NEW YORK STATE GEOLOGICAL ASSOCIATION

62nd ANNUAL MEETING
SEPTEMBER, 1990

FIELD TRIP GUIDEBOOK

WESTERN NEW YORK AND ONTARIO

Fredonia State University College

NEW YORK STATE GEOLOGICAL ASSOCIATION

62nd Annual Meeting

September 28-30, 1990

FIELD TRIP GUIDEBOOK

Gary G. Lash, Editor
Department of Geosciences
State University College
Fredonia, NY 14063

Department of Geosciences

State University of New York College at Fredonia

Fredonia, New York 14063

cover: Barcelona Light, Aimée Tourgée

TABLE OF CONTENTS

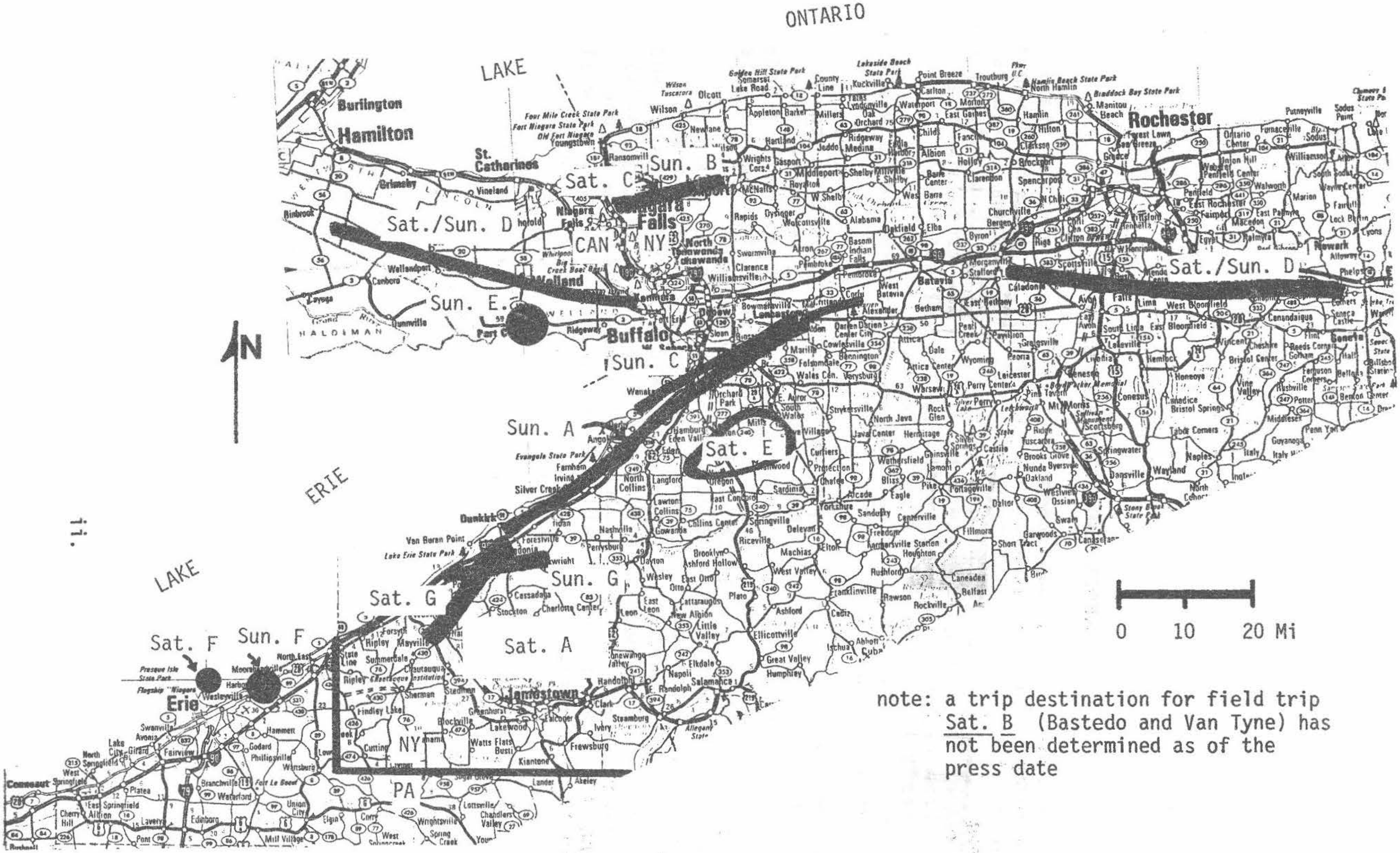
TRIP Sun. B: STRATIGRAPHY, STRUCTURE, AND HYDROGEOLOGY OF THE LOCKPORT GROUP: NIAGARA FALLS AREA, NEW YORK
Dorothy Tepper, William Goodman, Michael Gross, William Kappel and Richard Yager

TRIP Sun. C: PALEOECOLOGY AND SEDIMENTOLOGY OF SHORT-TERM SEALEVEL FLUCTUATIONS RECORDED WITHIN THE LOWER WANAKAH SHALE MEMBER, LUDLOWVILLE FORMATION
Keith B. Miller

TRIP Sun. E: SHALLOW WATER REEFS OF THE MIDDLE DEVONIAN EDGECLIFF MEMBER OF THE ONONDAGA LIMESTONE, PORT COLBORNE, ONTARIO, CANADA
Thomas Wolosz

TRIP Sun. F: A FRESH LOOK AT THE SEDIEMNTS WITHIN THE TERRACES AND CLIFFS THAT PARALLEL THE LAKE ERIE SHORE IN ERIE COUNTY, PENNSYLVANIA
David Thomas and Raymond Buyce

TRIP Sun. G: PADDLING UP A MELTWATER CHANNEL: A LATE-WISCONSINAN ICE-MARGINAL CRUISE NEAR FREDONIA, NEW YORK
Randy J. Woodbury and Michael D. Jensen



note: a trip destination for field trip <u>Sat. B</u> (Bastedo and Van Tyne) has not been determined as of the press date

GENERAL FIELD TRIP MAP FOR THE 62ND ANNUAL MEETING OF THE NEW YORK STATE GEOLOGICAL ASSOCIATION (field trip locations are only approximate)

PREFACE AND ACKNOWLEDGEMENTS

On behalf of the Department of Geosciences and the State University College at Fredonia, I welcome you to the 62nd Annual Meeting of the New York State Geological Association. We have attempted to provide you with a variety of trips that display the geological diversity of western New York. The trips run the gamut from oil and gas exploration to glacial and coastal processes to paleoecological and paleoenvironmental considerations of various stratigraphic units exposed in western New York and Ontario.

I wish to thank the field trip leaders and particularly those who made life easier by submitting their manuscripts on time. Mr. Jack Ericson of the College Archives provided the cover figure of Barcelona Light. I am grateful to Nancy Jagoda, our department secretary, for her invaluable assistance in this undertaking. Fredonia State students Blaine Smith, Kurt Schober, Kevin McGovern, and Richard Newman assisted in all phases of preparation of the guidebook. Eileen and Jocelyn Lash, Walther Barnard and Jason David Pacos helped to print and collate various chapters of the guidebook. Last but certainly not least, Janet Wentz, Tom Gestwicki and Dave Lillie of the Campus Printshop provided valuable technical assistance as well as much needed space.

DEVONIAN STRATA AND PALEOENVIRONMENTS: CHAUTAUQUA COUNTY REGION: NEW YORK STATE

GORDON C. BAIRD and GARY G. LASH
Department of Geosciences
State University of New York College at Fredonia
Fredonia, New York 14063

INTRODUCTION

The history of the study of the Paleozoic stratigraphic divisions in New York State spans, at least, two centuries, and it records the early development of important concepts centrally germane to the study of sedimentary basins worldwide (see Tesmer, 1989). The works of James Hall, John Clarke, Henry S. Williams, Joseph Barrell, and George Chadwick are well known to most sedimentary workers, as are those of numerous subsequent workers. In particular, the sedimentary sequences of the Allegheny Plateau Region ("Southern Tier" region) served as important sources of information for refinement of the facies concept and for the widespread recognition of the westwardly prograding Catskill Delta Complex which is the primary Paleozoic story recorded in Southern Tier bedrock deposits (see Chadwick, 1924, 1933; Cooper, et al., 1942; Caster, 1934; Woodrow, 1985).

The first significant geologic work relating to the Chautauqua County region was presented in James Hall's (1843) Survey of the Fourth Geologic District. Subsequent stratigraphic studies in this area include Clarke (1903), Chadwick (1923, 1924), Caster (1934), Pepper and de Witt (1950, 1951) and de Witt and Colton (1953). In northwest Pennsylvania, significant synthetic contributions include I.C. White (1881), Butts (1906-1908), Chadwick (1925), Caster (1934), Pepper et al. (1954). In Ohio, deposits equivalent to parts of the New York and Pennsylvania Upper Devonian section (Chagrin Shale) have been the subject of recent paleoenvironmental studies (see Weidner and Feldmann, 1983; Schwimmer and Feldmann, 1990). The present authors particularly build from the comprehensive work of Tesmer (1963) and the subsequent work of Murphy (1973) and Burrier (1977).

Both prior- and ongoing studies of the present authors are directed to the Canadaway-Chadakoin succession exposed on the Lake Erie lowland plain and adjacent escarpment. One of us (Baird) is currently tracing Chadakoin strata from Ohio northeastward along the escarpment to the vicinity of Fredonia, and he is also examining eastward (upslope) facies changes associated with the base of the Dunkirk Shale Member. Lash is currently working on the origins of interbedded siltstone-shale deposits and massive siltstone units (Laona and Shumla members) within the Canadaway Formation.

Despite the earlier regional mapping syntheses of White, Butts, Caster, and Tesmer, the Devonian bedrock geology of the western Southern Tier and northwest Pennsylvania region is not as well known as the classic Middle Devonian and lower Upper Devonian (Frasnian) succession to the north. One reason for this is the apparent paucity of widespread mappable event-beds and discontinuities in the thick high Devonian (Famennian) succession; this problem, coupled with the sparsely fossiliferous nature of many of the thickest units may have discouraged many from attempting to map or study these deposits. A second problem is the relative scarcity of long, clean, continuous stratigraphic sections in the Allegheny Plateau region southeast of the Erie Lake escarpment.

During this field trip we hope to dispel any impressions that this region is a geological "Empty Quarter"-that these deposits are "monotonous" and "unmappable;" we intend to show that, through preliminary mapping of certain units within the Late Devonian (Famennian) Canadaway and Chadakoin formations, we have already found additional mappable key beds, critical to refinement of the existing stratigraphy. In addition, reconnaissance examination of creek and Lake Erie shore sections has turned up both sedimentological and paleontological surprises.

In the present paper we examine more closely beds and facies associated with the boundary between the Hanover and Dunkirk members; in particular we document the occurrence of a black shale-roofed discontinuity near the base of the Dunkirk and its complex eastward passage to extinction. Higher in the Canadaway Formation we examine a distinctive basinal facies ("zebra-facies") characterized by thin repeating alternations of minimally bioturbated black and green shale beds. We report the distinctive and problemmatic occurrence of pyritic microspheres and partial microspheres which occur at hundreds of levels within the Canadaway Formation and which display evidence of exhumation and hydraulic concentration at many stratigraphic levels. In addition, we discuss the biota and distinctive preservation of pyritic fossils in the Corell's Point Goniatite Bed within the Gowanda Shale Member. Moreover, we will discuss the inferred depositional history of massive Canadaway siltstone units, particularly exemplified by the Laona Siltstone (see Stops 1 and 4), which are important thin stratigraphic markers in Chautauqua County. Discussion of the basal Dunkirk-, Corell's Point Bed-, pyritic microsphere-, and Laona Siltstone-problems are respectively complemented by field trip stops 1 to 4.

The stratigraphy and facies gradients within the Chadakoin Formation are described and discussed with respect to refined mapping of the Foerstia (Protosalvinia) zone across northwest Pennsylvania and western Chautauqua County; discovery of a thin zone of concentrated Foerstia, sp. (probable reproductive structure of an extinct fucoid alga) in Twentymile, Chautauqua, and Prendergast creeks, allows more precise matching of the New York and Pennsylvania Chadakoin sections and it has led to the recognition of mappable key beds and intervals particularly within the Ellicott Shale Member.

Finally, we will examine and discuss the coarse, nearshore facies of the Cattaraugus Formation. In particular, we will discuss the anomalously thick lentils of the Panama and other similar "white pebble" conglomerate units in the New York Cattaraugus section as to their inferred depositional setting and transport history. Stops 6 and 8 are respectively directed to the examination of the interval of concentrated *Foerstia* within the Ellicott Shale and to examination of the Panama Conglomerate within the Cattaraugus Formation at Panama Rocks near Panama, Chautauqua County.

LATE DEVONIAN PALEOGEOGRAPHIC, PALEOENVIRONMENTAL, AND TECTONIC SETTING

Using the Late Devonian reconstruction of Scotese et al., 1985, the western New York study area was positioned at about 3° to 5° South latitude in what must have been a tropical climatic regime which was moist, or, at least seasonally moist (Fig. 1). A broad epicontinental sea covered the study area, but a prograding shoreline existed to the immediate southeast of the Cattaraugus-Chautauqua county region. A rising collisional overthrust complex, the "Acadian Mountains" was present in the eastern Appalachian region; this uplift apparently was a response to convergence of the Armorican Plate or Avalon terranes with Laurussia and it involved diachronous northeast-to- southwest oblique collision with extensive strike-slip and transpressive tectonic activity (see Ettensohn *et al*. 1988).

Uplift of the Acadian Mountains was accompanied by the formation of a large foreland basin to the west and northwest of the overthrust belts (in the present directional sense); erosion of the tectonic uplands resulted in gradual filling of the trough with terrigenous deposits to form the Catskill Delta Complex (Fig. 2). The first evidence of major orogeny-related basin-filling is recorded in the Lower-Middle Devonian (Uppermost Eifelian-Lower Givetian-age) Marcellus Formation in eastern New York; by the time maximal basin-filling and coastal progradation were affecting westernmost New York during the latest Devonian (Famennian) the Catskill Delta had prograded hundreds of kilometers to the west and southwest (Ettensohn, 1985; Woodrow, 1985). By the lower Mississippian, basin-filling was largely complete in New York State, but epicontinental marine conditions persisted in northwest Pennsylvania and Ohio (Fig. 2).

The Devonian epicontinental sea was apparently deepest to the west and southwest of the prograding delta complex and the water column was typically stratified in the basin center during the Late Devonian with prominent and periodic development of salinity- and/or temperature-related density layers (pycnoclines) in deeper areas. Tropical upwelling, influxes of fresh water from deltas, and thrust load-related deepening of the foreland basin are all factors which may have caused the basin stratification (Thayer, 1974; Byers, 1977; Ettensohn, 1985; Woodrow, 1985) anoxic conditions in the basin with the consequent deposition of organic-rich, laminated muds below the pycnocline. Transgressive rises of the pycnocline apparently produced, in ascending order, the Oatka Creek, Geneseo, Middlesex, Rhinestreet,

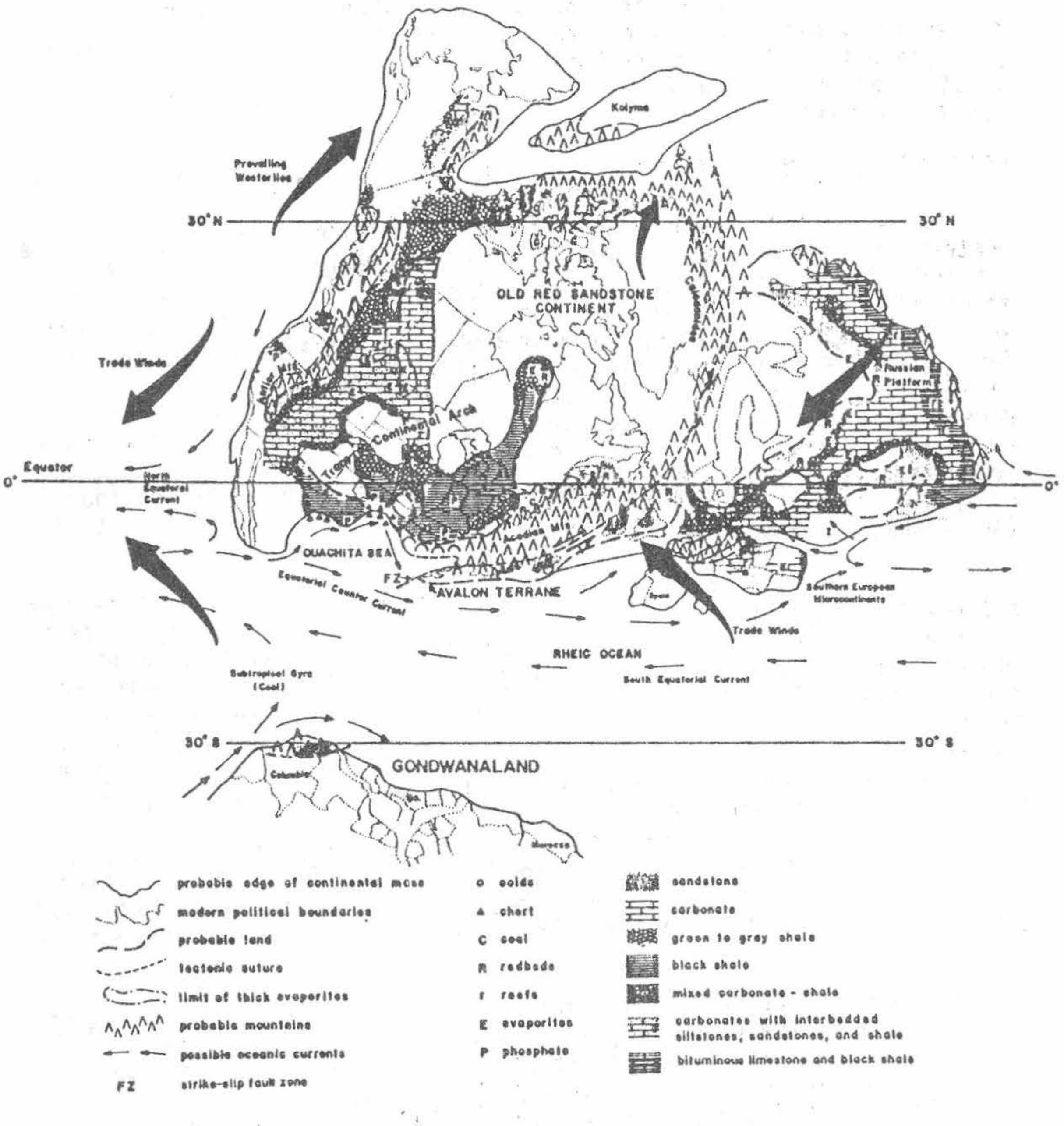


Fig. 1.-Late Devonian paleogeography (from Ettesohn et al., 1988).

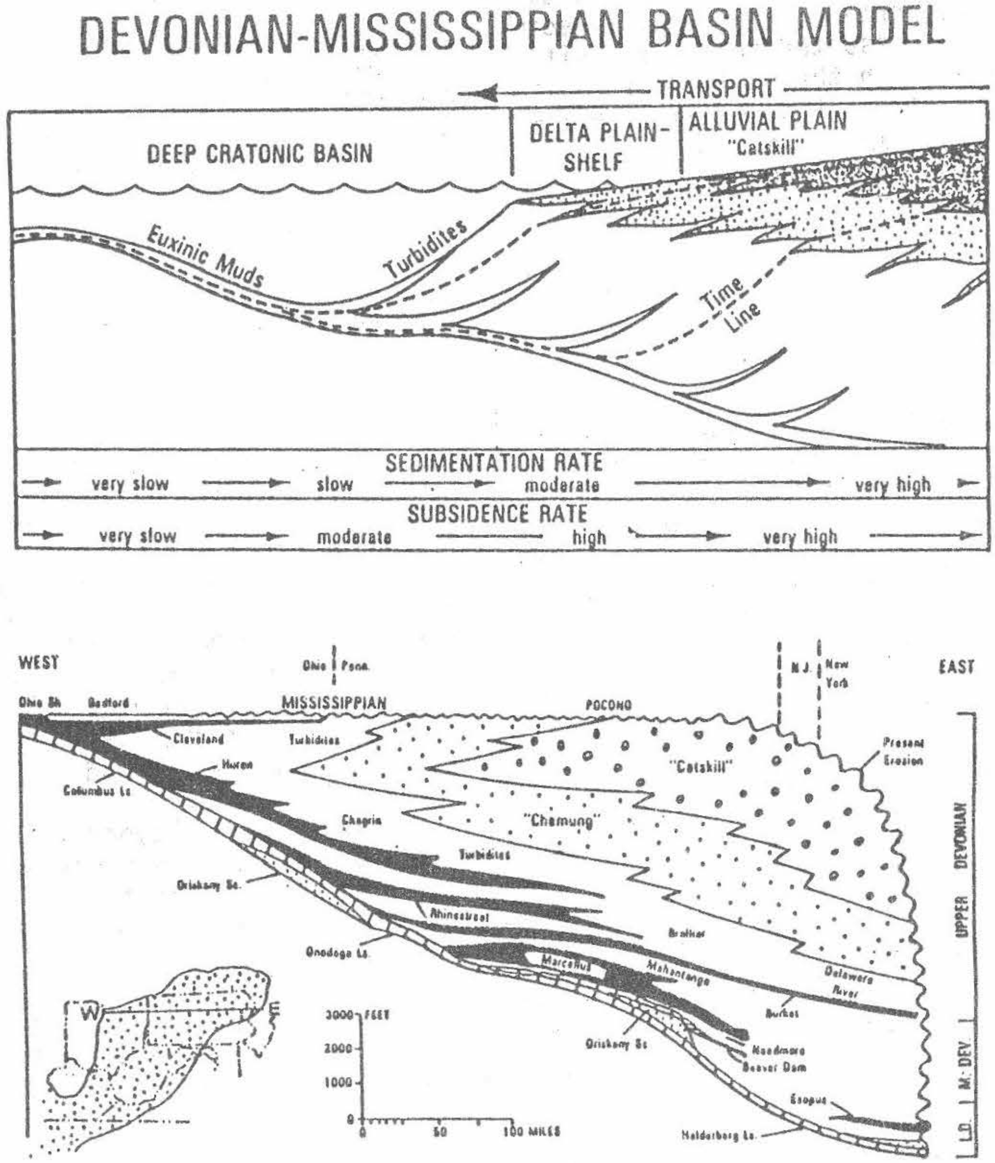


Fig. 2. Idealized east-west cross-section through Catskill Delta Complex: depositional model (top) and simplified stratigraphy (bottom). From Broadhead et al., (1982).

Pipe Creek, and Dunkirk black shales as well as many thinner unnamed black shale units. During times of relative sea level fall large areas of the epicontinental seafloor became variably oxygenated; such conditions resulted in the deposition of green-grey, bioturbated muds, sometimes with the shell debris of various bottom organisms. The vertical alternation of major black and grey-green shale unit in the Upper Devonian prodelta and basin deposits in western New York probably records temporal, cyclic changes in sea-level and paleoclimatic conditions which are superimposed on longer-term tectonic and progradational processes.

STRATIGRAPHIC FRAMEWORK AND FACIES UNITS

Stratigraphy

Upper Devonian (Famennian) deposits in the Chautauqua County - northern Erie Co., Pennsylvania region include approximately 600 meters (1,900 ft.) of marine, nearshore marine, and nonmarine deposits (see Fig. 3). We will examine only a small fraction of this overall succession at eight field stops; the lowest level we will visit is the Hanover Shale-Dunkirk Shale contact near the Frasnian - Famennian stage boundary and the youngest unit will be the Panama Conglomerate within the Cattaraugus Formation. Generally the stratigraphic sequence becomes coarser from the base upwards and there is an overall regressive lithologic transition from basinal black shale facies at the base, through a long prodeltaic succession of interbedded siltstone and bioturbated grey mudstone layers, to siltstone- and sandstone - dominated, shell-rich delta platform deposits at the top. Beyond the reach of our day-long field trip venue are red bed tongues in eastern Chautauqua, Cattaraugus and Allegheny counties which approximate the western limit of nonmarine conditions at the time the Cattaraugus Formation was deposited (Tesmer, 1963). In principle, most of the layers within the Famennian sequence individually grade eastward-southeastward ("upslope") into coarser, shoreward equivalent deposits, and westward-northwestward ("downslope") into finer basinal facies reflecting the generalized Waltherian facies pattern of this part of the Catskill Delta (Fig. 2).

The Famennian stratigraphic succession in Chautauqua County includes three formations (Fig. 3), using the nomenclature of Tesmer, 1963; Buehler and Tesmer, 1963; Tesmer, 1974, 1975; these include in ascending order the Canadaway, Chadakoin and Cattaraugus formations which are unconformably overlain by the Mississippian Knapp Conglomerate in New York, (Tesmer, 1963, 1975). There are still younger Devonian divisions in Ohio; the black Cleveland Shale and still higher basal strata of the grey Bedford Shale yield Devonian clymeniid goniatites (see House, et al, 1986).

The Canadaway Formation is composed mainly of fissile, organic-rich black shales, usually interbedded with green bioturbated or non-bioturbated mudstone and siltstones. With the exception of the Corell's Point Goniatite Bed and a few other, similar pyrite-rich,

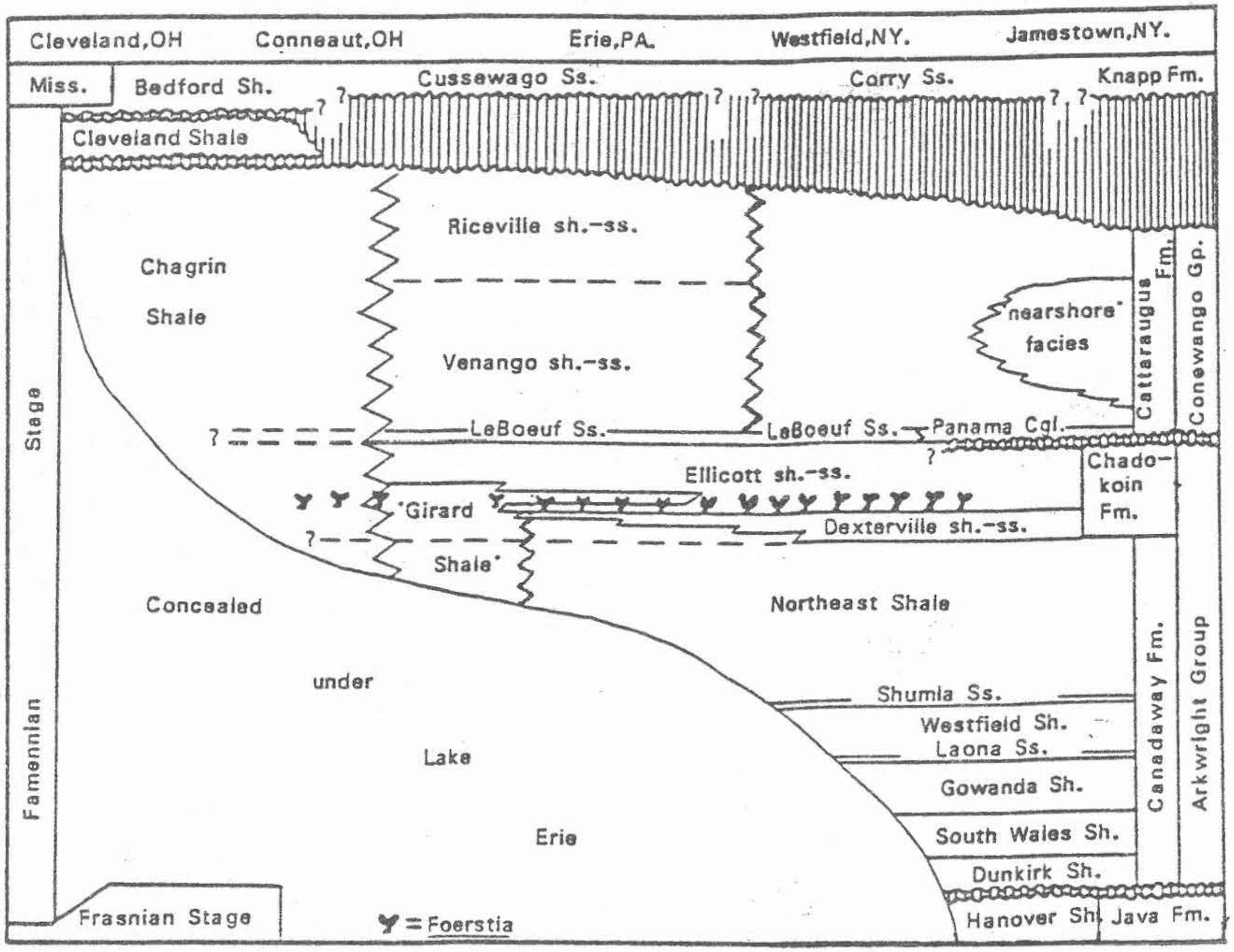


Fig. 3. Chronostratigraphic cross-section of Famennian deposits in Lake Erie region Modified from Rickard (1975), to include equivalent Devonian units in northwest Pennsylvania and Ohio.

concretion-bearing green mudstone units, this 300 meter (1,000 ft.)-thick interval contains few shelly fossils. Some green mudstone and siltstone-rich intervals, however contain numerous trace fossils, some of which are undescribed. Significant fossils include zonally-important goniatitic cephalopods (see Stop 3), rare fish fossils (see Stop 2), and hexactinellid ("glass") sponges which are reported from the Northeast Shale Member (Tesmer, 1963).

The Chadakoin Formation, including the Dexterville and Ellicott members, is distinctly siltier and richer in shelly fossils; this 120 meter (375 ft.)-thick division is the lowest unit composed of the silty, fossiliferous "Chemung" magnafacies (*sensu* Rickard, 1975) which represents well oxygenated, storm-influenced bottom conditions on the uppermost prodelta slope and platform (see Sutton, et *al*. 1974; Thayer, 1974; Sutton and McGhee, 1985; Burrier, 1977). Spiriferid, productid, and rhynchonellid brachiopods, bryozoans, bivalves, and diverse trace fossils are conspicuous componants, particularly in the upper two thirds of the formation. Significant fossils include glass sponges, horseshoe crab body-and trace fossils, plus diverse echinoderms, including inadunate crinoids, undescribed archaeocidaroid echinoids and stelleroids, as well as the algal fossil *Foerstia* (see Stop 6). Swaley siltstones, siltstone para-ripples, gutter casts, and abundant tempestitic siltstone beds and coquinites attest to frequent storm-wave impingement on the substrate (Burrier, 1977; this paper).

In Chautauqua County, the 215 meter (650 ft.)-thick Cattaraugus Formation begins with the first beds of quartz-pebble (orthoquartzitic) conglomerate or thick-bedded white sandstone which appear above Chadakoin mudstones and siltstones. Tesmer, 1963, indicates that the basal Cattaraugus quartz-pebble-rich sandstones corresponds to the level of the Panama Conglomerate, a unit of highly variable thickness within the county (see Stop 8). Recent mapping by Baird in western Chautauqua County suggests that the Panama Conglomerate correlates to a thinner, white (quartz-rich) sandstone (LeBoeuf Member) which can be observed near Sherman and Summerdale (Fig. 3). Post-Panama Cattaraugus deposits are, on average, siltier and sandier than those within the Chadakoin and additional, higher conglomeratic units are reported, particularly in eastern Chautauqua County and in Cattaraugus County (Tesmer, 1963, 1975). Cattaraugus deposits are usually very fossiliferous; moldic spiriferid, productid, and rhynchonellid brachiopods, bryozoans, bivalves, and crinoid stem debris are commonly observed at the numerous, small weathered outcrops which dot the upland parts of the county. A very large, triangular spiriferid brachiopod, *Sphenospira alta*, characterized by an extremely broad interarea (see Schwimmer and Feldmann, 1990) can be found in Cattaraugus siltstone beds in creeks south of Sherman. Cattaraugus fossils include many species different from those in the underlying Chadakoin Formation suggesting a significant period of erosion and nondeposition preceding Cattaraugus deposition (see Caster, 1934). Again, numerous features such as hummocky cross-stratification and tempestitic beds attest to strong storm-wave influence in Cattaraugus platform settings.

Facies Units

Five key types of spectral magnafacies occur within the regressive sequence of the Catskill Delta (see Rickard, 1975); the divisions relevant to this report include: in respective shoreward-order: "Marcellus", "Portage," "Chemung," and "Cattaraugus." (The nonmarine "Catskill" (Red bed) magnafacies is exposed east and southeast of the region we will visit). The Marcellus magnafacies, also called "Cleveland" magnafacies by Caster (1934) includes the well known organic-rich ("black") shale units; these deposits include laminated, black shales, exemplified by the Dunkirk Member, which contain no benthic organisms, or, at least, extremely few benthos. Anoxic bottom conditions are believed to explain the lack of bottom organisms and the lack of benthos. Most workers attribute the pervasive water stratification to maximal inferred water depths for these deposits; depth estimates for this facies usually are pegged in the "lower infralittoral" range (50 to 200 meters) for black shales in the Appalachian Basin (Bowen et al., 1974,; Thayer, 1974; Broadhead, et al. 1982). Bowen, et al, 1974, describe this facies as representing nondeltaic outer slope and basin conditions.

The Portage magnafacies includes mudstone, siltstone, and fine sandstone deposits which typically lack benthic shelled organisms or yield only low diversity assemblages of benthic and pelagic organisms (Rickard, 1975). Deposits of this type represent dysoxic (poorly oxygenated) bottom conditions which could only support soft-bodied or thin-shelled bottom organisms (see Rhoads and Morse, 1971; Savrda and Bottjer, 1987; Wignall, 1990). Typically thin siltstone or sandstone beds with sole marks often characterized by groove- or flute casts, alternate with bioturbated grey mudstone, non-bioturbated grey mudstone, and discrete thin black shale beds. Relative proportions of these various lithologies vary with stratigraphic level between the Dunkirk Shale and the base of the Chadakoin Formation, but there is an overall regressive stratigraphic trend from black shale-dominated sections at the base to bioturbated silty mudstone and siltstone-dominated sections at the top.

Siltstone beds in this magnafacies include gravity-flow deposits that appear to have a turbiditic appearance but there are many beds which have complex internal bedding which may reflect deep storm-wave impingement or other bottom-current processes (see discussion: Stops 1, 3, 5). This magnafacies is believed to record sedimentation along the lower-to-middle prodelta slope and the proximal part of the offshore shelf near the delta. (Sutton, et al., 1974; Thayer, 1974).

Fossiliferous mudstone, siltstone, and sandstone deposits within the Catskill Delta complex are largely grouped under the magnafacies designation "Chemung" (see Rickard, 1975); these deposits yield abundant and sometimes diverse communities of filter-feeding and deposit-feeding benthic organisms. Coquinitic storm deposits are abundant as are soft-sediment deformation features such as ball-and pillow structures, and there is abundant evidence of bottom smothering of organisms (see Stops 6,7). The entire Chadakoin Formation and much of the Cattaraugus Formation in western Chautauqua County and equivalent Venango Formation in Erie County, Pennsylvania fall within this magnafacies.

Nearshore deposits are referred to the "Cattaraugus" magnafacies by Rickard (1975); in such sequences red beds alternate with grey mudstone and sandstone intervals representing nearshore marine or brackish-water settings. This magnafacies usually contains low diversity fossil assemblages, and such features as flat-pebble conglomerates, ball-and-pillow deformation, and locally abundant plant debris are frequently observed. River-influenced nearshore shelf, estuarine-marine, and intertidal marsh-mudflat conditions are believed to be recorded by such sediments. Nearshore marine Cattaraugus deposits accumulated only a short distance west of the floodplain-fluvial channel paleoenvironments recorded in the Catskill magnafacies (Rickard, 1975). The Cattaraugus magnafacies is well developed within the Cattaraugus Formation in the upland regions south of Randolph, in the Alleghany State Park, and around Olean (see discussions in Caster, 1934; Tesmer, 1974; Rickard, 1975).

STRATIGRAPHIC UNITS

Member divisions:

Descriptions of component members of the upper Java, Canadaway, Chadakoin, and Cattaraugus formations are included below in order of stratigraphic succession. Descriptions are brief for units not examined and/or not visited on the field trip.

Java Formation:

Hanover Shale Member (See Figs. 4, 5; Stop 2). The Hanover Member, which is the uppermost unit containing fossils of the Frasnian Stage, was originally described by Hartnagel (1912) for exposures which occur near Silver Creek in Hanover Township. This 28 meter (90 ft.)-thick unit is composed primarily of green-grey mudstone with accessory discrete thin black shale beds and zones of small calcareous, phosphatic and pyritic nodules. The Hanover yields few shelly fossils, but is usually intensely bioturbated to the degree that some grey mudstone intervals have a massive, weakly-bedded appearance. It is distinctive in lacking many discrete siltstone beds which is the normal condition for late Devonian Portage-type magnafacies. The Hanover changes eastward into siltier, upslope deposits (Wiscoy Member) southeast of the type section at Java Village. Hanover outcrops are numerous and good in the region between Dunkirk Harbor and Warsaw, Wyoming County.

Hanover fossils include a mixture of benthic and pelagic fossils; goniatites and carbonized driftwood fragments are common at Silver Creek. Small rugose corals and crinoid ossiclesoccur in the nodular beds at numerous levels. The Frasnian-Famennian stage boundary position of the end-Frasnian extinction event has been tentatively pegged several meters below the top of this unit (Kirchgasser, pers. comm. 1989).

Canadaway Formation:

Dunkirk Shale Member (See Figs. 4, 5; Stop 2). The Dunkirk Member, named by Clarke (1903) for exposures at Dunkirk Harbor (see Stop 1), is a hard, radioactive, black shale unit which is about 12.5 meters (40 ft.)-thick in the Dunkirk area (Tesmer, 1963). This unit is southwestwardly correlative with the basal Huron Shale in Ohio (Woodrow, et al., 1988), and tongues of black Dunkirk can be traced eastward to the vicinity of Hornell, Steuben County (Pepper and de Witt, 1951).

The Dunkirk is a hard, prominently-jointed black shale which contains large (1-5 foot-diam.) septarial limestone concretions and numerous amoebiform pyrite nodules. Usually this unit is laminated, but close examination reveals the presence of small-scale bioturbation at some levels which suggest that low levels of bottom oxygenation occurred during some of the time of black mud accumulation. Benthic taxa include the linguloid brachioped Barroisella. Pelagic organisms and transported debris include: conodonts, algal spores, fish fossils and carbonized driftwood.

At Point Gratiot (Stop 1) the 17 cm (6.5 in.)-thick basal bed of Dunkirk, separated from the main overlying continuous black shale sequence by a 13.5 cm (4.5 in.)-thick bioturbated grey mudstone unit, yields abundant carbonized plant debris plus the remains of both articulated and disarticulated large armored fish (Fig. 4A). The top of the 17 cm grey band marks a submarine discontinuity which contains a lag bed of detrital pyrite, small fish bones, wood debris and conodonts (Baird and Brett, 1986).

As this discontinuity is traced northeastward to the vicinity of Java Village, Wyoming County, more and more beds progressively appear beneath it such that this break passes essentially to extinction (Fig. 5). Several black shale beds alternating with bioturbated and non-bioturbated grey-green mudstone make up an intervening sequence several meters-thick (Fig. 5). Most of the thin black shales, however, display minor diastemic basal contacts with the grey-green beds marked by minor detrital pyrite. Collectively, these diastems truncate underlying units as they are traced southwestward such that higher diastems apparently downcut through lower ones until only the single 13.5 cm grey-green bed remains at this locality (Figs. 4, 5). This illustrates the pervasiveness of submarine erosion associated with the transgression grey-to-black facies change beneath even very thin black

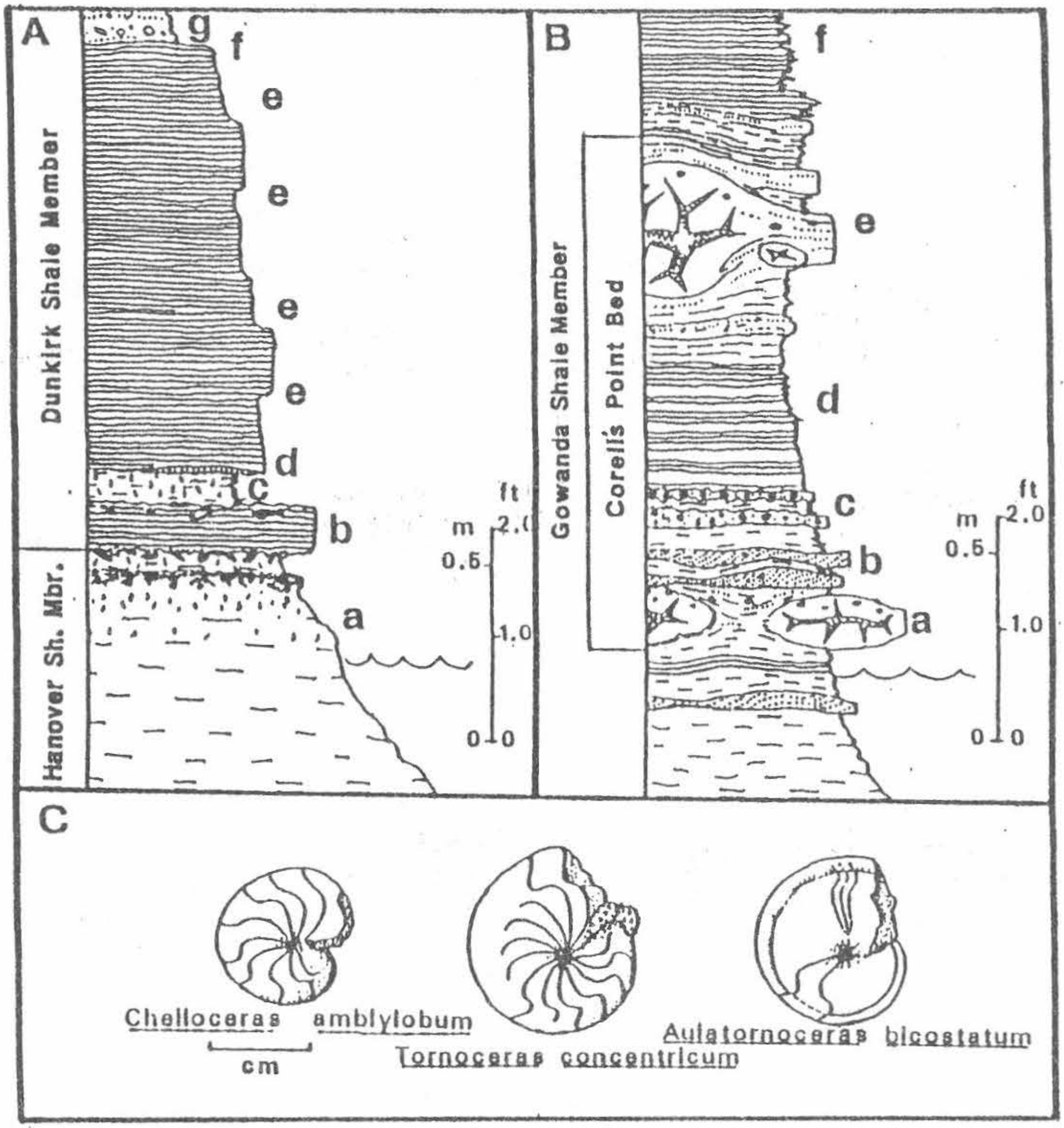


Fig. 4. Stratigraphic sections for the Point Gratiot and Corell's Point lake shore localities and key goniatite fossils from the Corell's Point Bed. A) Section at Point Gratiot (STOP 2). Lettered units include: a) intensely bioturbated grey-green mudstone; b) wood-and bone-bearing black shale bed; c) bioturbated grey-green mudstone bed; d) detrital pyrite lens along black shale-roofed discontinuity; e) laminated black shale; f) striated glacial scour contact; g) bedded glacial "till"; B) Section at Corell's Point (STOP 3). Lettered units include: a) septarian concretions containing fossiliferous pyritic steinkerns; b) siltstone beds containing unfossiliferous pyrite nodules, c) bioturbated siltstone beds yielding numerous pyrite nodules, pyritic fossil steinkerns and non-pyritic fossils; d) interbedded black and grey-green shale ("zebra" facies); e) large septarian concretions with occasional pyritic steinkerns and abundant auloporid corals; f) black shale; C) Key zonal goniatites from Corell's Point Bed (After House, 1962, 1965; Kirchgasser, 1974).

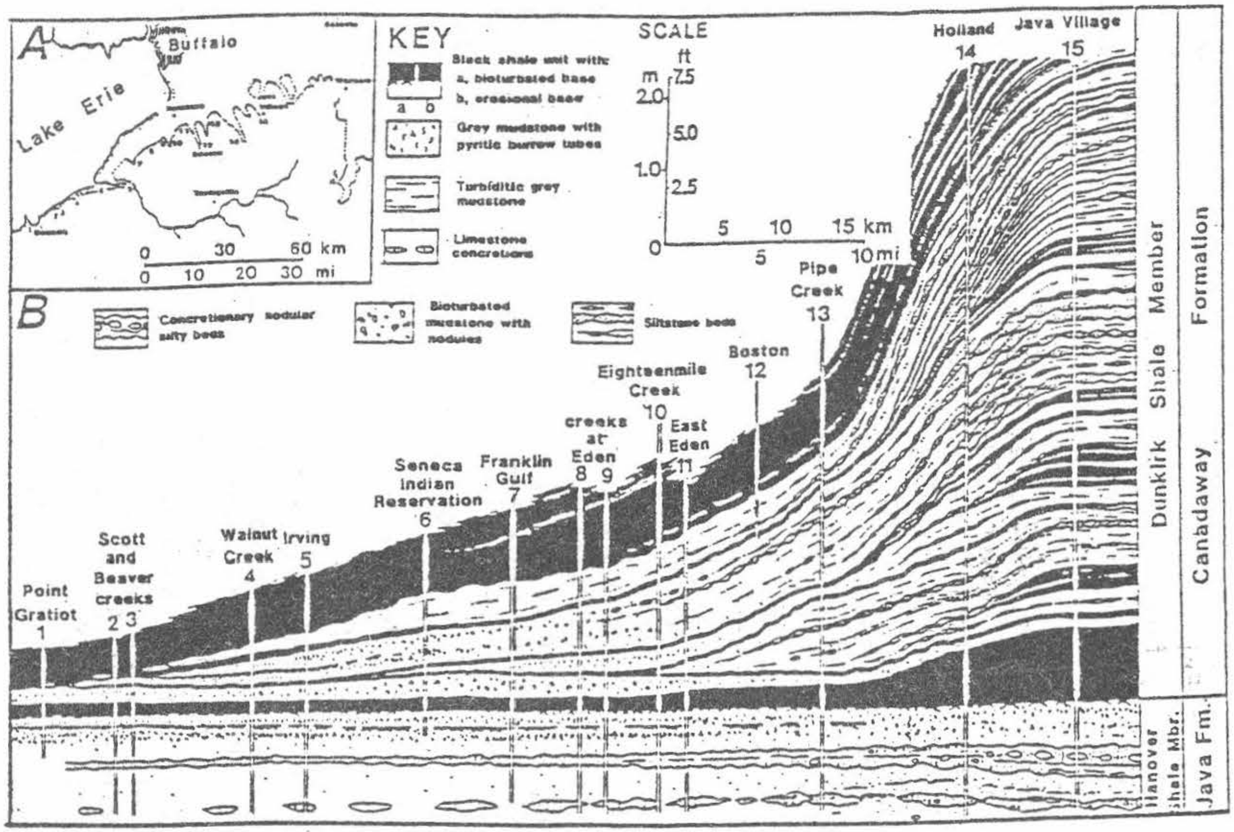


Fig. 5. Stratigraphic transect of basal Famennian strata (uppermost Hanover Shale Member and lowermost Dunkirk Shale Member) between Point Gratiot, Chautauqua County (STOP 2) and Java Village, Wyoming County. Note conspicuous northeastward thickening of basal Dunkirk as turbiditic grey mudstone facies appears within section. Many black shale beds in expanded (splayed) northeastern Dunkirk deposits have erosional bases. Southwestward convergence of these black shale beds involves, in part, the apparent downcutting by one or more of these beds into lower strata such that only one remaining grey mudstone bed is visible within the Dunkirk at the Lake Erie shore (see STOP 2).

shale units. It also illustrates the distinctive character of condensed sedimentary deposits involving alternations between dysoxic and anoxic facies.

South Wales Shale Member. Above the Dunkirk Member is an 18 to 25 meter (60-80 ft.)-thick interval of interbedded black and dark grey shale, grey-green mudstone and occasional flaggy siltstone beds designated the South Wales Member by Pepper and de Witt, 1951, for exposures in a small tributory to the east branch of Cazenovia Creek, three miles south of South Wales in southern Erie County. Large septarian concretions are associated with horizons of grey-green bioturbated mudstone. Macrofossils are rare but carbonized wood and rare fish fragments can be collected from this unit.

Gowanda Shale Member (See Figs. 4B, C, 6, 7; Stops 1, 3, 4). Above the South Wales Member is a 37-72 meter (120-230 ft.)-thick interval of interbedded black shale, marginally black shale, grey mudstone, and siltstone beds, designated by Hartnagel (1912), for exposures near Gowanda, Cattaraugus County. This unit is dominated by black shale in the lower half and contains numerous siltstone beds in the upper third. Thin intervals of bioturbated grey mudstone with both large and small calcareous concretions occur at several levels; these yield auloporid corals, bivalves, cephalopods and driftwood.

The best known of the fossil-bearing levels is the regionally mappable Corell's Point Bed (see House, 1966, 1968; Kirchgasser, 1974; Tesmer, 1974), which is, in reality, two-closely spaced pyrite nodule-and concretion-rich beds which yield uncrushed pyritic steinkerns of goniatites and orthoconic nautiloids (see Fig. 4B, C; Stop 2). The Corell's Point Bed is believed to be traceable as far to the northeast as Holland, Erie County and the south branch of Cattaraugus Creek east of Dayton (House, 1966, 1968); at the latter locality it occurs at an elevation of 965 feet which is anomolously high suggesting the possibility of significant structural upwarping of beds southeast of Gowanda. This Corell's Point Bed locality has yet to be examined by the present authors.

The upper Gowanda is characterized by finely interbedded discrete beds and laminae of black shale, non-bioturbated grey mudstone, bioturbated grey mudstone, and both lenticular and evenly-bedded siltstone (Fig. 6). Contacts are usually sharp between these divisions due to minimal bioturbation and these deposits have a striking banded appearance particularly along clean lake sections (see Stops 1, 3, 4). This type of deposit is herein designated by an informal name, "zebra facies," for economy of description; it is characteristic of parts of the Gowanda, Westfield, and Northeast shale members.

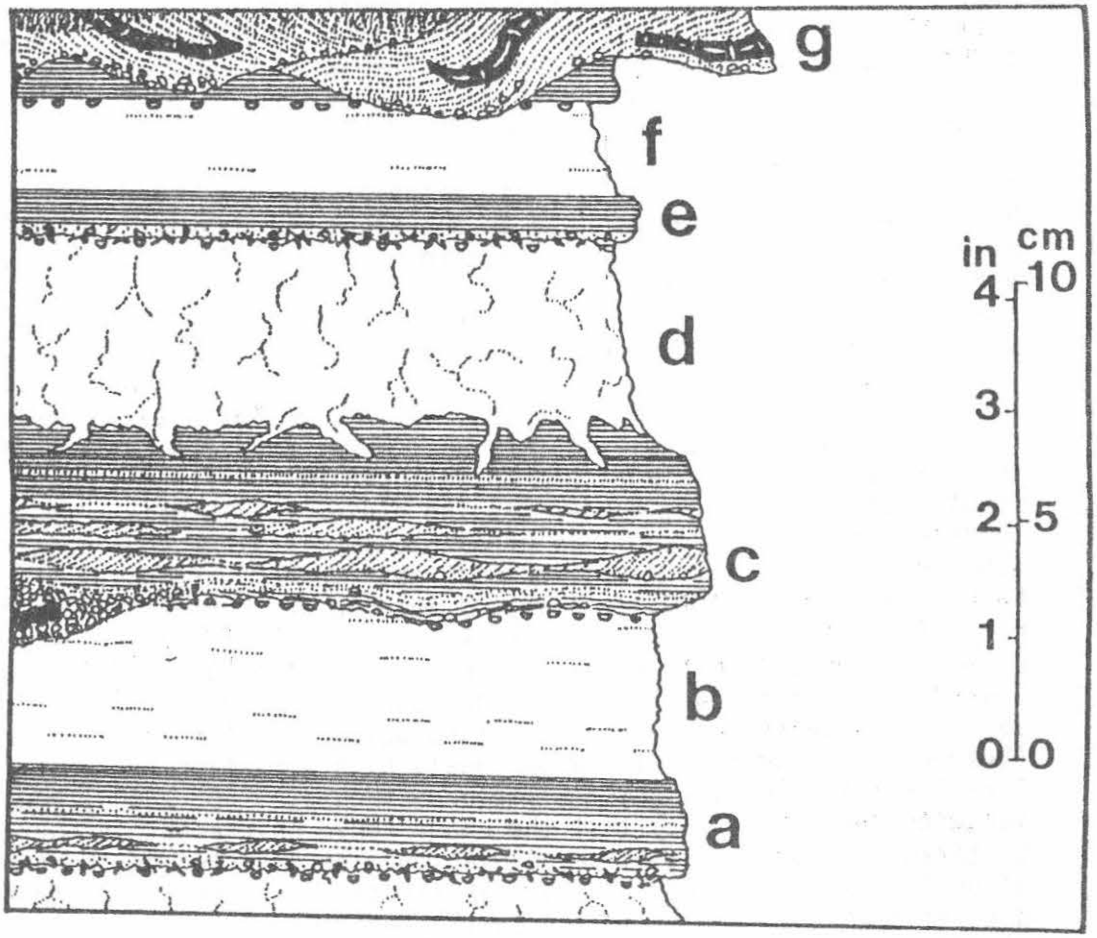


Fig. 6. Association of sediment-types and bedding relationships in finely interbedded grey mudstone-, black shale, and thin siltstone deposits ("zebra facies") typical of the lower and middle parts of the Canadaway Formation (see text; STOPS 1, 3, 4). Units shown include: a) black shale bed with in-situ partial, pyritic microspheres below the base and exhumed and reoriented microspheres at the base; b) turbiditic grey-green mudstone bed with in-situ partial microspheres below top; c) silty black shale bed with minor erosional contact at base. Both whole and partial reworked microspheres fill a scour depression on the erosion surface; d) bioturbated grey-green mudstone layer. Burrowers have penetrated the underlying black mud layer and have piped grey mud down into this bed; e) thin black shale bed similar to a; f) turbiditic mudstone layer similar to b; g) thicker siltstone bed recording significant bottom erosion by non-turbiditic currents prior to sediment accumulation. An underlying black shale layer is partly-to completely removed by erosion and both detrital pyritic microspheres and wood debris are concentrated within this siltstone bed.

The zebra facies records a complex succession of depositional events in an anoxic to lower dysoxic slope-to-basin setting. Although many of the grey-green mudstone and siltstone beds appear to be of turbiditic origin, some beds may have formed as a result of bottom current processes (see Stanley, 1987). Most unusual are the frequent occurrences of sand-sized pyritic microspheres at or near the bases of the thin black shale beds (Figs. 6, 7). These microspheres and partial microspheres may be the result of bacterial sulfate reduction within gas bubbles which were trapped beneath the black mud layers or under bacterial mats (Fig. 7). Reorientation of geopetal partial microspheres within many black shales due to current scour or bioturbation suggests that these black shales were deposited rapidly and that some black mud substrates could support burrowing organisms (see discussion; Stop 4).

Laona Siltstone Member (See Stops 1, 4). Above the Gowanda Member is a resistant, falls-forming siltstone unit designated the Laona Siltstone by Beck (1840) for the exposure we will see at our first stop at Laona, Chautauqua County. The Laona, ranging from 1 to 7.5 meters (3-25 ft.) in thickness, usually consists of quartzose siltstone beds which are about one foot in thickness; the most massive beds are usually at the base with thinner siltstone beds and interbedded shales towards the top. Presently, the Laona can be traced with certainty from Barcelona at the Lake Erie shore to Little Indian Creek near Perrysburg. An exposure near Gowanda originally identified as the younger Shumla Siltstone (Tesmer, 1963) now appears to be Laona on lithologic and thickness criteria (Tesmer, Pers. comm.). The anomolously high elevation recorded for the Corell's Point Bed southeast of Gowanda (House, 1966, 1968), however, suggests the possibility that the Laona could be present in northwest Cattaraugus County at a higher elevation than expected on the assumption of uniform southward dip of Devonian strata. However, no Laona deposit has yet been observed in that area (Tesmer, pers. comm.).

The Laona is unfossiliferous at most localities but near Nashville, Chautauqua County, a coquinite layer rich in brachiopods and bivalves was discovered by Tesmer, 1963. One of the present authors (Lash) is studying the Laona Member; based on thin section study and examination of sedimentary structures, he concludes that is unit is a non-graded gravity flow deposit. Tesmer's (1963) faunal list for this unit is interesting because it contains numerous taxa found in oxic upper prodelta and platform facies and which are distinctly absent from synjacent shales. It appears possible that these fossils may be allochthonous, and were transported downslope in one or more turbiditic flows into the anoxic or minimally oxygenated basin setting. We will see the Laona at Stops 1 and 4.

Westfield Shale Member (See Stop 4). Above the Laona Member is a 31 to 68 meter (100-220 ft.)-thick sequence of interbedded black and dark grey shale, grey mudstone, and thin siltstone beds designated the Westfield Shale Member by Chadwick (1923), for exposures on Chautauqua Creek near

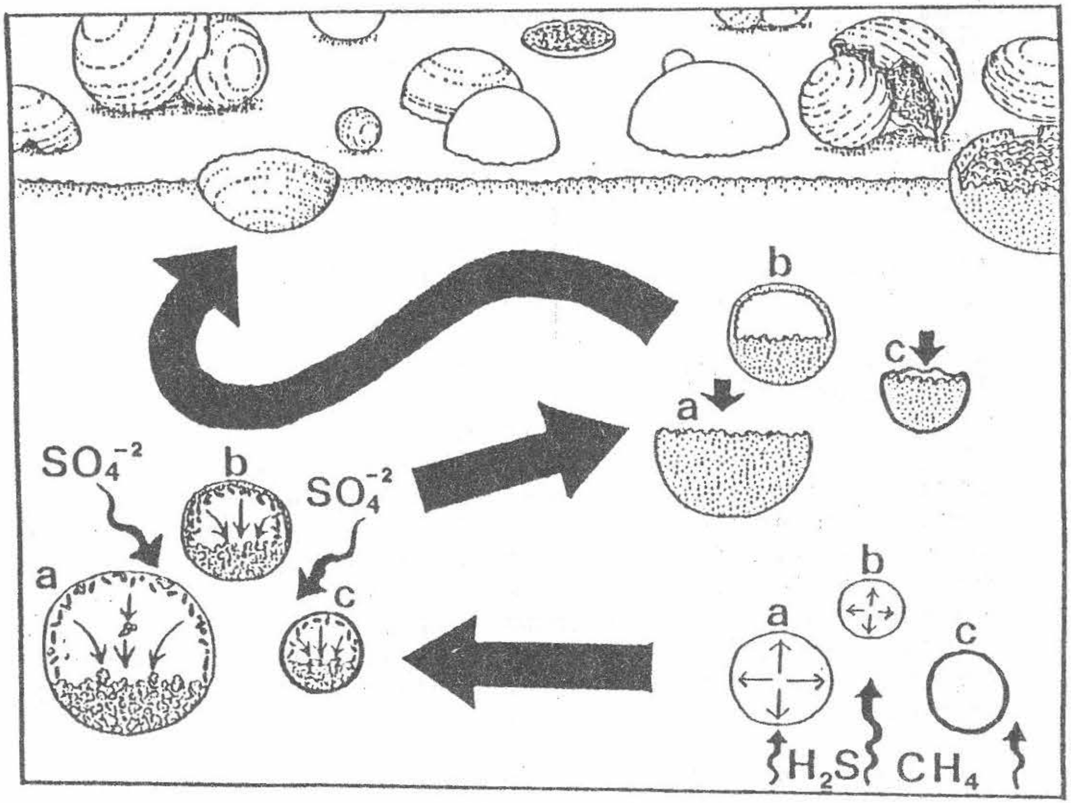


Fig. 7. Model for genesis of early diagenetic microsphere and partial (geopetal) microsphere pyrite. Sequential stages protrayed include: formation of bubbles (or spores) in near-surface turbiditic mud by rising decay gases; partial filling of bubble cavaties a and b sealing off flat floored geopetal cavity inside; and exhumation of shiny complete and partial spheres during scour events preceding deposition of black muds and/or turbiditic sediments (not shown here to facilitate viewing of multiple orientations of detrital pyrite grains on scour surface). Sphere c is a Sporangites (Tasmanites) spore which undergoes sequential partial filling by pyrite, collapse due to loading, and exhumation with presumed destruction of cutical cover. This latter interpretation is not favored here because no in-situ sphere and partial sphere pyrite is observed within spore cutical coverings.

Westfield, Chautauqua County. The Westfield is traceable from the Lake Erie Shore near the New York/Pennsylvania state line to the vicinity of Perrysburg where its bounding units (Laona and Shumla members) ceased to be present. The Westfield is lithologically similar to the Gowanda Member. We will examine Zebra facies in the lower Westfield at Stop 4.

Shumla Siltstone Member. Above the Westfield Member is an interval of thin-bedded siltstones up to 11 meters (35 ft.)-thick which is designated the Shumla Siltstone Member by Clarke, 1903, for beds outcroping along Canadaway Creek at Shumla, Chautauqua County. The Shumla is traceable from the Lake Erie Shore near the New York/Pennsylvania border to the vicinity of Perrysburg where it apparently pinches out (Tesmer, 1963). Lithologically the Shumla is similar to the Laona Member and it is usually unfossiliferous. As with the Laona, this unit consists of gravity flow units. We will not observe this member on the field trip.

Northeast Shale Member (See Stop 5). The Northeast Shale Member, named for exposures near Northeast, Pennsylvania by Chadwick, 1923, is a 125 to 188 meter (400-600 ft.)-thick unit composed mainly of interbedded silty grey mudstone and flaggy siltstone beds with a subsidiary component of dark grey and black shale beds in the middle and lower parts of the member. Calcareous "Cone-in-Cone" concretions occur at several levels within this unit (see Woodland, 1964; Gilman and Metzger, 1967). Shelly fossils are rare to absent in Chautauqua County Northeast exposures but become more common in equivalent beds in Cattaraugus County (Tesmer, 1963). However, several types of trace fossils are common in the Northeast Member; at Stop 5 they are visible as bas-relief (hypichnial) features on the soles of many siltstone beds.

The Northeast Member records part of the upward-transitional (regressive) record from anoxic to pervasively dysoxic facies. The greater percentage of siltstone beds in this unit is consistent with its more shallow and shoreward inferred depositional setting as compared with underlying Canadaway shale divisions. Preliminary examination of this unit by the present authors suggests that it could be further subdivided into mappable siltstone-rich, siltstone-poor, and black shale-rich stratigraphic divisions. We will briefly examine the Northeast Member at Stop 5.

Chadakoin Formation

Background:

Chadwick proposed the name Chadakoin for a sequence of interbedded lenticular grey siltstone and grey mudstone deposits exposed in quarries along the Chadakoin River at Dexterville (now the eastern part of Jamestown). Tesmer (1963) subsequently formalized this name to formation status. The two component members of the Chadakoin are, in ascending order, the Dexterville Siltstone and the Ellicott Shale.

Dexterville Siltstone Member (see Fig. 8). The Dexterville Member was proposed by Caster, 1934, for approximately 31 meters (100 ft.) of interbedded lenticular, grey siltstone and grey mudstone deposits exposed in quarries along the Chadakoin River at Dexterville, New York. This unit is distinguished from the underlying, sparsely fossiliferous Northeast Member, by the appearance of thicker and more numerous lenticular siltstone beds and by the appearance of pavements and coquinite concentrations of brachiapods, bryozoans and bivalves. As such this unit marks the upward change to "Chemung magnafacies" of Rickard, 1975.

Dexterville strata in the Jamestown, Cherry Creek, and Randolph areas are rich in the brachiopod "Pugnoides" duplicatus which is restricted to the Dexterville (Tesmer, 1963). Burrier (1977), however, reported this brachiopod within only the basal part of the stratigraphic section mapped by him as Dexterville on Chautauqua Creek. "Pugnoides" has not yet been observed in Pennsylvania Dexterville exposures by the present authors and it has not yet been conclusively confirmed at Chautauqua Creek. Shell-rich Dexterville can be traced southwestward ("downslope") into the Erie, Pennsylvania area. Southwest of Erie, this unit changes to a "Northeast Shale"-type appearance as it grades from outer shelf-upper slope "Chemung" magnafacies into the slope-to-basin "Portage" magnafacies.

Ellicott Shale Member (See Stops 6 and 7). The Ellicott Member was named by Caster, 1934, for exposures "along Hunt Road" in Ellicott township west-southwest of Jamestown. A few exposures exist in the vicinity of this road between Jamestown and Ashville (see the railroad cut northwest of the Sugar Grove-Hunt Road intersection at Lakewood: Tesmer, 1974: Field Trip Stop 5), but a "type section" for this unit is virtually nonexistent. However, what has come to be understood as "Ellicott" in Chautauqua County and in Erie County Pennsylvania includes some of the most fossiliferous facies in the region. Moreover, it may prove to be the most useful for establishing refined correlation of time-synchronous event-strata along prominent depth-related facies gradients from New York into northeast Ohio.

Crudely defined, the Ellicott Member is characterized by complexly-interbedded lenticular siltstone, mudstones, and coquinites. It is actually very similar to the Dexterville lithologically except that, at least, two siltstone-dominated ("Dexterville"-like) intervals alternate with three mudstone-dominated intervals (see Fig. 8). Coquinites and shell pavements are more abundant and fossil diversity generally increases upward. Distinctive fossils include the large rhynchonellid brachiopod Paurorhyncha newberryi which is abundant in a thick, upper Ellicott siltstone-dominated interval along a west-tributary of Chautauqua Creek, and the distinctive carbonaceous fossil Foerstia (Protosalvinia) which is abundant at the 1310-1330 foot elevation of the main channel of Twentymile Creek and at the 1380-1400

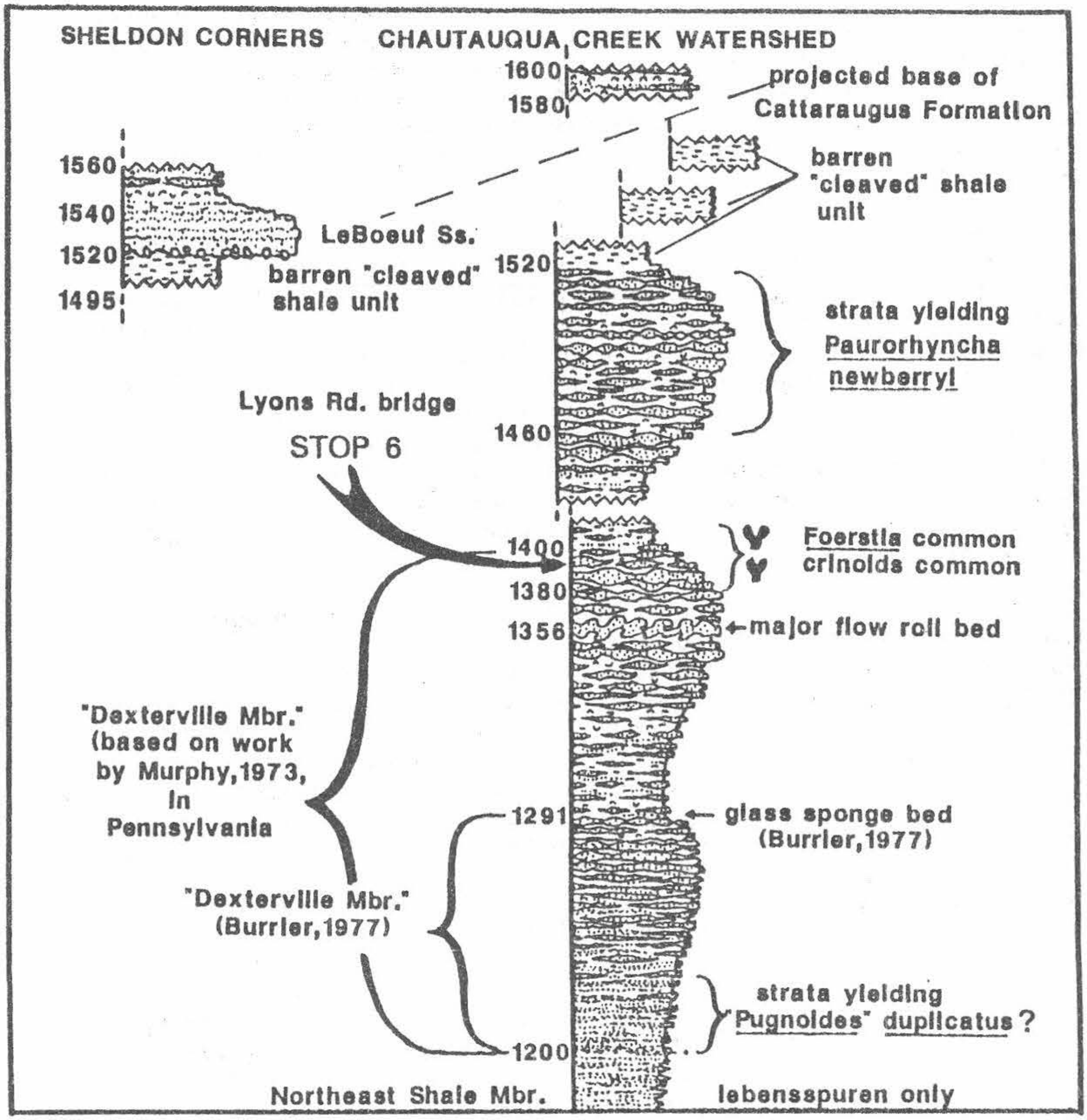


Fig. 8. Stratigraphy of Chadakoin Formation on Chautauqua Creek and its tributaries and also on nearby, small, unnamed, northwest-flowing tributory of Twentymile Creek, 0.7 mi. northeast of Sheldon Corners on the South Ripley 7.5 minute Quadrangle. Stratigraphic reconstruction is based partly on work of Tesmer (1963) and Burrier (1977) but information above elevation of 1380 ft. is largely to entirely based on recent mapping by Baird (see text). Newly recognized divisions in this area (Foerstia zone, Paurorhyncha-rich siltstone unit, sparsely fossiliferous to unfossiliferous variably fractured("cleaved") shale unit, and LeBoeuf Sandstone) are all well developed in the Union City, Erie, and Albion areas in Pennsylvania. Elevations (in feet) are from 7.5 minute topographic sheets.

feet elevation along Chautauqua Creek (See Stop 6: Figs. 8, 9). Foerstia is a probable reproductive structure of a fucoid alga (see Phillips, et al., 1972; Schopf and Schwietering, 1976). It is an extremely important and widespread zonal marker for North American Famennian deposits.

In the process of tracing the zone of abundant Foerstia from Conneaut Creek in Ohio northeastward to the vicinity of Twentymile Creek near Ripley, Chautauqua County, New York, using the report of Murphy, (1973), as a guide, one of the present authors (Baird) discovered that the "Dexterville-Ellicott" boundary of Murphy (1973), is much higher stratigraphically than that of Tesmer (1963), and Burrier (1977) (Fig. 8). It appears that Murphy's (1973), "Dexterville-Ellicott" boundary, coincident with his zone of abundant Foerstia, projects into Chautauqua Creek just below the defunct Lyons Road Bridge (See Stop 6) more than 31 meters (100 ft.) above the "Dexterville-Ellicott" boundary identified by Burrier (1977). As such, the "Ellicott" of Burrier (1977) includes about 80 meters (260 ft.) of strata on Chautauqua Creek (three "shale" and two "siltstone" divisions with most of the top shale unit missing), and the "Ellicott" of Murphy, 1973, encompasses 47 meters (150 ft.) of section (two "shale" and one "siltstone" division) at this section (see Fig. 8). Given this discrepancy, the present authors currently refer to the Chadakoin Formation along the Girard, Pennsylvania-Chautauqua Creek transect as including a total of three mappable siltstone units (the lowest being Dexterville (sensu Burrier, 1977) and three "shale" units (the base of the middle "shale" corresponding to the zone of densest Foerstia concentrations). Moreover, the top "shale" unit, marked by a widespread erosional discontinuity at its base and coarse basal Cattaraugus deposits (LeBoeuf-Panama members) at its top, is found to contain sparsely fossiliferous facies very different from that of underlying Chadakoin beds. These units are currently informal divisions, but, with continued work, they may be assigned formal stratigraphic names.

These six divisions can be traced southwestward to Little Elk Creek near Girard, Pennsylvania and three have tentatively been followed into northeast Ohio. The coquinite-, echinoderm-, and Foerstia-rich interval, coinciding with the "Dexterville-Ellicott" boundary of Murphy (1973), in the Chautauqua Creek-Erie, Pennsylvania region (see Stop 6), has tentatively been located along Conneaut Creek, southwest of Conneaut, Ohio where Foerstia and a low diversity brachiopod assemblage occur in association with interbedded grey and black shales in facies reminiscent of the Northeast Member in New York. If this boundary is an isochron, which we believe it is, then it will be possible to study fossil associations across major facies belts along time-controlled paleoenvironmental gradients through this region.

Cattaraugus Formation

Overview. The highest Devonian formational division was proposed by Clarke, (1902), to include strata from the Panama Conglomerate up to the base of the Knapp Conglomerate of Mississippian age. As such, this unit encompasses approximately 210 meters (650 ft.) of section in Chautauqua County which includes quartz pebble conglomerates, quartz-rich, but often micaceous sandstones, marine shell coquinites at many levels, lenticular siltstone beds, and both green and red shales (Tesmer, 1963). Most Cattaraugus beds are variably marine but "red beds" are reported in the hills south of Jamestown and in the eastern part of Chautauqua County. These are too far to the southeast to be visited on the present field trip. The strata from the Panama-level upwards to the base of the Mississippian are largely correlative with the Chagrin Formation of northern Ohio (see Chadwick, 1925; Caster, 1934; Pepper et al., 1954; Tesmer, 1963).

Numerous conglomeratic divisions have been traced by various workers over the past century providing some potential for the establishment for internal subdivisions, but the stratigraphy is still imperfectly known. Outcrops along creeks and roads in areas underlain by Cattaraugus strata are usually extremely poor due to the tendency for massive beds to migrate into creek channels through the process of slump and creep (See Stop 8). The best Chautauqua County outcrops observed by the present authors are along small creeks in the highest drained elevations west of Cherry Creek and south of Sherman.

Panama Conglomerate Member (See Stop 8). At the base of the Cattaraugus Formation is the Panama Conglomerate, first named by Carll, (1880), for approximately 21 meters (70 ft.) of quartz pebble conglomerate and buff sandstone near Panama, New York (See Stop 8). The Panama Conglomerate appears to be lenticular and discontinuous in the county. The coarse nature of this unit and its discontinuous distribution suggests that it may locally fill channels. The Panama has been correlated with the Leboeuf Sandstone which marks the top of Chadakoin deposits in the Erie-Albion area in Pennsylvania (Caster, 1934; Tesmer, 1963, 1974). One of us (Baird) has recently located the basal Cattaraugus sandstone, corresponding to the LeBoeuf thickness and lithologic character, both at Sheldon Corners (Fig. 8) at 1520 feet elevation and east of Summerdale on a tributory of Wing Creek at 1600 feet, in western Chautauqua County; this unit both thickens and coarsens southeastward assuming the typical Panama conglomeratic sandstone appearance south of the hamlet of Stedman Corners.

PROBLEMS AND QUESTIONS

Work Agenda:

Key questions pertaining to the Dunkirk-through Cattaraugus stratigraphic succession include the following:

Canadaway Formation

1. What process or set of processes accounts for the thin black and grey shale alternations comprising the zebra facies? What is the temporal significance of these beds and the sharp boundaries between them; are the black bands deposited slowly or rapidly? If many black bands were rapidly deposited, how were they deposited and what consistency was the organic-rich, black mud when it was transported? Which beds are turbiditic, which were produced by bottom-current processes, and which accumulated through slow background deposition from suspension?

2. What events and processes produced thick, non-graded siltstone units such as the Laona and Shumla? Are such depositional events linked in a process sense to slump-sheet and ball-and-pillow units observed in the shallower facies of the overlying Chadakoin Formation? Could these slump and/or gravity-flow beds be linked to ancient disturbances such as major storms or seismic events?

3. How far east can the Corell's Point Bed be accurately traced? Is the anomolously high reported occurence of this unit at 965 feet on South Cattaraugus Creek by House (1968) correct? If it is correct, can the Laona siltstone be located on this creek using the Corell's Point Bed as a control? If the Corell's Point and Laona units are truly this elevated, what impact does this have on stratigraphic correlations of higher divisions in this region? What impact does it have on oil and gas drilling prospects-is this the surface expression of the Bass Island structural trend?

4. Can the thick Northeast Shale Member be subdivided into smaller mappable divisions? Are turbidite bundles in the Northeast lenticular, as predicted in some submarine fan models, or are they sheet-like and widespread?

Chadakoin Formation

1. How far to the east can the key Chadakoin faunal and lithologic markers be mapped; can the zone of concentrated *Foerstia* and the *Paurorhyncha newberryi*-rich interval be traced across Chautauqua County into Cattaraugus County. Do the major siltstone-rich intervals retain their identity across this same interval; do these ultimately change to quartz pebble-rich sandstone facies if traced far enough?

2. What is the character and extent of structural folding and/or faulting of surface rocks in Chautauqua County, particularly along the "Bass Islands" structural trend? Recent mapping of upper Chadakoin and lower Cattaraugus divisions by Baird reveals anomolously high elevations for the Panama Conglomerate between Stedman and Panama in an area of intense oil and gas-action. Does this fold have a northeast-southwest trend? Does it connect to the aforementioned region of anomolously elevated Canadaway strata southeast of Gowanda?

3. What is the paleoenvironmental explanation for the uppermost Chadakoin, sparsely-fossiliferous, characteristically sheared ("cleaved") siderite-rich, unnamed mudstone division? Is this a regressive, nearshore facies unit? Does it correlate to the "Tanner Hill" redbed tongue reported below the Panama Sandstone at Warren, Pennsylvania (See Caster, 1934).

4. What is the impact of the Frasnian-Famennian extinction event on the slope and platform facies of the Catskill Delta? Is there a major restructuring of platform communities between the Frasnian Wiscoy Member and the Famennian Caneadea-Ellicott sequence? Are there corresponding changes in the depositional regime?

Cattaraugus Formation

1. This formation is not well understood; a need exists for basic mapping of key beds, but this work should probably commence in the largely-equivalent Chagrin Shale in Ohio, where outcrops are larger and better, and proceed eastward into Pennsylvania and New York. It appears that lower Cattaraugus strata offer some potential for mapping in Chautauqua County, but upper Cattaraugus beds may be mappable only through combined field and subsurface methods.

2. What is the origin of the thick conglomerate units? Do these overlie unconformities produced by sea level lowstand events? If so, the conglomerates should be lenticular but regionally mappable as is indicated by previous workers. If sea level oscillations are temporally superimposed on the progradation process, is there a predictable shoreward facies trend in the regressive sandstone and conglomerate units that can be documented? Miller (1974), suggests that this is the case, but more work is clearly needed to document dynamic changes which should be visible and measurable. Can neap-spring tidal cyclicity be identified in the cross-bedding laminations within this unit?

REFERENCES CITED

Aigner, T., 1985, Storm depositional systems. Dynamic stratigraphy in modern and ancient shallow marine sequences: Lecture notes in Earth Science, V.3, Springer-Verlag, New York, Heidelberg, Berlin, 174 p.

Alvarez, L.W., Alvarez, W., Asaro, F., and Michel, H.V., 1980, Extraterrestrial cause for the Cretaceous-Tertiary extinction. Experimental results and theoretical interpretation: Science, V. 208, p. 1095-1108.

Babcock, L.E., 1982, Paleontologic and sedimentologic character of Corell's Point faunal assemblages (Upper Devonian; Famennian), Southwestern New York State (abstr.): Amer. Assoc. Petrol. Geologists Bull., V. 66, No. 8, p. 1164.

Baird, G.C. and Brett, C.E., 1986, Erosion on an anaerobic seafloor: significance of reworked pyrite deposits from the Devonian of New York State: Palaeogeog. Palaeoclim. Palaeoecol., V. 57, p. 157-193.

Baird, G.C., Brett, C.E., and Kirchgasser, W.T., 1988, Genesis of Black Shale-roofed discontinuities in the Devonian Genesee Formation: In McMillan, N.J., Embry, A.F. and Glass, D. J., eds., Devonian of the World: Volume II: Sedimentation, Canadian Soc. Petroleum Geologists, Calgary, p. 357-376.

Beck, L.C., 1840, Report of the Mineralogical and chemical department of the survey of New York. New York Geol. Surv. Ann. Rpt., V. 4, p. 57-58.

Bowen, Z.P., Rhoads, D. C., and McAlester, A.L., 1974, Marine benthic communities in the Upper Devonian of New York: Lethaia, V. 7, p. 93-120.

Brett, C.E. and Baird, G.C., 1982, Upper Moscow-Genesee stratigraphic relationships in western New York: evidence for erosive bevelling in the Late Middle Devonian. <u>In</u> Buehler, E.J., ed., Field Trip Guidebook, 54th Meeting, New York State Geological Assoc., Buffalo, New York, p. 19-63.

Brett, C.E., Speyer, S.E. and Baird, G.C., 1986, Storm-generated sedimentary units: tempestite proximality and event stratification in the Middle Devonian Hamilton Group of New York: <u>In</u> Brett, C.E., ed., Dynamic stratigraphy and depositional environments of the Hamilton Group (Middle Devonian) in New York State, Part I. New York State Museum Bull., V. 457, p. 129-156.

Broadhead, R.F., Kepferle, R.C., and Potter, P.E., 1982, Stratigraphic and sedimentological controls of gas in shale-example from the Upper Devonian of northern Ohio. American Assoc. Petroleum Geologists, V. 66, p. 10-27.

Buehler, E.J. and Tesmer, I.H., 1963, Geology of Erie County, New York: Buffalo Soc. Natural History, Bull. V. 21, No. 3, 118 p.

Burrier, D. G., 1977, The paleoecology of the Chadakoin Formation of Chautauqua County (unpublished masters' thesis): S.U.N.Y. College, Fredonia, New York, 129 p.

Butts, C., Pre-Pennsylvanian stratigraphy of Pennsylvania: Pennsylvania Topographic and Geol. Surv. Rept. for 1906-1908, p. 190-204.

Byers, C. W. 1977, Biofacies patterns in euxinic basins; a general model: In Cook, H.E. et al., eds., Deep-water carbonate environments. Soc. Econ. Mineralogists Paleontologists Spec. Pub., V. 25, P. 5-17.

Carll, J.F., 1880, The geology of the oil regions of Warren, Venango, Clarion, and Butler counties: 2nd Pennsylvania Geol. Surv. Rept. 13, 58 p.

Caster, K.E., 1934, The stratigraphy and paleontology of northwestern Pennsylvania; Part 1, Stratigraphy: Bull. American Paleont. B. 21, No. 71, p. 19-37.

Chadwick, G. H., 1923, Chemung stratigraphy in western New York: Abstr. Geol. Soc. America Bull., V. 34, p. 68-69.

Chadwick, G. H., 1924, The stratigraphy of the Chemung Group in western New York: New York State Museum Bull., V. 251, p. 149-157.

Chadwick, G. H. 1925, Chagrin Formation of Ohio. Geol. Soc. America Bull., V. 36, p. 455-464.

Chadwick, G. H., 1933, Great Catskill Delta, and revision of Late Devonian succession: Pan-American Geologist, V. 60, p. 91-107, 189-204, 275-286, 348-360.

Clarke, J.M., 1902, Preliminary statement of the paleontological results of the areal survey of the Olean Quadrangle: New York State Museum Bull., V. 52, p. 524-528.

Clarke, J.M., 1903, Classification of the New York Series of geologic formations: Univ. State New York Handbook, V. 19, 1st edition.

Cooper, G. A., and others, 1942, Correlation of the Devonian sedimentary formations of North America: Geol. Soc. America Bull., V. 53, p. 1729-1794.

Cuomo, M.C. and Bartholemew, P.R., 1987, Identification of modern and ancient pellets using electron microscopy and microanalysis. Abstr. Geol. Soc. America, p. 63.

Cuomo, M.C. and Rhoads, D.C., 1987, Biogenic sedimentary fabrics associated with pioneering polychaete assemblages: Modern and ancient. Jour. Sed. Petrology, V. 57, No. 3, p. 537-543.

de Witt, W., Jr. and Colton, G., 1953, Bedrock geology of the Silver Creek quadrangle, New York: U.S. Geological Surv., Geol. Quad. Map GQ30.

Ettensohn, F. R., 1985, The Catskill Delta Complex and the Acadian Orogeny: A model: In Woodrow, D. L. and Sevon, W. D., eds., The Catskill Delta. Geol. Soc. America. Spec. Publ. No. 201, p. 39-50.

Ettensohn, F. R. et al., 1988, Characterization and implications of the Devonian-Mississippian Black-Shale sequence, eastern and central Kentucky, U.S.A.: Pychoclines, transgression, regression, and tectonism: In McMillan, N.J., Embry, A.F., and Glass, D.J., eds., Devonian of the World: Volume I: Regional Synthesis: Canadian Soc. Petroleum Geologists, Calgary, p. 323-345.

Gilman, R.A. and Metzger, W. J., 1967, Cone-in-cone concretions from western New York: Jour. Sed. Petrology, V. 37, No. 1, p. 87-95.

Goodfellow, W. D., Geldsetzer, H.H.J., McLaren, D. J., Orchard, M.J. and Klapper, G., 1988, The Frasnian-Famennian extinction: current results and possible causes, In McMillan, N.J., Embry, A.F. and Glass, D.J., eds., Devonian of the world: Volume III: Canadian Soc. Petroleum Geologists, Calgary, p. 9-22.

Hall, J., 1843, Geology of New York: Part IV, Comprising the survey of the Fourth Geological District: Albany, New York, 525 p.

Hartnagel, C.A., 1912, Classification of the geologic formations of the State of New York: New York State Museum Handbook V. 19, 2nd Edition.

House, M.R., 1962, Observations on the ammonoid succession of the North American Devonian Journal Paleontology, V. 36, p. 247-284.

House, M. R., 1965, A study in the Tornoceratidae: the succession of Tornoceras and related genera in the North American Devonian. Phil. Trans. Royal Soc. London, Ser. B, V. 250, p. 79-130.

House, M.R., 1966, Goniatite zonation of the New York State Devonian: In Buehler, E.J., ed., Field Trip Guidebook, 38th meeting New York State Geol. Assoc., Buffalo, New York, p. 53-57.

House, M. R., 1968, Devonian ammonoid zonation and correlation between North America and Europe: In Oswald, D.H., ed., International Symposium on the Devonian System, Calgary, Alberta Soc. Petroleum Geologists, V. II, p. 1061-1068.

House, M. R., Gordon, M., Jr., and Hlavin, W. J., 1986, Late Devonian amononoids from Ohio and adjascent States: Jour. Paleontology, V. 60, No. 1, p. 126-144.

Kidwell, S.M., 1986, Taphonomic feedback in Miocene assemblages: testing the role of dead hard parts in benthic communities: Palaios, V. 1, No. 3, p. 239-255.

Kidwell, S.M. and Aigner, T., 1985, Sedimentary dynamics of complex shell beds: implications for ecologic and evolutionary patterns, In Bayer, U. and Seilacher, A., eds., Sedimentary and evolutionary cycles, Springer Verlag, Berlin, p. 382-395.

Kirchgasser, W. T., 1974, Notes on the ammonoid and conodont zonations of the upper Devonian of southwestern New York, In Peterson, D.N., ed., Guidebook: Geology of Ne York State, 46th Meeting: Fredonia, New York State Geological Assoc., p. B9-B13.

McGhee, G.R. Jr., 1982, The Frasnian-Famennian extinction event: a preliminary analysis of Appalachian marine ecosystems. Geol. Soc. America Spec. paper, No. 190, p. 491-500.

McGhee, G.R., Jr., 1988. Evolutionary dynamics of the Frasnian-Famennian extinction event. In McMillan, N.J., Embry, A.F. and Glass, D. J., eds., Devonian of the World: Volume III: Paleontology, Paleoecology, and Biostratigraphy: Canadian Soc. Petroleum Geologists, Calgary, p. 23-26.

Miller, W. H., 1974, Petrology of Devonian Cattaraugus Formation and related conglomerates, Cattaraugus and Chautauqua counties, New York (Unpubl. master's thesis): State University of New York at Buffalo.

Murphy, J. L., 1973, Protosalvinia (Foerstia) zone in the Upper Devonian sequence of eastern Ohio, northwestern Pennsylvania, and Western New York. Geol. Soc. America Bull., V. 84, p. 3405-3410.

Pepper, J. F. and de Witt, W., Jr., 1950, Stratigraphy of the Upper Devonian Wiscoy Sandstone and the equivalent Hanover Shale in western and central New York: U.S. Geol. Surv. Oil and Gas Invest., Chart 37.

Pepper, J. F. and de Witt, W., Jr., 1951, Stratigraphy of the Upper Devonian Perrysburg Formation in western and west-central New York. U.S. Geol. Surv. Oil and Gas Invest. Chart OC-45.

Pepper, J. F., de Witt, W., and Demerest, D. F., 1954, Geology of the Bedford Shale and Berea Sandstone in the Appalachian Basin: U.S. Geol. Surv. Prof. paper, V. 259, 111 p.

Phillips, T. L., Niklas, K.J., and Andrews, H.N., 1972, Morphology and Vertical distribution of Protosalvinia (Foerstia) from the New Albany Shale (Upper Devonian): Rev. Paleobotany and Palynology, V. 14, p. 171-

Rhoads, D. C., and Morse, J.W., 1971, Evolutionary and ecological significance of oxygen-deficient marine basins. Lethaia, V. 4, p. 413-428.

Rickard, L.F., 1975, Correlation of the Silurian and Devonian rocks of New York State (Map. Chart Serv.): New York State Mus. Sci. Serv., No. 24, 16 p.

Sames, C. W., 1966, Morphometric data of some recent pebble associations and their application to ancient deposits: Jour. Sed. Petrology, v. 36, No. 1, p. 126-142.

Savrda, C.E. and Bottjer, D. J., 1987, The exaerobic zone, a new oxygen-deficient marine biofacies: Nature, V. 327, p. 54-56.

Schopf, J. M. and Schwietering, J.F., 1976, The Foerstia zone of the Ohio and Chattanooga Shales: U.S. Geol. Survey Bull., p. 1294-

Schwimmer, B.A. and Feldmann, R.M., 1990, Stratigraphic distribution of brachiopods and bivalves in the Upper Devonian (Famennian) Chagrin Shale in the Cuyahoga River Valley, northeast Ohio: Kirtlandia, No. 45, p. 7-31.

Scotese, C.R., Van der Voo, R., and Barrett, S. F., 1985, Silurian and Devonian base maps: In Chaloner, W. G. and Lawson, J. D., eds., Evolution and environment in the Late Silurian and Early Devonian. Phil. Trans. Royal Soc. London, V. B309, p. 57-77.

Stanley, D. J., 1987, Turbidite to current-reworked sand continuum in Upper Cretaceous rocks, U.S. Virgin Islands: Marine Geol., V. 78, p. 143-51.

Sutton, R.G., Bowen, Z. P., and McAlester, A.L., 1974, Marine shelf environments of the Upper Devonian Sonyea Grove of New York: Geol. Soc. America Bull., V. 81, p. 2975-2992.

Sutton, R.G., and McGhee, G. R., Jr., 1985, The evolution of Frasnian marine "community-types" of south-central New York: In Woodrow, D. L. and Sevon, W.D., eds., The Catskill Delta, Geol. Soc. America Spec. Publ., V. 201, p. 211-224.

Tesmer, I.H., 1963, Geology of Chautauqua County, New York: Part I, Stratigraphy and Paleontology (Upper Devonian), New York State Museum Bull., No. 391, 65 p.

Tesmer, I.H., 1974, A brief description of Upper Devonian units to be observed on Chautauqua County Field Trip. In Peterson, D.N., ed., Guidbook: Geology of New York State, 46th Meeting: Fredonia, New York State Geological Assoc., p. B1-B8.

Tesmer, I.H., 1975, Geology of Cattaraugus County, New York. Buffalo Soc. Natural Sciences Bull., V. 27, 105 p.

Tesmer, I.H., 1989, History of geology of westernmost New York State (1604-1899): Omni Press, Madison Wisconsin, 214 p.

Thayer, C. W., 1974, Marine paleoecology in the Upper Devonian of New York: Lethaia, V. 7, p. 121-155.

Weidner, W.E. and Feldmann, R.M., 1983, Paleoecological interpretation of echinocarid arthropod assemblages in the Late Devonian (Famennian) Chagrin Shale; northeastern Ohio: Jour. Paleontology, v. 59, p. 986-1004.

White, I.C., 1881, The geology of Erie and Crawford Counties: Pennsylvania Geol. Surv. 2d, Q4, 406 p.

Wignall, P.B., 1990, Observations on the evolution and classification of dysaerobic communities: In Miller, W., III, ed., Paleocommunity temporal dynamics: The long-term development of multispecies assemblies. Paleontological Soc. Spec. Pub., No. 5, p. 99-111.

Woodland, B.G., 1964, The nature and origin of cone-in-cone structure: Fieldiana: No. 4, p. 187-305.

Woodrow, D. L., 1985, Paleogeography, paleoclimate, and sedimentary processes of the Late Devonian Catskill Delta: In Woodrow, D. L. and Sevon, W. D., eds., The Catskill Delta. Geol. Soc. America Spec. Papr. No. 201, P. 51-64.

Woodrow, D. L., Dennison, J.M., Ettensohn, F.R., Sevon, W.T., and Kirchgasser, W. T., 1988, Middle and Upper Devonian stratigraphy and paleogeography of the central and southern Appalachians and eastern midcontinent, U.S.A.: In McMillan, N. J. Embry, A.F., Glass, D.J., eds., Devonian of the world: Volume I: Regional synthesis: Canadian Soc. Petroleum Geologists, Calgary, p. 277-304.

ROAD LOG
DEVONIAN STRATA AND PALEOENVIRONMENTS:
CHAUTAUQUA COUNTY REGION: NEW YORK STATE

TOTAL MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Leave the Fredonia campus at the Temple Street exit. Turn left (south) onto Temple Street.
0.7	0.7	Intersection with Route 20. Proceed straight ahead on Water Street.
0.8	0.1	Cross Canadaway Creek. Gowanda Shale exposed in vicinity of bridge.
0.9	0.1	Bear left at intersection of Water and Liberty Streets.
1.3	0.4	Several levels of river terraces are exposed on each side of bus at this point associated with earlier and higher stages of Lake Erie.
1.8	0.5	Cross Canadaway Creek bridge. Good exposures of the upper part of the Gowanda Shale by bridge are continuous up to falls at Laona.
2.2	0.4	Cross railroad tracks.
2.4	0.2	Enter Laona, New York. Well defined stream terraces can be seen on the right side of the bus. An abandoned oxbow lake can be observed 8 meters (25 ft.) above the present level of the creek.
2.6	0.2	Intersection with Webster Road. Park cars on pull-off to right just before this intersection.

STOP 1. LAONA SILTSTONE MEMBER

Laona Siltstone over Gowanda Shale at waterfall 200 feet southwest of car park at bridge (Overlook Stop Only; we will be able to sample from this division at STOP 4). Note the sharp contrast between the conspicuously banded ("zebra" facies) of the Gowanda and the massive nature of the overlying Laona. The Laona and the higher Shumla siltstone are believed to be gravity-flow units which record downslope movement of silt or fine sand into the anoxic to minimally dysoxic lower slope-basin setting recorded by the Gowanda-Westfield-Northeast shale members.

Total Miles	Miles From Last Point	Route Description
5.2	2.6	Return to Temple Street entrance to the Fredonia campus, but proceed straight (north) on Temple Street towards Dunkirk.
5.6	0.4	Y-Junction; turn right (north) on Brigham Road.
6.1	0.5	Cross New York State Thruway
6.7	0.6	Enter City of Dunkirk
6.9	0.2	Pass Al-Tech Corp. factory (to right).
8.1	1.2	Junction with Route 5; turn left (towards southwest).
8.3	0.2	Turn right onto Point Drive North.
8.7	0.4	Turn right onto Cedar Beach parking area.

STOP 2: HANOVER SHALE-DUNKIRK SHALE CONTACT AT POINT GRATIOT

Point Gratiot is the type section for the Dunkirk Shale Member. Along the northeast side of the point the lower part of the Dunkirk is visible as is the uppermost 1 to 2 meters of the underlying Hanover Shale Member (Fig. 4A). The bioturbated grey Hanover contrasts strongly with the laminated brownish black beds of the Dunkirk. Pyritic burrow tubes are conspicuous in the uppermost Hanover and in the thin, 13.5 cm (4.5 in.), recurrent Dunkirk grey mudstone interval above the basal 15-17 cm (6.0-6.5 in.)-thick basal Dunkirk black bed, but no body fossils occur in the grey lithology except for pyritic goniatite steinkerns and wood debris; this grey sequence is classic for slowly deposited dysoxic facies (Rhoads and Morse, 1971; Byers, 1977).

The Dunkirk Shale is non-bioturbated at most levels but there is some evidence of internal burrowing as can be seen in the basal black bench. Body fossils in the Dunkirk are usually uncommon; the linguloid brachiopod *Barroisella campbelli* occurs at some levels and both wood debris and fish fragments have been found. The top of the basal black bench is particularly good for carbonized logs and fish debris (Fig. 4A); when viewing is optimal, a thirty foot-long tree trunk is visible on the surface. In the spring of 1986 one of us (Baird) and several S.U.C. Fredonia geology students recovered a partial skeleton of the giant arthrodire *Dunkleosteus* from this bed.

Although a bioturbated, conformable contact between Hanover and Dunkirk can be seen below the basal Dunkirk black bed which dips southward below the water at this section, the top of the 13.5 cm Dunkirk grey mudstone bed marks a submarine erosion surface which is "roofed" by black Dunkirk deposits; this contact is marked by abundant

Total Miles	Miles from last point	Route Description

STOP 2, continued

detrital pyrite in the form of reworked burrow tube fragments, nodules, and occasional goniatite steinkerns derived from the underlying grey mudstone interval which is conspicuously rich in pyritic tubes and nodules (Baird and Brett, 1986). Additional reworked components include fish bones, conodonts, and carbonized wood fragments. The abundance of wood and conodonts in the basal few centimeters of the overlying black shale suggests that this interval is unusually condensed, recording the descent of numerous logs from the water surface over a long period of time. This wood-rich unit also records bioturbation, which, at this section, was sufficient to flux the broken pyritic tubes into random orientations. Bottom erosion is believed to be due to sediment starvation on the seafloor during relative sea-level rise. Scouring of the substrate may have been accomplished by rare deep storm waves, bottom currents in the basin, or internal waves impinging a regionally sloped substrate (Baird and Brett, 1986; Baird et al., 1988). Exhumed and concentrated detrital pyrite was preserved (not oxidized) in the anoxic to minimally dysoxic Dunkirk bottom setting.

In the early 1980s George McGhee and Edward Olsen sampled this section for the occurrence of iridium, a trace element believed by some to mark catastrophic extraterrestrial collision events in the rock record (see Alvarez et al., (1980). The Late Devonian Frasnian-Famennian ("F-F") extinction event, one of five greatest Phanerozoic extinctions, was, until recently, believed to be coincident with the base of the Dunkirk. This event is similarly thought by some to be the possible record of an impact (Goodfellow, et al., 1988). However, no Iridium-rich horizon was found (McGhee, 1982, 1988). Recent bistratigraphic work indicates that the F-F boundary is somewhere within the upper Hanover (Klapper, Kirchgasser pers. comm.) at a level which has not been precisely identified.

Note the "bedded" till above the Dunkirk and the striated bedrock contact below this "till."

		Leave parking lot. Turn left (south) onto Point Drive North.
9.1	0.4	Intersection with Route 5. Turn right (to southwest).
10.1	1.0	Cross Canadaway Creek.
13.7	3.6	County Fly Ash Dump to the left (see Field Trip G, this volume).
14.8	1.1	Bridge over Little Canadaway Creek.
15.9	1.1	Entrance to Lake Erie State Park.

Total Miles	Miles from last point	Route Description
17.9	2.0	Bridge over Slippery Rock Creek. Exposures include the South Wales Member at the lake shore level and Gowanda Member at the bridge and upstream.
18.8	0.9	Bridge over Corell Creek. Exposures are in the Gowanda Shale Member. The Corell's Point pyrite-goniatite bed is exposed approximately 120 feet upstream from the bridge.
19.9	1.1	Turn right (northwest) onto property of trailer court. Park at beach by boat ramp and proceed on foot across small creek and along shore for approximately 100 yards.

STOP 3: CORELL'S POINT PYRITE-GONIATITE BED IN GOWANDA MEMBER

The Corell's Point Pyrite-Goniatite Bed, encompassing two regionally mappable levels of calcareous septarian concretions, is present at several creek localities in this county but this shore exposure is the best place to examine fossils and sedimentary structures (Fig. 4 B,C). At least two, laterally discontinuous beds rich in pyrite nodules and locally abundant pyritized cephalopod steinkerns can be traced within the lower septarian concretion zone along the shore and around the small headland. Goniatites include *Cheiloceras amblylobum*, *Tornoceras concentricum*, and *Aulatornoceras bicostatum* (Fig. 4C); these belong to the zone of *Cheiloceras* (II) in the Famennian (see House, 1966; Kirchgasser, 1974). Orthoconic cephalopods are also common; these are commonly encrusted by a reptate auloporid coral. Slightly-curved conical shells less than one inch in length may be bactritid cephalopods or coleolid tubes. Bivalves, including *Lunulicardium eriense*, *Praecardium multicostatum* and *Loxopteria corrugata* occur with the cephalopods but these are usually preserved as non-pyritic composite molds often with a faint organic patina which may be a remnant of the periostracum layer. Driftwood, usually partly carbonized and partly permineralized by pyrite, occurs with the other fossils. Spectacular large *Zoophycos* spreiten in the Corell's Point Bed indicate that this trace had become important in offshore, dysoxic facies by the Famennian.

Babcock (1982), believed that the fauna of this bed was selectively preserved by turbiditic smothering events; some beds in this unit have shallow sole marks and display lamination similar to those in flaggy siltstone beds elsewhere in the Canadaway Formation. However, evidence of periods of reduced sediment influx is shown by intense bioturbation at some levels and by the tendency for auloporid corals to not only colonize partly-buried cephalopods but to extend colonial growth onto the adjacent seafloor. The history of this unit is complex and the presence of so many fossils at this level is suggestive of an episode of increased bottom oxygenation and reduced average turbidity.

Total Miles	Miles from last point	Route Description

STOP 3, continued.

The Corell's Point Bed is well exposed in Corell's Creek, Slippery Rock Creek near Brocton, Little Canadaway Creek near Lamberton, Canadaway Creek upstream from Route 20, and Walnut Creek at Forestville (House, 1966, 1968).

Examine the Gowanda strata both below and above the Corell's Point Bed; notice the numerous brown-black shale beds which alternate with grey-green mudstone to produce the conspicuous striped banding along the shore (Fig. 4 B). This basinal "zebra" facies will be discussed in greater detail at STOP 4.

		Return to the bus. Leave trailer park and turn right (southwest) onto Route 5.
21.1	1.2	To the left (south) observe the elongate glacial landforms which characterize this specific part of the lake plain. These deposits may represent reworked sand from older sand deposits of higher lake stages.
21.6	0.5	Bridge over unnamed creek south of golf club. Exposures of Gowanda Shale include a 4.5 m (15 ft.) waterfall at the lake and a conspicuous fossil wood concentration in a siltstone bed between the falls and Route 5.
23.6	2.0	Bridge over Bournes Creek. Park on shoulder and enter property on left (south) of Route 5.

STOP 4: GOWANDA, LAONA, AND WESTFIELD MEMBERS

This outcrop on Bournes Creek affords an opportunity to examine the Laona Siltstone over the Gowanda Shale and to sample the lower part of the still-higher Westfield Shale upstream from the waterfall. The Laona is expressed here as three massive, but relatively thin, siltstone beds which are separated by intervals of interbedded black and green shale ("zebra" facies). It is much more poorly developed here than at STOP 1 or in creeks between here and Barcelona (see road log below); apparently the Laona gravity-flow event involved channelization of the mobile silt such that this unit develops the thick, massive character only locally. Here it appears to be represented by three discrete flow events which must have occurred at separate times.

Total Miles	Miles from last point	Route Description

STOP 4, continued.

We will hike upstream to examine the lower Westfield section which is characterized by good development of the finely interbedded black and green-grey shale depoits ("zebra" facies) that are characteristic of much of the Canadaway succession (Figs. 6, 7). The black shale bands, typically less than one inch-thick, are usually laminated and they can be quite silty; some appear to grade laterally into siltstone beds, commonly containing wood debris. Grey mudstone beds sometimes pinch out laterally allowing black beds to merge. The grey mudstone deposits are fine grained and minimally bioturbated such that they appear structureless on a fresh break surface. The grey beds appear to be composed of turbiditic mud; there appears to be a spectral gradient from the grey mudstone layers through silty mudstone beds with incipient sole mark development to cross-laminated siltstone beds with good sole marks; at an initial glance, this would appear to be reasonable suite of bed-types that one might expect to see in an ancient prodelta slope environment where turbiditic silt was scarce with respect to mud.

However, complex internal cross-lamination of many siltstone beds suggests that processes other than turbidity currents may have produced them and that some beds may record multiple depositional events (see Stanley, 1987). Moreover, it appears that, at least, some of the thin black shale beds do not simply record stagnant anoxia and slow accumulation of organic-rich mud. Beneath many, if not most, thin black shales beds is a millimeter-thick zone of pyritic microspheres and partial microspheres which appear to have formed in situ within the uppermost, water-rich grey mud layer (Figs. 6, 7). Most of these are partial spheres with brilliant curved exteriors and flat tops marked by dull polyframboidal granular surfaces (Fig. 7). Some bear crenulated shiny exteriors reflecting compressive burial affects following formation of the pyrite.

These sand-size microspheres apparently record early diagenetic, bacterially-mediated pyrite formation within spherical spaces. We suspect that surfaceward migration of CO_2 or methane may have produced bubbles which became trapped under the black mud layers or under bacterial mats associated with the black muds (Fig. 7). Sulfidic activity resulted in the partial-to-complete filling of the bubbles leaving the partially filled ones as useful geopetal indicators.

Within the basal few millimeters of many black shale beds, the microspheres are again visible. However, partial microspheres are randomly reoriented, indicating that they have been disturbed after formation. Evidently microbioturbation during black mud deposition served to reorient the pyrite in some instances. In other black shales, the microspheres are concentrated in lag concentrations; this suggests that these black shale layers may have been deposited under the influence of tractional currents. Cuomo and Bartholemew (1987) and

Total Miles	Miles from last point	Route Description

STOP 4, continued

Cuomo and Rhoads (1987), have observed that both modern and ancient black muds are extensively pelletized by bottom organisms. If this is true, then the pelletized black mud may have been easy to mobilize through current action because it would have been like "sand" or "silt". It is then possible that some of the thin black bands in the Canadaway Formation may be analogous to some of the tractional siltstone beds to which they sometimes laterally intergrade. Study of these complex and somewhat enigmatic deposits remains an on-going project.

In-situ microspheres are abundant beneath some of the black bands at this stop. Reoriented and hydraulically concentrated microspheres are locally common along the soles of siltstone beds where they have collected into scoured depressions, (corresponding to "positive" flutes, grooves, and excavated burrows) along such surfaces.

Total Miles	Miles from last point	Route Description
		Return to bus and proceed ahead (to SW) on Route 5.
24.6	1.0	Unnamed creek to right (north). Thick Laona section is present downstream from road.
26.0	1.4	Enter Barcelona, New York.
26.1	0.1	Pass Barcelona Harbor complex. Excellent shore exposures into the Gowanda Shale are developed north of dock area. This is a particularly clean section for observing "zebra" facies. Stone lighthouse was first to be illuminated by natural gas in the U.S.A.
26.2	0.1	Intersection of Route 5 and Route 394. Turn left (south) onto Route 394.
26.4	0.2	Pass Thruway entrance on left.
26.7	0.3	Entering Westfield.
27.7	1.0	Intersection of Route 394 and 20 in Westfield. Continue straight towards Mayville.
28.2	0.5	Crossing Glacial Lake Whittlesey beach berm.
28.6	0.4	Leave Westfield.
28.8	0.2	Cross Little Chautauqua Creek Gorge.
29.0	0.2	Turn right onto Gale Road.

Total Miles	Miles from last point	Route Description
29.45	0.45	Turn right onto car park before bridge and disembark.

STOP 5: NORTHEAST SHALE ALONG CHAUTAUQUA CREEK

At this brief stop we will examine a cutbank section and associated loose siltstone slabs which are typical of the Northeast Member. The Northeast is a typically "flaggy" sequence of thin grey siltstone beds interbedded with silty grey mudstone. Some siltstone layers appear to be turbidites which lack the basal graded division of the typical Bouma-sequence. Many beds display good sole marks with groove- and tool-marks. Networks of the trace fossil *Planolites* are common, and a trace resembling *Zoophycos* occurs more rarely. Some sole marks show hydraulic concentrations of exhumed and transported pyritic microspheres and partial microspheres in groove- and flute-casts; these were current-scoured depressions into which the coarser and heavier pyrite material selectively accumulated as currents waned. Cone-in-cone concretions occur at several levels within the Northeast Shale (Woodland, 1964; Gilman and Metzger, 1967); fragments of such concretions are occasionally observed in creek rubble at this stop. *In-situ* shelled fossils are rare to absent here as at many other sections of the lower Northeast Member. This facies records a dysoxic prodelta setting below the reach of most storm waves.

		Return to bus. Turn left from carpark onto Gale Road.
29.9	0.45	Junction with Route 394; turn right (south) towards Mayville.
31.4	1.5	Start to cross lip of prominent lake escarpment marking edge of Allegheny uplands.
34.4	3.0	Enter Mayville.
34.8	0.4	Junction of Route 394 and Route 430. Turn right (southwest) onto Route 430.
38.9	4.1	Turn right (west) onto Nettle Road (turn onto Nettle Rd. coincides with left (southward) turn of Route 430 towards Sherman).
39.0	0.1	Turn right (northwest) from Nettle Road onto Lyons Road. Group of buildings is hamlet of Summerdale.
39.1	0.1	Cross Summerdale Road.

Total Miles	Miles from last point	Route Description
40.0	0.9	Chautauqua Creek State Forest to right.
40.6	0.6	Lyons Road dead end at Chautauqua Creek (bridge is out). Park bus and disembark. LUNCH STOP.

STOP 6: FOERSTIA (PROTOSALVINIA) ZONE WITHIN ELLICOTT SHALE MEMBER

Exposures of the Chadakoin Formation on Chautauqua Creek are the most complete in New York; the lower division (Dexterville Siltstone Member) is entirely exposed, and approximately 75 to 80 percent of the overlying Ellicott Shale Member can be observed on this creek and along its upper tributaries (Fig. 8). Recent discovery of about 38 meters (120 ft.) of additional measureable Ellicott in a side tributory bordering Lyons Road allows us to observe a major siltstone division yielding the distinctive large rhynchonellid brachiopod Paurorhyncha newberryi (Figs. 8, 9): which has not been recorded prior to now within the county (Tesmer, 1963). In addition, discovery of the important algal taxon Foerstia (Protosalvinia) in the section at this stop (see text) allows us to establish equivalency of this outcrop with key parts of the Ohio, Antrim, New Albany, and Chattanooga black shales between New York and Oklahoma.

According to correlations and measurements of Tesmer (1963), and Burrier (1977), the Lyons Road section is within the Ellicott Shale approximately 33 meters (105 ft.) above the Dexterville Member. Mapping by Baird shows that the Dexterville-Ellicott boundary of Murphy (1973), corresponds to a level in Chautauqua County which is approximately 33 to 38 meters (105-125 ft.) above that of Burrier (see text). The Lyons Road section correlates to Murphy's Dexterville-Ellicott lithologic transition and the zone of abundant Foerstia that are characteristically associated with it; the boundary between Murphy's two members is approximately at the 1400 foot-elevation (Fig. 8, 9) which is below the old road shoulder (parking area for group) adjacent to the defunct bridge. This level also corresponds to an upward transition from siltstone-dominated "distal platform" Ellicott facies of Burrier (1977), to his mudstone-rich "prodelta" deposits at the upper end of his measured section. We will cross the mowed field area and collect from shell-rich, silty deposits near the boundary between the siltstone-dominated and upstream shale-dominated intervals.

Downstream from the bridge vicinity is siltstone-dominated, fossiliferous delta platform ("Chemung") facies of the Late Devonian (Rickard, 1975). The most conspicuous features are lenticular siltstone and fine sandstone beds interspersed with grey mudstone layers and coquinitic shell accumulations. Unlike the thinner and more continuously even-bedded slope and basin siltstone beds in the Canadaway Formation, these layers pinch and swell and there is abundant evidence of erosional truncation of older beds by younger ones. Storm layers dominate the section; tempestites, pararipples, occasional thick siltstone beds with hummocky cross-stratification, and gutter casts are

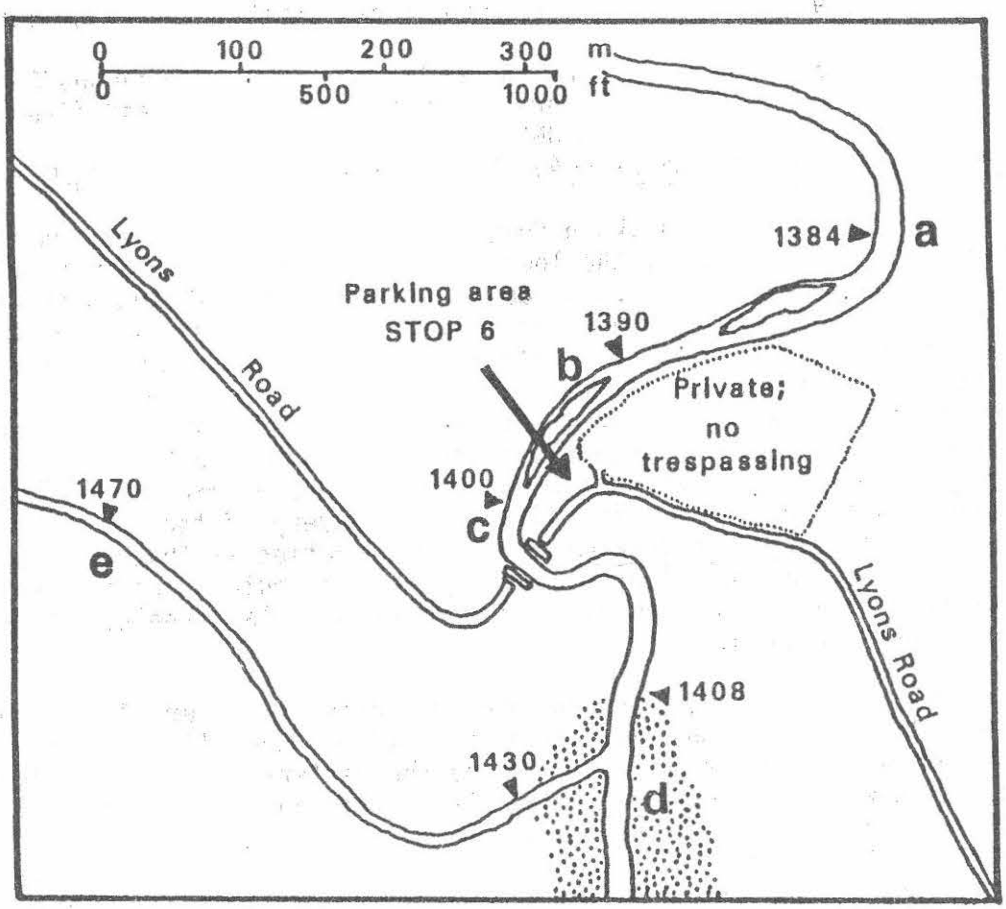


Fig. 9 Chautauqua Creek Devonian exposures in the vicinity of Lyon's Road crossing (see STOP 6). A) major flow roll bed; B) Interval of abundant Foerstia; B-C) top of siltstone-dominated interval corresponding to Dexterville-Ellicott boundary identified by Murphy (1973) in Pennsylvnia. Burrier (1977) interprets upward change from the siltstone-dominated unit to mudstone-dominated facies (at elevation 1395) as a transgressive change from "distal shelf" to "prodelta" conditions. The mudstone-dominated facies unit (C) near the defunct bridge yields echinoderms, Foerstia, and dictyosponges. D) No bedrock exposure; a Quaternary buried valley filled with stratified drift underlies the creek in this area; E) Strata yielding the large rhynchonellid brachiopod Paurorhyncha newberryi are exposed in large side tributary.

Total Miles	Miles from last point	Route Description

STOP 6, continued.

all evidence for the impingement of storm waves on the substrate. Good tempestites display a weakly graded concentration of reworked shells, overlain by laminated siltstone with low inclination cross-stratification displaying internal bedding discordances (see Aigner, 1985). Gutter casts are erosional runnels, usually less than a meter in length which often have interiors similar to tempestites; siltstone- and shell-filled gutters are the result of scour- and filling-processes involving both oscillatory and unidirectional currents during storms (Aigner, 1985). These runnels are presently casted with siltstone and coquinites and they display "tool marks" produced by the bouncing and rolling of shells, wood, and other debris during the storms. Sole marks under many siltstone beds show disarticulated brachiopod shells visible as concavities on the sole surface; this is the result of mud "sheltering" beneath the stationary shell during the storm which is subsequently buried by silt. Later differential weathering removes the sheltered mudstone leaving the shell as a negative feature on the siltstone surface. Chadakoin storm beds decrease in size and complexity downslope into Pennsylvania and Ohio along a depth-controlled proximality gradient (see Aigner, 1985); this gradient can be better assessed through refined stratigraphic mapping within this formation.

Not only are the thick proximal storm beds represented in this section but thin, mud-dominated layers produced by smaller storms are common, particularly upstream in the shaley Ellicott interval. These distal tempestites record the transport and settle-out of mud which often smothered bottom organism as currents dissipated (Brett, *et al.*, 1986). Excellent examples of articulated delicate fossils, including crinoids, echinoids, and glass sponges, found at the Lyons Road section attest to numerous bottom-smothering events recorded in this outcrop.

Abundant fossils belonging to a few standard brachiopod and bivalve genera can be collected here. Brachiopods including the rhynchonellid *Camarotoechia contracta*, the spiriferid *Cyrtospirifer nucalis*, and a productid *Productella speciosa* are ubiquitous, as are the bivalves *Mytilarca chemungencis* and *Leptodesma potens*. In particular, notice the partially dissolved character of the brachiopod shells. Evidently, carbonate undersaturation in the near surface muds and/or oxidation of pyrite in these same muds caused many of these shells to undergo variable amounts of dissolution. Fossils to look for include hexactinellid sponges, articulated inadunate crinouds and undescribed archaeocidaroid echinoids. Small fossils include the fucoid algal structure *Foerstia* and abundant black, chitinous polychaete jaw elements called scolecodonts. The best *Foerstia* can be collected from shelly beds in the vicinity of- and downstream from the junction point of the north-trending, gently sloped path with Chautauqua Creek (Fig. 9). These resemble 0.5 to 1.0 mm-diameter black "tar splatter" marks on bed surfaces which frequently are branched in a characteristic "Y" pattern. A few bedding surfaces are crowded with this important fossil.

Total Miles	Miles from last point	Route Description
		Return to bus. Head back uphill on Lyons Road to Summerdale.
42.2	1.6	Junction of Lyons Road with Nettle Road at Summerdale. Proceed straight (south) on Lyons Road.
42.3	0.1	Lyons Road converges into Route 430. Proceed south on Route 430 to Sherman.
46.2	3.9	Enter village of Sherman.
46.8	0.6	Junction of Route 430 with Route 76 in Sherman. Turn left (south) onto Route 76.
45.9	0.3	Junction Route 76 with New Route 17. Turn left (east) onto Route 17.
48.1	0.9	Cattaraugus exposure on left side of road.
49.5	1.4	Ellicott Shale on right and left side of road for next half mile. It is rich in sideritic lentils and characterized by some pyrite-rich mudstone beds, fossils are scarce.
52.2	2.7	Cross Prendergast Creek. Ellicott Shale along creek below bridge. *Foerstia*-bearing strata equivalent to the section at STOP 6 occur along this creek north of (downstream from) Route 17 above the Stedman-Sherman road crossing at an elevation of 1380-1390 feet.
52.5	0.3	Ellicott Shale on left and right side of road for next 0.25 mile. Stop vehicles at west (downhill) end of section.

STOP 7: STRATA OF UPPERMOST ELLICOTT MEMBER

This will be a brief stop to examine beds in the highest Chadakoin Formation. In the lower part of the cut and, particularly in the drainage ditch, one can observe brachiopod coquinites associated with storm (tempestite) beds. The characteristic upper Ellicott rhynchonellid brachiopod *Paurorhyncha newberryi* occurs at one or two levels in this roadcut. Of particular interest are shells stacked like dishes due to the oscillatory motion of storm-currents on the substrate (see Kidwell and Aigner, 1985; Kidwell, 1986). The elevation of the upper (eastern) end of this outcrop is approximately 10 to 15 meters (30-50 ft.) below the projected level of the Panama Conglomerate which can be seen on the hill immediately south of this area (Tesmer, 1963).

Total Miles	Miles from last point	Route Description
		Return to bus. Proceed east on Route 17.
53.0	0.5	Exit Route 17 for Route 32.
53.3	0.3	Turn right (south) on Route 32 towards Panama.
55.9	2.6	We are crossing the northern limit of the "Bass Islands" structural trend of drillers. The base of the Panama Conglomerate is anomolously elevated to approximately 1740 feet in this immediate area. Note abundant evidence of drilling activity (blue pumps and tanks) in nearby fields.
58.4	2.5	Enter Panama, N.Y.
58.8	0.4	Junction of Route 32 with Route 474 in Panama. Turn right (west) onto Route 474.
59.0	0.2	Junction of Route 474 with Rock Hill Road. Turn left up the hill towards Panama Rocks Park.
59.2	0.2	Enter Panama Rocks Park on left.

STOP 8: PANAMA CONGLOMERATE MEMBER OF CATTARAUGUS FORMATION

The Panama Conglomerate, named by Carll (1880), for 21 meters (70 ft.) of quartz-pebble conglomerate and quartz-dominated sandstone exposed at Panama, Chautauqua County, is a conspicuous lenticular unit at the base of the Cattaraugus Formation (see Tesmer, 1963, 1974). It is one of several conglomerate-rich units within the Cattaraugus Formation of Chautauqua and Cattaraugus counties. The Panama is correlative with the LeBoeuf Sandstone in Erie County, Pennsylvania and in western Chautauqua County.

The Panama Conglomerate Member is composed of cross-bedded, quartz pebble conglomerate, quartz-rich sandstone and sandstones with dilute (sand-supported) pebble concentrations. Conglomerates are composed mainly of quartz pebbles with minor components of jasper and metamorphic rock. Pebbles are usually less than one inch in diameter, well rounded, and often display a distinctly flattened discoidal to prolate shape. Rare fossils described from the Panama include *Cyrtospirifer chemungensis*, *Camarotoechia contracta*, *Ptychopteria sao*, as well as gastropods, (Tesmer, 1963; Miller 1974).

Miller (1974) interpreted the Panama to be a "beach" or "near beach" deposit with the main sediment source area to the southeast. One line of evidence he used was the tabular shape of many quartz pebbles in

STOP 8, continued.

the Panama as opposed to the more ovoidal pebble shapes he observed in the "fluviatile" post-Devonian Knapp and Olean conglomerates (see also work of Sames, 1966); abrasive imbricate stacking and sliding of pebbles in a beach-foreshore environment was cited to explain the tabular pebble condition. In addition, the distinctly bipolar (NW-SE) paleocurrent pattern for measured cross-bedding sets within Panama (Miller, 1974) appears to reflect tidal current processes acting in a coastal setting. The occurrence of brachiopods and marine mollusks in the Panama is further evidence of marine influence in this deposit.

A major question that geologists confront when examining deposits such as these is the problem of transporting and accumulating the vast numbers of quartz pebbles in this outcrop; clearly, a great volume of source terrain must have been denuded to produce it because vein quartz, chert nodules, and pegmatitic quartz make up only small volumes of normal lithosphere. This problem is partly obviated by the fact that the Panama Rocks outcrop is really more a thick lentil or channel feature which attenuates laterally to thin Panama facies dominated by sandstone (Tesmer, 1963; Miller, 1974). Hence, the volume of Panama quartz pebbles is really much less than this outcrop would initially suggest. Many of these pebbles may have been recycled from conglomerates in older Paleozoic units. Conglomerates are conspicuous features in the Ordovician Bald Eagle and Juniata divisions, the Silurian Oneida-Shwangunk and Tuscarora deposits, and in Middle through Upper Devonian proximal deltaic deposits of the Catskill Delta; Acadian overthrusting events probably exposed some or all of these eastermost sedimentary divisions to erosion, thus freeing up older pebbles for recycling. Post-Chadakoin sea level-fall and westward-northwestward advance of the paleoshoreline in western New York and northwest Pennsylvania probably served to introduce large numbers of quartz pebbles into the Chautauqua County region.

Many questions remain unanswered concerning these conglomerate units and the non-conglomeratic facies in between them. There are very few places where basal Cattaraugus (Venango) coarse facies can actually be seen in place rather than as slump blocks in creek beds or on wooded hillsides; it is essential to observe upper and lower contacts of units such as this to establish any sense of context for these units as geological events.

For those particularly interested in landscape and groundwater processes Panama Rocks offers a superb view of slope creep in action as well as weathering processes along joint systems developed in this deposit. The numerous "Dens" and "Alleys" between blocks also provide opportunity to examine the important role of "root pry" in forcing blocks apart. This park is one of several such tourist areas developed in conglomerate-rich deposits in this region.

Total Miles	Miles from last point	Route Description
		Return to bus. Return downhill on Rock Hill Road to Junction with Route 474.
59.4	0.2	Junction with Route 474. Turn right (east) into Panama.
59.6	0.2	Junction of Route 474 with Route 32 in Panama. Turn left (north) and proceed towards Chautauqua Lake.
65.3	5.7	Pass beneath new Route 17. Continue straight (north) on Route 32.
66.2	0.9	Pass through Stedman. Continue straight. over Kent end moraine.
67.3	1.1	Cross Prendergast Creek and continue over Kent ground moraine to vicinity of Chautauqua Lake.
69.0	1.7	Junction of Route 32 with Route 394 at Chautauqua Lake south shore. Turn left (northwest) onto Route 394 towards Mayville. For the next several miles we will traverse over Pleistocene and Recent alluvial sand and silt.
70.7	2.8	Leave Chautauqua Lake shore area and proceed uphill (northwest) into center of Mayville.
71.5	0.8	Junction of Route 394 and Route 430 in Mayville. Proceed straight on Route 394 to Westfield.
77.5	6.0	Junction of Route 394 and U.S. 20 in Westfield. Turn right (northeast) onto Route 20 and proceed to Fredonia.
79.3	1.8	Leave Westfield (commercial area bordering village).
81.1	1.8	Ascend Lake Warren beach berm. Rt. 20 follows the top of this beach ridge for the next 8 miles. Numerous vineyards and orchards are developed on the sandy soil, as are numerous sand and gravel pits.

Total Miles	Miles from last point	Route Description
83.6	2.5	Enter town of Portland.
84.7	1.1	Enter town of Brocton.
86.0	1.3	Leave town of Brocton.
88.2	2.2	Cross Little Canadaway Creek. Resistant siltstone beds within Gowanda Shale hold up waterfall 100 feet downstream from road.
91.3	3.1	Enter village of Fredonia.
91.8	0.5	Cross Canadaway Creek. Monument (enscribed boulder) to left marks position of the first successful gas well which was drilled in 1825).
92.0	0.2	Junction of Temple Street and U.S. 20. Turn left (northwest) onto Temple.
92.7	0.7	Temple Street entrance to State College: Fredonia campus.

END OF TRIP

GEOLOGY AND OIL AND GAS EXPLORATION IN WESTERN NEW YORK

Jerold C. Bastedo (Ecology and Environment, Inc.)
and
Arthur M. Van Tyne (Van Tyne Consulting)

HISTORY OF OIL AND GAS DRILLING IN WESTERN NEW YORK

SEEPS

Long before any drilling took place, local Indians had found natural oil and gas seeps at several places in western New York. The first recorded observation by white men of such an occurrence was in 1627. In a letter written in that year a Franciscan priest described his visit to the Cuba oil spring in Allegany County. He was taken there by a band of Indians who had used the oil at various times for medicinal, fuel and ceremonial purposes.

In 1669 Indians took the French explorer LaSalle and his party to a natural gas seep which they knew of in western Ontario County. While there, they lighted the gas and watched it burn. The event was recorded in a diary by one of the party and this account was later published.

FIRST WELLS

Possibly the first gas well drilled anywhere was located in the bed of Canadaway Creek where it flows through Fredonia. In 1821, William A. Hart, drilled a hole into the shale there to a depth of 70 feet and obtained a flow of gas. He piped and sold the gas to various places of business in the Village and it was also used to light street lamps.

According to Herrick (1949) the Hart well was still producing gas in 1858. From 1821 to 1858 others produced gas for commercial use, with varying success, from gas seeps in the Fredonia area. In 1858, searching for an additional supply of gas, Preston Barmore and Elias Forbes made a location for a new gas well. This was near a gas seep in Canadaway Creek about one mile north of the Hart well. The new well consisted of a dug cavity about 30 feet deep in the bottom of which two holes were drilled to depths of 100 and 150 feet. The holes found gas and the well was hooked up to supply gas to the Village.

In August of 1859 Edwin L. Drake made his great oil discovery near Titusville, Pennsylvania. That 69½ foot hole showed that oil was reservoired in rocks and could be recovered from rock in greater quantity than from a seep by drilling a well into it. After the Drake discovery men began to drill all over western Pennsylvania and southwestern New York looking for oil.

OIL WELL DRILLING

As already described, the earliest wells drilled in New York were attempts to find natural gas. According to Herrick (1949) the first actual oil test in New York was drilled on the Moore farm in northwestern Allegany County in 1860. The hole was drilled to a depth of 600 feet and found some oil but not enough to pay. In 1865 the hole was deepened to 900 feet but no further oil sands were found and the well was abandoned.

The first commercial oil production in New York came from a well drilled by Job Moses in 1865 in the Town of Carrollton in Cattaraugus County. The well was drilled on the Hall farm near the Village of Limestone and is in a northern projection of the Bradford oil field which lies mainly to the south in Pennsylvania. When completed, the well produced about seven barrels of oil a day from the Bradford Third sandstone.

In 1878 and 1879, O. P. Taylor found the first oil production in the Allegany field in Allegany County. This became the largest oil field in New York. Subsequent drilling, including the Richburg extension in 1881, enlarged the limits of the field and discovered a number of separate small pools. Some of these extend eastward across the Allegany County line for a short distance into southern Steuben County.

Chautauqua County is the only other New York County which has had significant oil production. The first drilling for oil took place in the southeastern part of the County in 1919 and continued for several years with negative results. The shallow Busti oil field was discovered in 1945. The following account is taken from an article by Van Tyne in the Guidebook for the 1974 NYSGA meeting at Fredonia: "In 1945 Todd M. Pettigrew, backed by the Thomas brothers of Chicago, began a drilling program to develop oil production in the Busti area located about four miles southwest of Jamestown. Several wells were drilled over a three year period and modest production was established from the Glade sand at about 700 feet in depth. However, the sand is tight and initial production declined rapidly making the operation uneconomic. By late 1948 the properties were abandoned or farmed-out to other interests who later abandoned them. An interesting aspect of the Pettigrew operation was the use of a "secret" electric device to aid in locating oil bearing rocks". After the invention of hydraulic fracturing of wells to stimulate production the development of the Busti area was revived and over 300 wells have been completed there.

GAS WELL DRILLING

ALLEGANY COUNTY.

In Allegany County the first deep gas wildcat was drilled on the Buell farm southwest of the Village of Hume in the northern part of the

County in 1899. The well was taken to a total depth of 3,326 feet but only a few shallow shows were found and the hole was abandoned.

"The first deep gas in Allegany County was found by a well on the Gilbert farm in the Town of Wirt. The location is slightly to the north and east of the Richburg oil field area. The well was completed in 1928 and produced gas from fractured Tully limestone at a depth of about 4,000 feet. It is now being used for the storage of natural gas." (from Van Tyne & Foster, 1980)

The deepest well in the County was completed in 1963 by Consolidated Gas on the Wolfer farm in the southwest part of the Town of Hume. The well ended in the Potsdam at a total depth of 7,560 feet and was unsuccessful. About one and one-half miles to the southwest in the same Town, Parsons Brothers had drilled a well in 1959 to a depth of 7,337 feet. That well was also unsuccessful.

Not counting the abovementioned two deep tests, only 17 Medina sandstone tests have been drilled in Allegany County. Four of these were completed as gas wells - three are small producers and one is shut-in.

CATTARAUGUS COUNTY.

The first deep gas test in the County was drilled on the Vinton tract in the Village of Gowanda in northwestern Cattaraugus County in 1883. The well was drilled to a total depth of 1,700 feet, through the Onondaga, but found only a show of gas in the upper shales. It was later deepened to the Medina but was still unsuccessful.

"The first deep gas here was discovered in the Medina sandstone by a well drilled in 1898 at Skinner Hollow in the Town of Otto. Deep drilling in later years resulted in the discovery of the Perrysburg and Nashville (Medina) Fields in 1924 and 1926 in the Town of Perrysburg and adjoining Town of Hanover in Chautauqua County.

The first Oriskany sandstone gas production in Cattaraugus County was from the Allegany State Park Field, discovered in 1955. The discovery well, one of nine eventually drilled on the Lockwood farm, came in for 8.4 million cubic feet of gas per day but could not be controlled. The well blew wild for several days until increasing salt water killed the gas flow. Later wells drilled further to the west re-established gas production from this Oriskany pinch-out type trap. The field is the largest Oriskany pinch-out field in New York. In 1957 the unusual Ischua Oriskany gas field was discovered. This unique one-well field produced from a small outlier of Oriskany sandstone separated from and north of, the main pinch-out line." (From Van Tyne & Foster, 1980)

The deepest well in the County, and the second deepest well in New York, was drilled in 1972 by Pennzoil-Amoco in the southeastern part of the County near the New York-Pennsylvania state line. The Enterprise-

Transit-State, or ET-1, well was completed at a depth of 11,680 feet in Pre-Cambrian rocks. No commercial gas was found in the deep rocks but the well was plugged back and produced a small amount of gas from the Oriskany sandstone after treatment.

In 1957 and 1958, Iroquois Gas drilled two deep tests near Perrysburg in the northwestern part of the County. No deep production was found and one well was plugged-back for use as a gas storage well and the other well was plugged and abandoned. In 1971 the same company drilled two more Theresa tests in the same general area but both were dry and abandoned. In late 1989 Fault Line Oil was drilling a 7,500 foot Theresa test in the Town of Ashford.

CHAUTAUQUA COUNTY.

The following material is taken from the account by Van Tyne in the Guidebook for the 1974 NYSGA meeting at Fredonia: "The first deep well was drilled by Alvah Colburn in 1871 at his mill located south of Main Street in Fredonia. The well, which was drilled to a depth of 1,256 feet, encountered gas between 130 to 300 feet, and reached the Onondaga limestone at 1,079 feet.

During the autumn of 1886, and the first half of 1887, the first Medina sandstone well in the county was drilled about one-half mile southeast of the village by the Fredonia Gas and Fuel Company. Additional Medina drilling did not take place until 1903 when the South Shore Gas Company and the Brocton Gas Company began to drill Medina wells in the Brocton and Silver Springs areas. Since 1903, drilling in Chautauqua County has continued on a sporadic basis with the majority of wells being drilled for Medina sandstone gas production.

The first of ten deep wells drilled in Chautauqua County (Note: to 1974; one additional deep hole was drilled in 1976) was the No. 1 Cassity of Frost Gas located one-half mile east of Dunkirk. That well was originally drilled to a depth of 2,052 feet in 1908 and was re-entered and drilled deeper to a depth of 4,035 feet in 1916. The Niehaus well, located two and one-half miles northeast of Dunkirk, was also an old well drilled deeper to 4,517 feet in 1949.

Since then, eight (Note: now nine) more deep wells have been completed, all ending in Cambrian sandstones except one which stopped in the Trenton. No well has been drilled to undoubted basement rocks in Chautauqua County, but the deepest well in the county, the No. 1 Harrington of Wolf's Head Oil located six miles northwest of Jamestown was completed at 7,694 feet in the Cambrian Potsdam sandstone which overlies basement rocks. No commercial gas or oil production has so far been found in the Cambro-Ordovician rocks of Chautauqua County."

ERIE COUNTY.

According to Bishop (1895) the first gas test in Erie County was drilled at Getzville in 1858 or 1859. Eventually, at least 20 to 30 wells were drilled in the area around Getzville. The Getzville gas production is quite interesting because it occurs at the shallow depth of 450 to 500 feet and is found in the Irondequoit limestone. This is the only gas occurrence in New York from this zone although it has also been productive to the west across the Niagara river in Ontario, Canada.

Only a few scattered wells were drilled for gas in Erie County from 1859 until the early 1880's. In 1883 the Buffalo Cement Company began the first systematic search for gas within the City of Buffalo (Bishop, 1895). Within a few years eleven wells had been drilled on their property which was located about one-half mile southwest along Main Street from the University of Buffalo.

In the latter 1880's and in the 1890's many more wells were put down in Buffalo and other parts of Erie County with varying success. By the turn of the century the natural gas industry was well established and drilling in Erie County has continued to the present day. To date, 22 deep wells testing for possible gas in Cambro-Ordovician rocks have been drilled in Erie County. Only one of these wells was moderately successful.

WYOMING COUNTY.

The first well in this County was drilled for oil by the Vacuum Oil Company in the Oatka Creek valley in the Town of Middlebury in 1877-1878. The well, known as the "Pioneer" well, found a salt bed and ended at a depth of 1,455 feet, in the Salina. In the latter 1800's many other salt wells were drilled in the Oatka Creek valley and in other parts of Wyoming County.

In 1883 a well was drilled to a depth of 1,960 feet at Attica. Although the well penetrated the Medina it was dry. In 1897 a well drilled 400 feet west of this well to a depth of 1,708 feet became the first gas producer in Wyoming County and the discovery well of the Attica (Medina) gas field.

Fifteen wells have been drilled to the Trenton, or deeper, in Wyoming County. The first of these was drilled at the K.R.Wilson plant in Arcade in 1946. The well was drilled to a total depth of 7,144 feet but was dry. In 1961 a well was drilled near the Village of Gainesville by Consolidated Gas. The well reached a depth of 7,182 feet and finished in the Potsdam as a dry hole.

In 1963, Trans-American Petroleum Corp. re-opened a 1961 dry hole located on the Strathearn farm in the Town of Middlebury. They drilled the well deeper from its original depth of 2,481 feet to a new total depth of 5,507 feet. This deepened well discovered gas in the Theresa formation, one of only a few such producers in western New York. Within a year, four other deep wells had been drilled in the vicinity, one of these to the Pre-Cambrian. Only one other well found gas production in the Theresa. None of the other eight deep tests in the County was successful.

REGIONAL GEOLOGY OF PRODUCING FORMATIONS

(Note: Much of what follows has been taken from the publication: "Inventory and Analysis of the Oil and Gas Resources of Allegany & Cattaraugus Counties, New York" by Van Tyne and Foster, 1980)

UPPER DEVONIAN SANDSTONES.

Oil and gas producing sandstones of Late Devonian age are found in Allegany, Cattaraugus, Chautauqua and Steuben Counties (Fig. 1). These are interfingering and lensing rocks that were deposited in shallow water at the distributary front of a large delta complex which was built out from the east during Middle and Late Devonian time. The sandstones have generally low porosity and permeability that does not appear to be related to regional trends or structures.

Lithologically, the sandstones are very fine to fine-grained graywackes consisting of angular quartz grains, rock fragments, and subordinate mica. Little or no feldspar is present. The inter-granular cementing material is largely silica.

Where productive they have an average porosity of 12 to 14 percent with an average permeability of 3 to 10 millidarcies (md.). Other values have been measured in core tests of the sandstones but these values appear to apply to most of the productive zones. Porosities in productive areas are generally in the above range, but production has been obtained in areas with permeabilities of 1 md. or less.

DEVONIAN BLACK SHALES.

Several black shale sequences occur within the Upper and Middle Devonian rocks of New York. The uppermost of these shales is the Dunkirk which lies at the base of the Perrysburg Formation. The Dunkirk outcrops and has been a gas producer in the Lake Erie area of Chautauqua County. East of Cattaraugus County, it thins and becomes grayer.

COMPOSITE PALEOZOIC STRATIGRAPHIC SECTION

FOR SOUTHWESTERN NEW YORK

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

† INCLUDES GLADE, BRADFORD 1st, CHIPMUNK BRADFORD 2nd, HARRISBURG RUN, SCIO, PENNY, & RICHBURG

INCLUDES BRADFORD 3rd, HUMPHREY, CLARKSVILLE, WAUGH & PORTER, & FULMER VALLEY

Figure 1.

The Rhinestreet may have some possible gas potential. In the Lake Erie shoreline area the Rhinestreet is a massive black shale with interbedded gray shale and argillaceous limestones in the upper section. To the east it contains thin silty interbeds which can be gas reservoirs.

The Middlesex is a highly organic brownish-black shale. It commonly consists of laminae of dark organic matter alternating with clays and silt-size quartz grains. When broken, it exudes an oily odor. There is no known gas production from this zone.

The Geneseo black shale overlies the Tully limestone, where present, and is cut out by erosion in far western New York. There is no gas production from this zone in New York.

The Marcellus black shale is the basal formation of the Hamilton Group which overlies the Onondaga limestone. This is a highly organic black shale and the most highly radioactive shale in the New York Devonian section. It may be recognized by a distinctive, strong, rightward deflection on a gamma ray log.

ONONDAGA LIMESTONE.

The Onondaga limestone varies from about 10 feet in Steuben County to more than 200 feet in Chautauqua County. This fossiliferous, medium gray limestone may contain oil and gas in fault-generated fractures or in primary porosity in fossil coral reefs. The faulted Onondaga limestone will be discussed in the "Bass Islands" section under <u>Recent Developments</u>.

In 1967, the first Onondaga pinnacle reef was discovered in a wildcat well drilled on the Cornell farm in the Town of Jasper, Steuben County, by the Wyckoff Development Company. An open flow of about one million cubic feet of gas per day (mmcfgpd) was recorded from the Onondaga reef. Gas was also found in the Oriskany sandstone and the final open flow from this well was about 7 million cubic feet of gas. The discovery was later designated as the Wyckoff Field. The Homer Banks well, drilled by the Sylvania Corporation, penetrated 196 feet of Onondaga reef but was completed and produced initially from the Oriskany sandstone (Fig. 2). The Oriskany produced over 912 mmcf of gas from 1968 to 1971. The well was re-completed in the Onondaga in 1972 and has produced over 2.8 billion cubic feet (bcf) of gas to date.

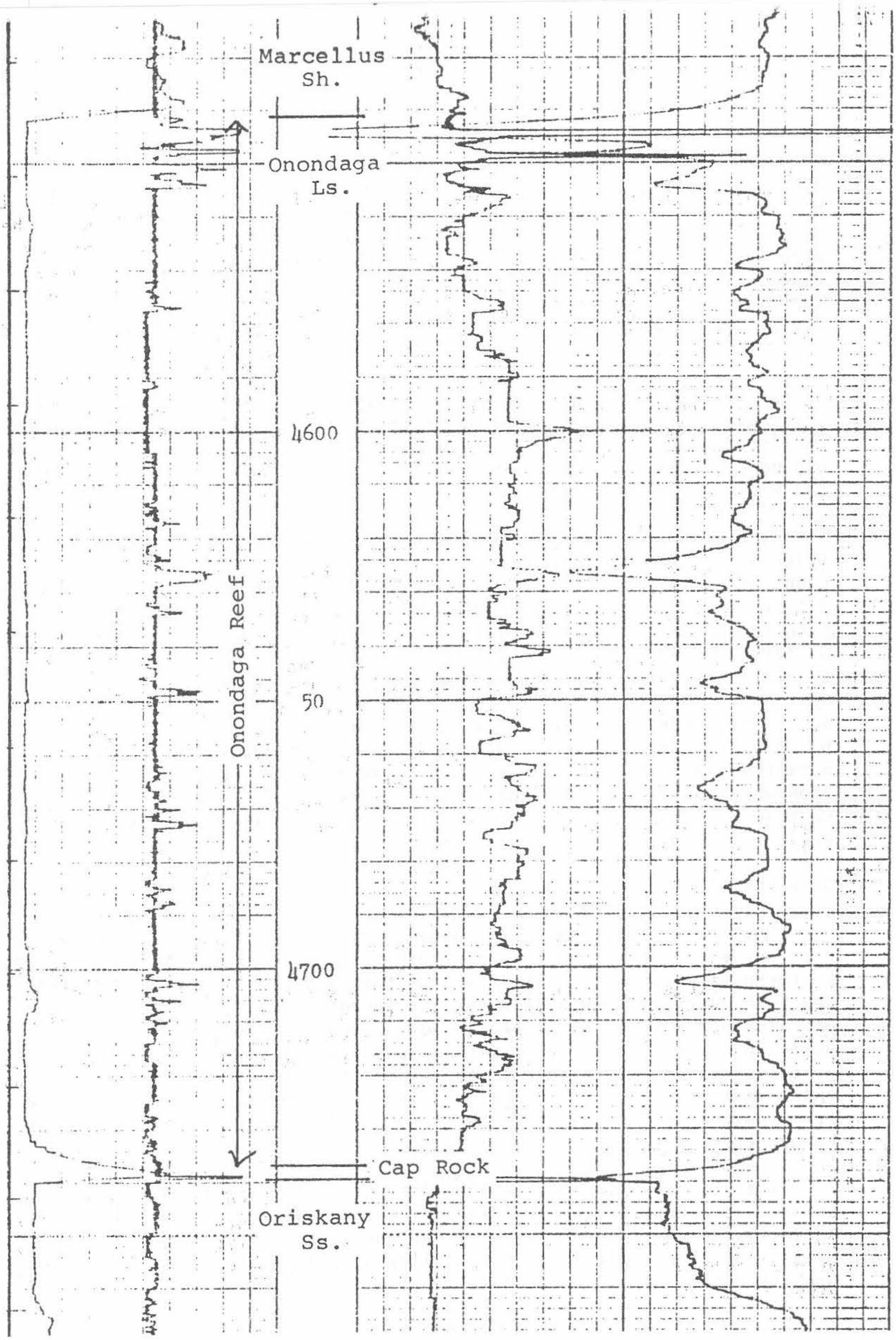


Figure 2. Geophysical log of an Onondaga pinnacle reef in the Homer Banks well, Wyckoff Field, drilled by the Sylvania Corporation in July 1967. After acidizing, the well had an open flow of 3,843 mcfgpd.

From 1968 to 1981, six more pinnacle reefs were discovered, five in New York and one in Pennsylvania. The pinnacle reefs project above the base of the Onondaga limestone to about 200 feet in thickness and may be more than 6,000 feet across. These Devonian age coral reefs created reservoirs that have had production from .5 bcf of gas in the Flatstone reef in Cattaraugus County to over 6.6 bcf from the Raish #1 well in the Adrian Reef field (Fig. 3) in the Town of Canisteo, Steuben County.

The Wyckoff Field has had production from both the Oriskany sandstone and the Onondaga reef. The Oriskany gas was produced from three wells from 1968 to 1974 and totaled over 4.2 bcf. The Onondaga reef production began in 1972 from three wells and over 4.8 bcf has been produced to date. A total of over 9 bcf of gas has been produced from the Wyckoff Field thus far.

The Onondaga limestone also contains numerous bioherms or reefs (Fig. 4) which are completely contained within the Onondaga. The bioherms may be from 80 to over 100 feet in thickness. Their lateral dimensions are uncertain as offset wells, 2,000 feet away, generally miss the bioherm. Several of these bioherms have been discovered in Cattaraugus, Chautauqua and Erie Counties. Many have been cased off because the primary target was the deeper Medina Formation (Fig. 5). Some have produced significant quantities of oil and gas, although they do not have the reserves of the larger pinnacle reefs. One reef well has produced several thousand barrels of oil and it is reported that more than 40 mmcf gas and over 1000 barrels of oil have been produced from another. These bioherms, which can be encountered at shallow depths, can augment production and add to the economic significance of a well if recognized.

ORISKANY SANDSTONE

The Oriskany is a grayish-white quartzitic sandstone varying from 0 to 35 feet thick in western New York (Fig. 6). It is present only in southern Allegany and southeastern Cattaraugus Counties in far western New York. Both gas and salt water are found in the Oriskany. The gas migrates up-dip and is trapped in any closed structurally higher position. These gas traps may be up-dip against faults, in closed domes or folds, or up-dip regionally to the north where the Oriskany sandstone thins and pinches out. The major Oriskany gas production, however, is from faulted anticlines where the associated fracturing develops a much higher effective porosity.

The Oriskany pinchout is the limit of the Oriskany sandstone in the subsurface. Several gas fields have been found along the pinchout line. These fields range in size from about one-quarter billion cubic feet to over 20 billion cubic feet in the Allegany State Park field and normally produce condensate with the gas.

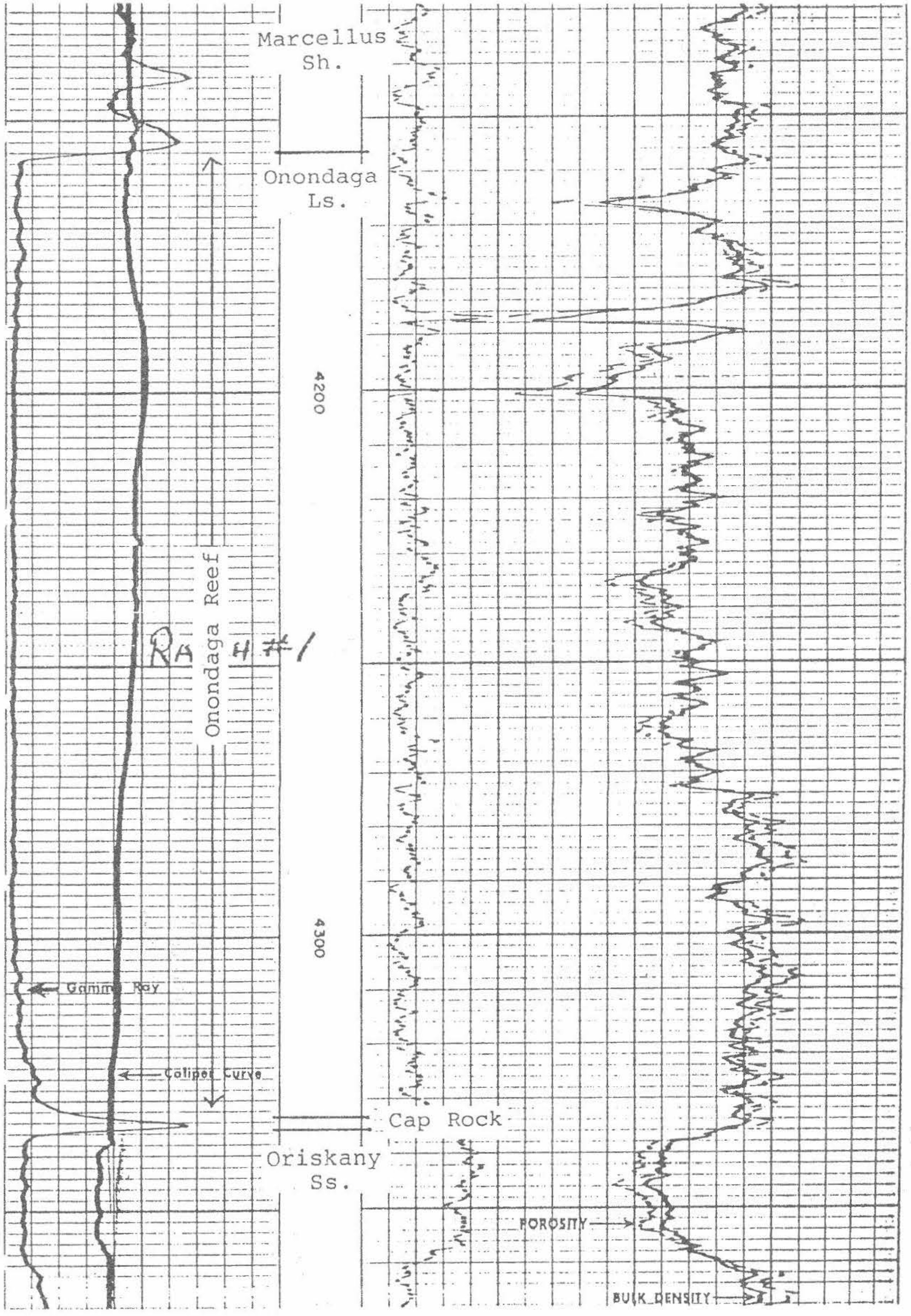


Figure 3. Geophysical log of an Onondaga pinnacle reef in the W.J. & L.I. Raish well, Adrian Reef Field, drilled by the Cabot Corporation in July 1971. The well had an estimated open flow of 11,200 mcfgpd.

Sat. B11

Figure 4. Geophysical log of an Onondaga bioherm in the L. Philips #1, drilled by Empire Exploration, Inc., in the Town of Villenova, Chautauqua County, N.Y.

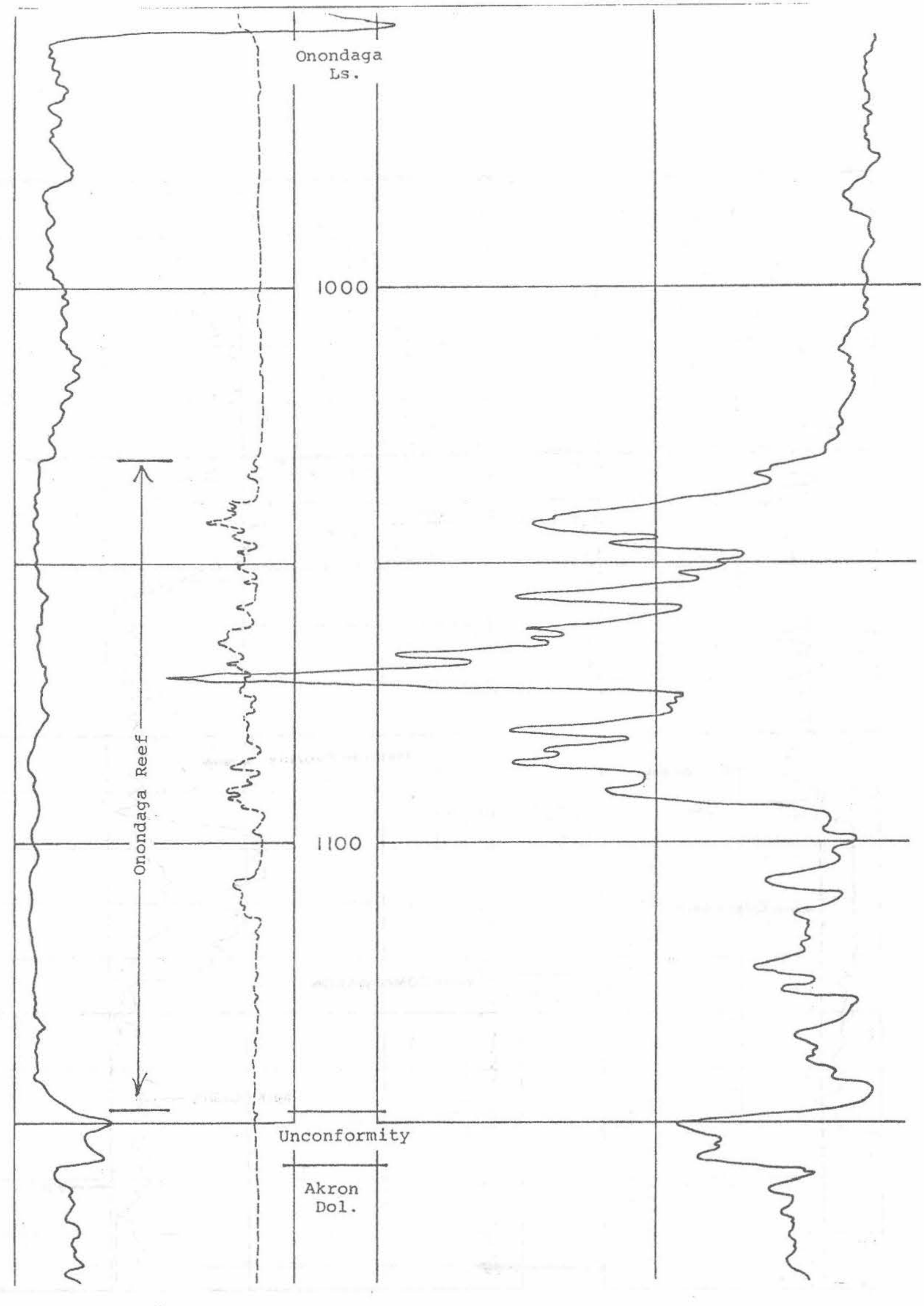


Figure 5. Geophysical log of an Onondaga bioherm in the E. Majka #1 well, drilled by Local Energy in the Town of Pomfret, Chautauqua County, N.Y.

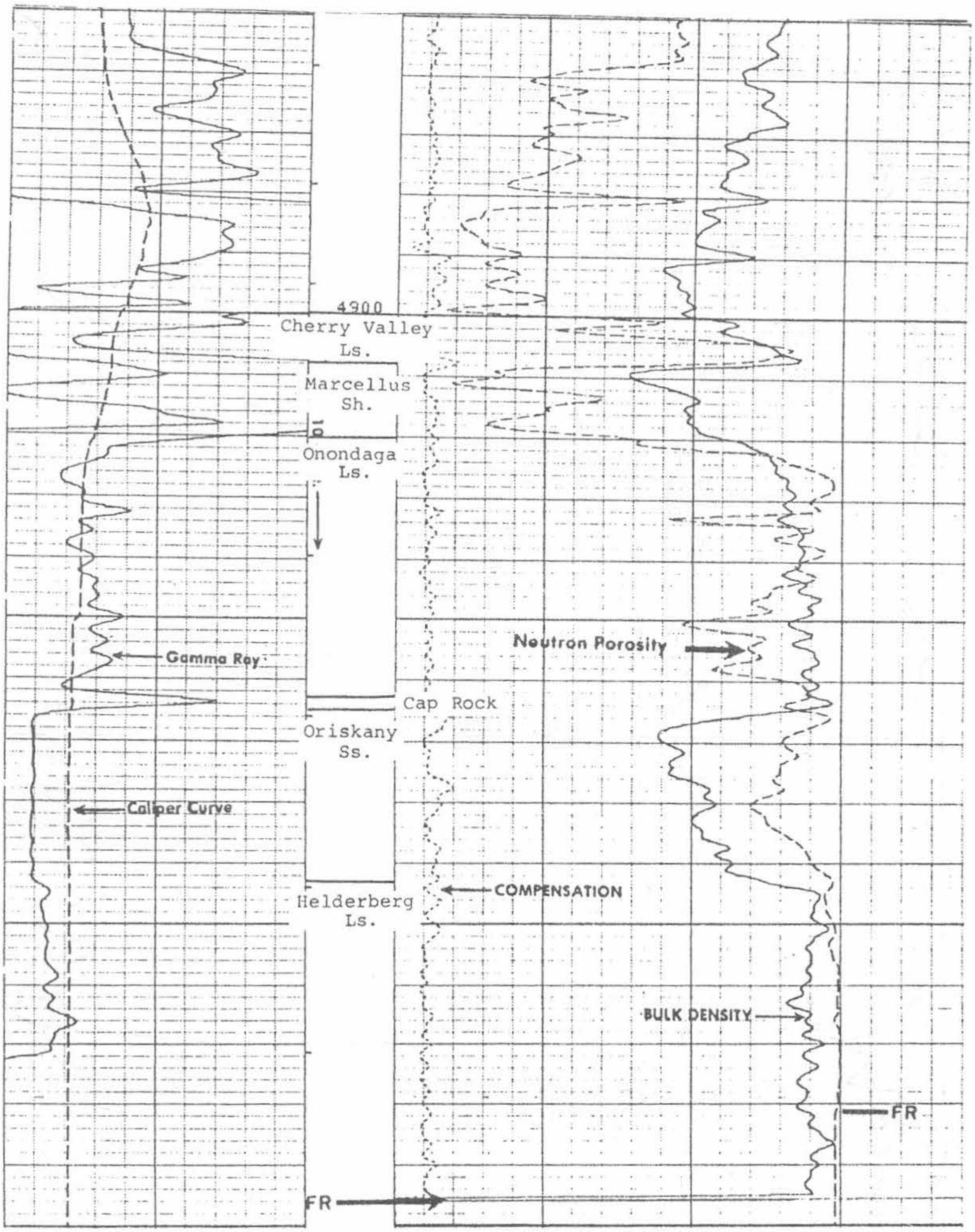


Figure 6. Geophysical log of a representative Oriskany well drilled by Penn York Energy Corporation in the Town of Willing, Allegany County, N.Y.

MEDINA GROUP.

In western New York the Medina Group is divided into the Grimsby sandstone (upper Red Medina), the Power Glen shale and the Whirlpool sandstone (lower White Medina) (Fig. 7). Medina Group rocks were deposited in near-shore, shallow marine environments following the deposition of the Queenston delta complex.

The Grimsby consists of fine to medium-grained red and mottled red and white sandstones with interbedded red and light green shales. Cross bedding, ripple marks, channel deposits, and mud cracks are common sedimentary features. In the Niagara Gorge, the Grimsby is up to 52 feet thick, increasing in thickness to the east in Genesee County to over 100 feet.

The Power Glen shale consists of finely laminated, gray to greenish gray shales with interbeds of lenses of grayish-white sandstones. This unit is referred to by drillers as the "shale break" between the overlying Red Medina sandstone and underlying White Medina sandstone. The Power Glen is about 36 feet thick in the Niagara Gorge, pinching out to the east.

The Whirlpool is a light gray to white, fine to coarse-grained quartz sandstone up to 25 feet thick in the Niagara Gorge. The Whirlpool sandstone has porosity zones in the subsurface in Erie, Chautauqua and the western portion of Cattaraugus Counties that are excellent producers of natural gas. The Whirlpool pinches out to the east with only the Grimsby remaining to represent the Medina Group.

The Medina has been the focus of the vast majority of the drilling programs in western New York. The high success rate of Medina wells has been one of the leading factors in this drilling. Grimsby wells may average 80-100 million cubic feet (mmcf) of gas in fifteen years of production. Whirlpool wells, in areas of enhanced porosity, may have much better production of up to 250 to 300 mmcf or more over fifteen years. Currently, several companies are concentrating their exploration activities in searching for the more lucrative Whirlpool wells.

ORDOVICIAN AND CAMBRIAN ROCKS

The red Queenston is the youngest Ordovician unit in New York. It forms a clastic wedge, probably of deltaic origin, thickening westward from a feather edge in eastern New York The lower Queenston has a gradational and erratic color boundary with the underlying green Oswego sandstone.

Because the source for Queenston sedimentation was to the east and south-east of its present location, the unit has a coarser clastic content in the east and consists of finer clastics to the west. In Erie County the Queenston is a medium red and grayish-red shale and mudstone with silty and very fine sandstone interbeds. Scattered light green

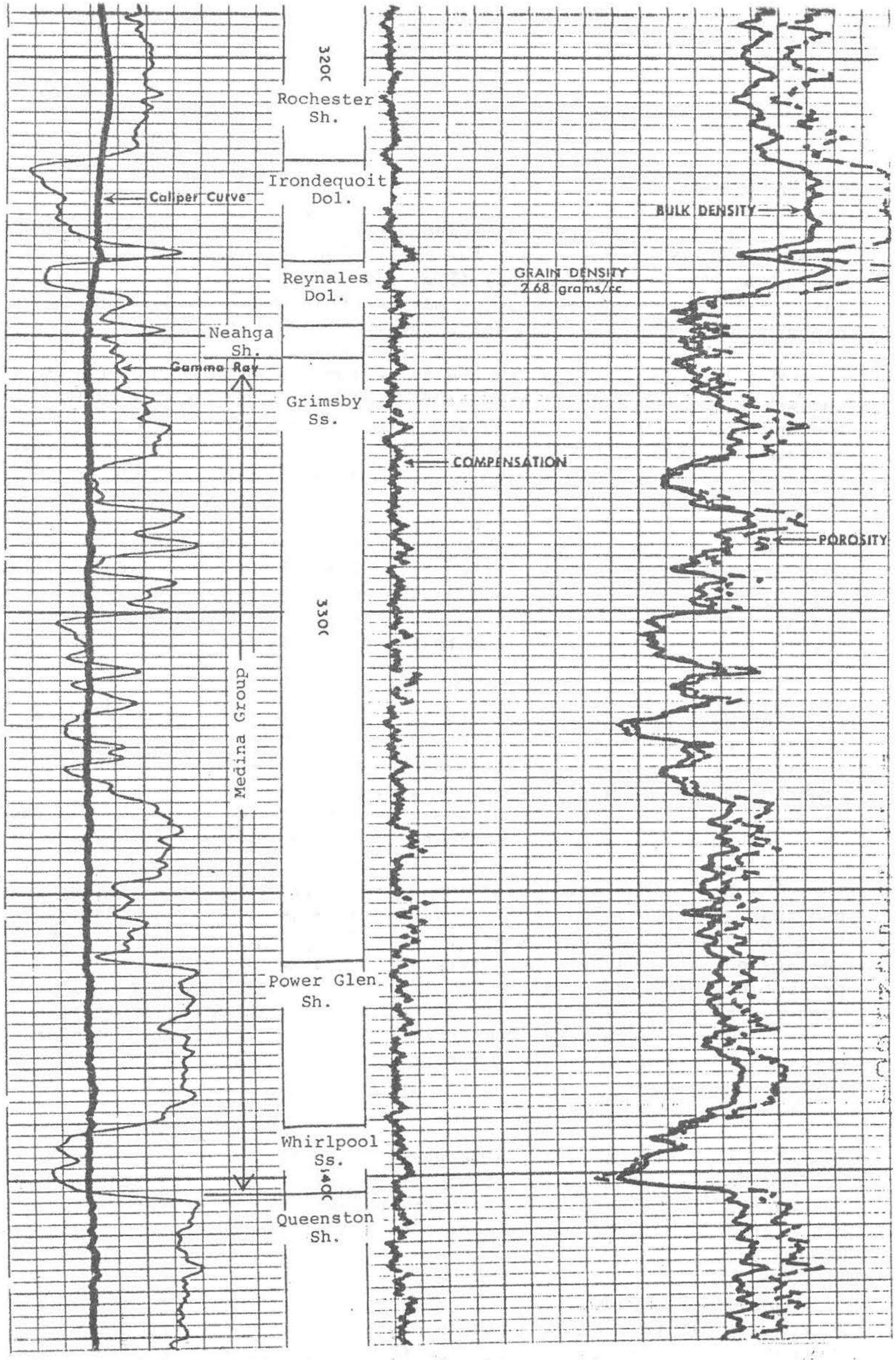


Figure 7. Geophysical log representative of a Medina well. Bemis Unit, drilled by Paragon in the Town of Westfield, Chautauqua County, N.Y.

shale is also present. The Queenston has produced gas from some of these interbedded sandy zones, mainly in its upper portion, in western New York.

The Trenton-Black River section is a brown and gray carbonate sequence varying in thickness from about 700 feet in Chautauqua County to about 1,100 feet eastward to Allegany County. The Trenton has produced gas from shallow wells south and west of the Adirondacks and occasional shows of gas have been reported in the Trenton from wells in western New York. Traps in lower Black River rocks could be formed by draping of porous beds over erosional remnants left on the underlying Lower Ordovician disconformity surface.

Sandstones in the Galway (Theresa) Formation often have shows of gas and salt water. However, a structural closure or fracture system would have to be present to form a gas trap. This is also true of the underlying Potsdam sandstone.

RECENT DEVELOPMENTS

"BASS ISLANDS" TREND

In February of 1981 Envirogas, Inc., while drilling the No. 8 Wassink well just west of the Village of Clymer in Chautauqua County, encountered a large flow of oil from a section of rock below the Onondaga limestone. The source of this oil proved to be fractured Akron dolomite. Later in 1981 various wells being drilled for Medina gas in northeastern Chautauqua County also unexpectedly encountered high flows of oil and gas from this and adjacent fractured zones. This resulted in some blowouts and fires which signaled that a new producing trend had been discovered.

Operators involved in the early phases of this discovery applied the name, "Bass Islands" to the productive zone because they were more familiar with the Ohio name for this Upper Silurian section than with the name Akron which is the equivalent rock unit in New York.

The wells in this trend occur in a structurally complex area found by Van Tyne in 1978. This occurred while he was doing subsurface structural mapping for the Eastern Gas Shales Project (Devonian black shales) of the United States Department of Energy. The work was done while he was head of the New York State Oil and Gas Research Office and acting as a contractor for the DOE.

By the use of Gamma Ray logs from an area of extensive Medina drilling in the Town of North Harmony, south of Chautauqua Lake, he was able to map a low-relief, highly thrust-faulted anticlinal feature. The structural similarity to other such features in the well-known Appalachian thrust belt farther southeast in New York and Pennsylvania was obvious. Further work revealed faulting in other wells south of

Chautauqua Lake. The presence of faults in two wells northeast of the lake and some surface evidence of faulting suggested that the feature could form a long, narrow northeast-southwest trending structure. This would extend completely across Chautauqua County, the northwest corner of Cattaraugus County and into southern Erie County. Subsequent Gamma Ray log correlation work in those areas bore out that assumption. The feature as mapped and appearing in Black Shale publications in 1980 is about 65 miles long and mostly about one and one-half miles wide (Fig. 8).

The structure consists of a series of linear, en-echelon thrust faults which form an anticline consisting of a complex of horst and graben blocks. The feature is a decollement structure with faults emanating from a glide plane in the Vernon "B". Little or no structure occurs from that zone down to the Queenston where the Medina sandstone wells are stopped.

Production occurs from highly fractured zones along the fault trends at depths from about 2,600 to 2,900 feet. Oil and gas has been found in the Marcellus shale, Onondaga limestone, and underlying Bois Blanc, Akron dolomite and Bertie dolomite. The best production has been in the Akron and Onondaga sections. The State Division of Mineral Resources has so far delineated 43 separate pools in this trend.

The oil produced is a high gravity, paraffin base crude. Gas cap production has been marked by the high flows which could be anticipated from such a highly permeable, fractured reservoir. Initial gas flows as high as 60,000 MCFGPD and oil flows up to 2,400 BOPD have been reported. At present, no new drilling is taking place in the trend and production is showing an extreme decline. Total gas production from the trend, through 1989, is estimated at 8½ billion cubic feet and oil production about 1.6 million barrels.

DEEP GAS DRILLING.

The search for deeper gas has focused on basement faulting and rift zones. Such faults often extend upwards into the Trenton and evidently provide pathways for fluid migration leading to dolomitization. Dolomitized limestones can form porosity traps and the faulting can provide extensive fracture systems for migration of dolomitizing fluids and for fracture porosity traps.

Several tests searching for such deep traps have been drilled along the north-south Clarendon-Linden fault system in Monroe, Orleans, Genesee and Wyoming Counties without success so far. In other areas, Columbia Gas has drilled three tests, and others one well, on deep rift features in northeastern Steuben and north-central Schuyler County. The second well in Steuben County was drilled to a depth of 7,961 feet by Columbia Gas in late 1985. Some deep gas was found and the well has been extensively treated and tested in the deep zones.

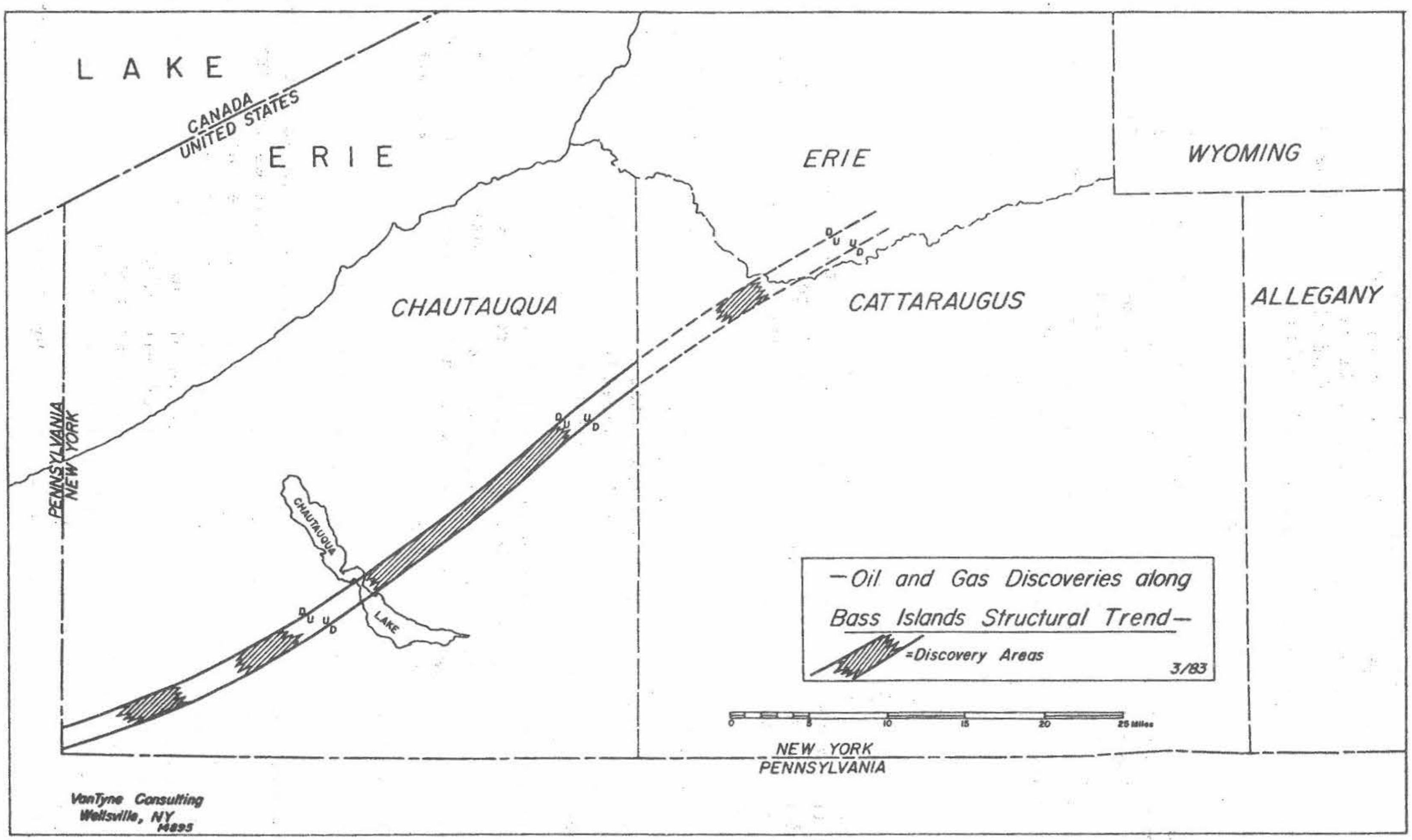


Figure 8. Map Showing Extent of the "Bass Islands" Structural Trend in New York. Only Faults Shown are Bounding Reverse Faults.

In Schuyler County Columbia Gas drilled two test wells in the Town of Reading to depths of 8,397 and 8,510 feet in 1986. Both found dolomitization of the deep limestones as well as gas shows. They were both treated and tested. However, only one of the wells has been completed and shut-in as a possible gas producer.

STORAGE FIELDS.

Storage fields are developed in formations where stratigraphic and/or structural traps have contained a significant amount of gas. These fields have been previously produced and must offer a large amount of storage capacity to justify development. The porosity and permeability of the storage formation is important for injection and withdrawal rates. Once a formation has been delineated and designated as a potential storage field, new wells are drilled for injection and withdrawal of gas and a compressor station is constructed to inject the gas.

There are 21 natural gas storage fields in New York. These have been developed in the Onondaga limestone-Akron dolostone, Oriskany sandstone and the Grimsby and Whirlpool sandstones of the Medina. Gas from local production and the southwestern United States is pumped into the storage fields during the summer months for withdrawal during the winter as required.

National Fuel Gas Corporation developed the first underground storage field in the United States in 1916. This field was originally discovered in the late 1800's and was produced until development as a storage field. National Fuel, or its subsidiary Penn York Energy Corporation, operates 15 of the 21 storage fields in New York State.

Although most of the gas storage in New York is for local consumption, some storage service, such as that offered by Penn York Energy Corporation, is provided for customers on the East Coast. The growing market for natural gas in the eastern United States will stimulate a demand for the development of additional storage fields in New York and neighboring states.

DRILLING AND COMPLETION METHODS.

The speed and efficiency of rotary drilling has replaced the cable tool rigs of the past. Today the drilling process involves the drilling of a surface hole and installation of a ten-inch surface casing which is cemented to surface. The surface casing would be set through the over-burden into bedrock for a total length of 250 to 500 feet through fresh water zones. In some valleys filled with glacially derived sediments more than 1,000 feet of surface casing may be required. A 7-7/8" hole will then be drilled to total depth (T.D.) and a 4½" production string run through the producing zones and cemented in place. The sand is then perforated through the casing opposite the productive zones, as determined by the analysis of geophysical logs, fracture treated, cleaned-up, and shut-in for pipeline hookup and production.

Completion techniques during the past few years have undergone several changes and improvements based on increased technological advancements available to the industry. The efforts put into the completion program are just as important as finding the productive zone and may determine whether a well produces up to its expectations.

ENVIRONMENTAL CONCERNS.

During the last twenty years environmental problems have been increasingly brought to the public's attention. The oil and gas industry is concerned about the environmental affects it may have relating to the exploration and production of oil and natural gas.

Well drilling activities, pipeline installations and the installation of surface and subsurface storage areas for oil and natural gas directly involve environmental considerations. The industry has taken measures to ensure that our vital natural resources are protected. Proper surface casing installation and cementing procedures are implemented to ensure protection of the groundwater from salt water, oil or natural gas. Drill cuttings and fluids and formation fluids are contained on site for proper disposal. Berm dikes are constructed around oil storage tanks to contain the contents should a tank rupture.

The New York State Department of Environmental Conservation, Division of Mineral Resources, issues drilling permits and regulates drilling and completion activities. Their responsibilities also include authorizing the plugging of new and old wells. The oil and gas industry works closely with the Mineral Resources Division in establishing proper procedures to protect the environment and with landowners to minimize disruptions to their land.

Environmental regulations have added additional costs to drilling programs and at times have been controversial. These added costs are a major concern that must be considered when developing an exploration or development program. Environmental considerations will continue to be important for the oil and gas industry in all future activities.

FUTURE DRILLING

Drilling in New York is presently about one-fourth that of the peak drilling years in the early to mid 1980's. About 150 wells were drilled in 1989. Oil well drilling is virtually non-existent so almost all drilling now is for gas. The great majority of gas wells are drilled for Medina gas in far western New York and Queenston gas production in central New York. Several Oriskany sandstone wells and several exploratory tests of various kinds are also drilled each year.

Drilling for Oriskany gas is confined to searching for new traps on known Appalachian fold and fault structures. High resolution seismic work and a better understanding of the tectonic geology of these features has helped in this search.

There has been little or no drilling for possible Onondaga reefs in recent years. Extensive seismic work has been done searching for more reefs but indications of reefing have so far not been found.

As already discussed, there is an increased interest in drilling prospects in deep Ordovician and Cambrian rocks. This could increase if there is an improvement in the wellhead price for gas. These rocks have rarely been targeted in most areas of western New York.

The outlook for an upward movement in wellhead gas prices is good but probably only on the basis of a small annual increase. The average wellhead price in New York is probably about $2.25 per MCF at present.

The price for oil is currently $24.50 per barrel (8/20/90) up from a low of $16.50 per barrel in late June. This rapid increase is due to the Iraq-Kuwait crisis. However, there is little or no oil left to be recovered from the old New York oil fields no matter what the price. For new tertiary recovery projects to be initiated the oil price would probably have to be at least $30.00 per barrel and would have to stay at that level, or higher, for several years.

ACKNOWLEDGEMENTS

Thanks are extended to the Division of Mineral Resources, New York State Department of Environmental Conservation, and especially to Tom Wickerham of the Region 8 office in Avon, for Onondaga reef production information. Roger Willis of Universal Well Services in Meadville, Pa. was helpful in supplying well completion information. Don Thompson of Empire Exploration in Buffalo provided information about gas storage fields. Thanks are due Twila Day of Quaker State Oil Refining Corp. in Titusville, Pa. for current information about Penn Grade oil prices.

REFERENCES

BISHOP, I.P., 1897, Petroleum and Natural Gas in Western New York: in New York State Museum Report 51-2, p. 9-63.

__________, 1895, The Structural and Economic Geology of Erie County: in 15th Annual Report of the State Geologist, p. 305-392, 6 Fig.

HERRICK, J.P., 1949, Empire Oil, New York, Dodd, Mead & Company, 474 p.

LUTHER, D.D., 1896, The Brine Springs and Salt Wells of the State of New York, and the Geology of the Salt District: in 16th Annual Report of the State Geologist, p. 169-226.

VAN TYNE, A.M., 1974, Geology and Occurrence of Oil and Gas in Chautauqua County, New York: in Guidebook, Geology of Western New York State, 46th Annual Meeting, New York State Geological Association, Fredonia, p. H1-H8.

__________, 1983, Oil and Gas Developments in New York in 1982: American Association of Petroleum Geologists Bulletin, v. 67, p. 1554-1557.

__________, 1983, Natural Gas Potential of the Devonian Black Shales of New York: Northeastern Geology, v. 5, Nos. 3 & 4, p. 209-216.

VAN TYNE, A.M. and FOSTER, B.P., 1980, Inventory and Analysis of the Oil and Gas Resources of Allegany & Cattaraugus Counties, New York: Southern Tier West Regional Planning and Development Board, Part I, p. 17-48.

TRIP B. - GEOLOGY AND OIL & GAS EXPLORATION IN WESTERN NEW YORK - Bastedo and Van Tyne.

There is no road log available for this trip because at the time this guidebook was being compiled, we couldn't be sure of what operations it might be possible to visit. We will be visiting the Drake Well Museum at Titusville, Pennsylvania and hope to make two or three other stops along the way. These stops will depend upon what, if any, drilling activities are being conducted in the area on the day of the trip.

SEQUENCE STRATIGRAPHY OF THE TYPE NIAGARAN SERIES (SILURIAN) OF WESTERN NEW YORK AND ONTARIO

By

CARLTON E. BRETT, WILLIAM M. GOODMAN AND STEVEN T. LO DUCA
Department of Geological Sciences
University of Rochester
Rochester, New York 14627

INTRODUCTION

Recent advances in sequence stratigraphy resulting from seismic profiling of continental shelf sedimentary prisms (Vail et al., 1977; Wilgus et al., 1988), combined with a renewed interest in cyclicity and periodicity in the sedimentary record (see, for example, Bayer and Seilacher, 1985; Einsele, 1985; Fischer et al., 1985; House, 1985) have had a revolutionary impact on the field of stratigraphy. Subdivision of sedimentary prisms into unconformity-bound, genetic stratal packages, or sequences, (sensu Vail et al., 1977; Fig. 1) has permitted seismic stratigraphers to make rather detailed intercontinental correlations and has provided a new genetic model of stratigraphic dynamics on continental shelves (see, for example, Van Wagoner et al., 1988; Posamentier et al., 1988a, b; Sarg, 1988; Galloway, 1989). If depositional sequences are produced, at least in part, by eustatic sea level fluctuations, then the principles of sequence stratigraphy should be equally applicable to epicontinental sea and foreland basin strata which appear to be highly influenced by high frequency, low magnitude sea-level oscillations (see, for example, Goodwin and Anderson, 1985; Busch and Rollins, 1984). However, to date, relatively few studies have attempted to discern unconformity-bound sequences in settings other than passive continental margins (notable exceptions include Nummedal and Swift, 1988; Cross, 1988). In order to test the applicability of sequence concepts to epicontinental strata, field geologists must begin to apply the model at an outcrop scale to a broad array of sedimentary environments. Only in this way can the generality of the sequence model be tested and peculiarities of continental interior and foreland basin sequences be distinguished from those characteristics specific to continental shelf deposition.

The Silurian strata of western New York and adjacent Ontario provide an excellent test case for the application of the principles of sequence stratigraphy in a foreland basin setting. The Silurian rocks of the northern Appalachian foreland basin region in New York State, Pennsylvania and adjacent to Ontario are a classic succession in the history of North American geology (Hall, 1839, 1852). Indeed, they include some of the first formally designated stratigraphic units in North America. These rocks, representing marine and paralic, mixed siliciclastic and carbonate facies, have recently been studied in detail both at the surface and in the subsurface of the Niagara region. Extensive drilling for USGS/EPA and industry supported groundwater studies have provided the impetus for a major revision of the classic Niagaran Series (Early and medial Silurian). Extensive regional correlation of outcrop and subsurface data is providing a broader overview of facies patterns in the northern Appalachian basin.

The Niagaran Series records diverse litho- and biofacies, ranging from nonmarine sandstones to deep-water shales and reefal carbonates. These strata span the early Llandoverian through the Ludlovian, an interval of about 10 to 15 m.y. Furthermore, the Niagaran series is internally divided by multiple regional unconformities that provide a basis for tentative subdivision of these strata into genetic depositional sequences (Fig. 4).

In this report we discuss the sequence stratigraphy of the Silurian Medina, Clinton, and Lockport Groups of westernmost New York and Ontario with emphasis on the following: 1) positions of major sequence bounding erosion surfaces; 2) positions of internal discontinuities associated with transgressions and offshore sediment starvation (i.e., marine flooding or downlap surfaces); 3) internal structure of sequences into component systems tracts (facies representing lowstand, transgressive, and highstand portions of large-scale cycles); 4) fine-scale subsequence and parasequence subdivisions and their bounding surfaces; and, finally, 5) lateral changes of facies within sequences and their implications for basin geometry and dynamics.

The Silurian strata of the Niagara region demonstrate the feasibility and utility of a sequence stratigraphic approach. Although these rocks do not perfectly fit all the details of the original sequence model, many constructs and concepts are applicable. Furthermore, detailed studies of foreland basin successions, such as this, should aid in refining the original sequence model and making it more broadly applicable.

GENERAL CONCEPTS OF SEQUENCE STRATIGRAPHY

As underscored repeatedly by the Exxon seismic stratigraphy group (see Wilgus et al., 1988), depositional sequences consist of genetically related packages of relatively conformable strata bounded by subaerial unconformities and their correlative submarine conformities. Sequences are comprised of internal subdivisions of smaller-scale units termed parasequences which commonly form genetic groupings referred to as parasequence sets (see, for example, Vail et al., 1977; Van Wagoner et al., 1988; Figs. 1, 2, herein). We introduce the term sub-sequence to refer to small scale sequence-like divisions that lack evidence for major erosion at their boundaries. The subsequences are composed of two or more parasequence sets and correspond approximately to the concept of systems tracts, proposed by Vail et al. (1977)(Van

Wagoner et al., 1988). Sequences can be subdivided into systems tracts, representing deposition during three phases of a major sea level cycle, lowstand (major sea level fall or regression), transgressive (initial sea level rise or transgressional, and highstand (still stand to early sea level fall). Systems tracts are identified by their position in the sequence and the stacking order of component parasequences (Figs. 1, 2).

Lowstand or continental margin systems tracts accumulate during periods of relative sea level lowstand when much of the continental shelf is subjected to erosion. Consequently, lowstand systems tracts are generally confined to deep sea fans and wedges or erosional valleys on the continental shelf (Fig. 1). Lowstand deposits are bounded at their base by the sequence bounding unconformity, a major erosional surface produced when the rate of sealevel far exceeds local subsidence rates, resulting in rapid sealevel drop.

Transgressive systems tracts are made up of retrograding (upward deepening) parasequences and form during times of relatively rapid sea level rise. Transgressive systems tracts are composed of relatively shallow water sediments, and typically are relatively thin and tend to become increasingly starved of siliclastic sediments upward. They are typically bounded at their bases by a sharp transgressive surface that represents a rapid increase in the rate of sea level rise and separates the transgressive systems tract from the underlying lowstand systems tract (Figs. 1, 2). This surface may be marked by erosional clasts from the underlying lowstand deposits as well as phosphatic and resistant skeletal material in some instances (see Baum and Vail, 1988). The base of the transgressive systems tract is often a very sharply demarcated surface in outcrops as noted by Baum and Vail (1988); it is commonly coextensive with a ravinement surface produced by erosional shore face retreat in a high energy environment. Consequently, the base of the transgressive systems tract typically displays shallow shelf sand sharply overlying non-marine or paralic sediments of the lowstand systems tract. The upper boundary of the transgressive deposits is sharply overlain by deeper water sediments at the base of the highstand systems tract. Transgressive systems tracts are generally relatively condensed near their upper boundaries.

The transgressive systems tract is bounded at its top by another discontinuity surface referred to variously as the downlap surface or surface of maximum sediment starvation (Figs. 1, 2). This surface is commonly associated with (overlain by) a highly condensed horizon containing phosphatic nodules, glauconite, conodonts and other chemically resistant allochems. The condensed section records an extended period of non-deposition or exceedingly low sedimentation rates. A point not usually emphasized, but one brought out by Baum and Vail (1988), is that the base of the highstand systems tract is typically recorded by relatively condensed sediments representing deeper water environments than those underlying the surface of maximum starvation or the downlap surface. Hence, the time of maximum relative water depth, and probably of maximum coastal flooding, actually occurs above the surface of maximum starvation and in the basal portions of the highstand systems tracts (Fig. 2). In many instances, this represents an anoxic basinal sediment which may be highly enriched in organic matter (e.g., condensed black shale facies).

As noted by Baum and Vail (1988) and others, the sediments near the top of the transgressive systems tract and those near the base of the highstand systems tracts are relatively "time rich" and, therefore, are often grouped together as the condensed section. Condensed sections are of extreme importance in that they are very widespread in deeper water areas, often commonly show a concentration of biostratigraphically important fossils (e.g., foraminifera or conodonts), and typically can be traced into marine flooding or transgressive surfaces in shallow marine and nonmarine environments high on the continental shelf. The condensed section and, particularly, the surface of maximum starvation correspond to the time of maximum rate of sea level rise.

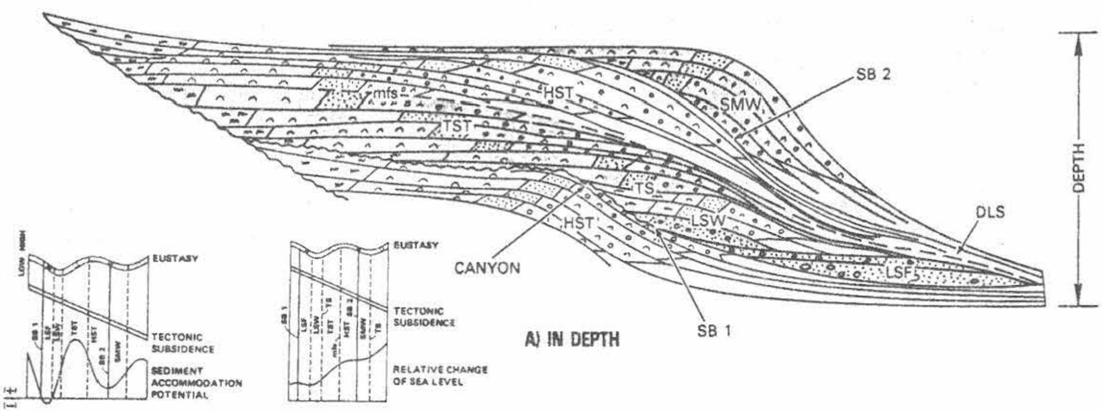
A point not brought out by most sequence stratigraphers is that the surface of maximum starvation (downlap surface) can also be a surface of extensive submarine erosion. This results both from the very low sedimentation rates associated with maximum rate of sea level rise, and from turbulent processes that apparently occur along water mass boundaries at the pycnocline (Ettensohn, 1987; Woodrow, 1985; Baird and Brett, 1986, 1988, and in press).

The remaining portions of the highstand systems tract, above the condensed section, consist of a generally static to upward shallowing (aggradational to progradational) set of parasequences culminating at the top in shallow water deposits (Figs. 1, 2). The upward shallowing may result either from sedimentary processes (e.g., progradation of sediment wedges) or more probably from initial lowering of relative sea level or at least diminishing rate of sea level rise. This results in a reduction of accomodation space and consequent progradation of strandline deposits. The highstand systems tract is typically bounded at its top by an erosional unconformity marking the lower boundary of the overlying sequence.

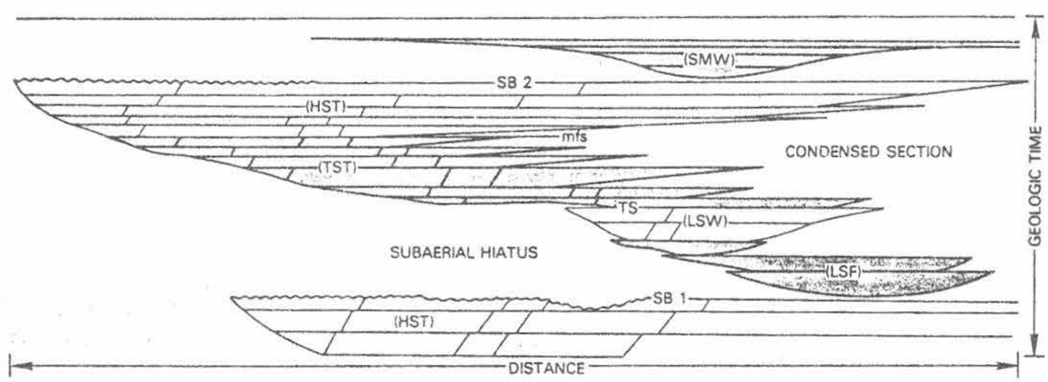
Figure 1. Idealized sequence stratigraphy model showing key surfaces and systems tracts. Sequence boundary 1 (SB-1) is a type 1 (incised) boundary, SB-2 is a type 2 sequence boundary. Note that the sequences are separated by an unconformity representing exposure during times of maximum rate of sea level fall - this unconformity can be traced to continuity offshore. This correlative conformity is a time boundary. From Baum and Vail (1988).

SEQUENCE STRATIGRAPHY DEPOSITIONAL MODEL

SHOWING SURFACES, SYSTEMS TRACTS AND LITHOFACIES



A) IN DEPTH



B) IN GEOLOGIC TIME

LEGEND

SURFACES

(SB) SEQUENCE BOUNDARIES
- (SB 1) = TYPE-1
- (SB 2) = TYPE-2

(DLS) DOWNLAP SURFACES
- (mfs) = maximum flooding surface

(TS) TRANSGRESSIVE SURFACE
- (First flooding surface above maximum regression)

SYSTEMS TRACTS

HST = HIGHSTAND SYSTEMS TRACT
TST = TRANSGRESSIVE SYSTEMS TRACT
LST = LOWSTAND SYSTEMS TRACT
- LSF = LOWSTAND FAN
- LSW = LOWSTAND WEDGE

SMW = SHELF MARGIN WEDGE SYSTEMS TRACT

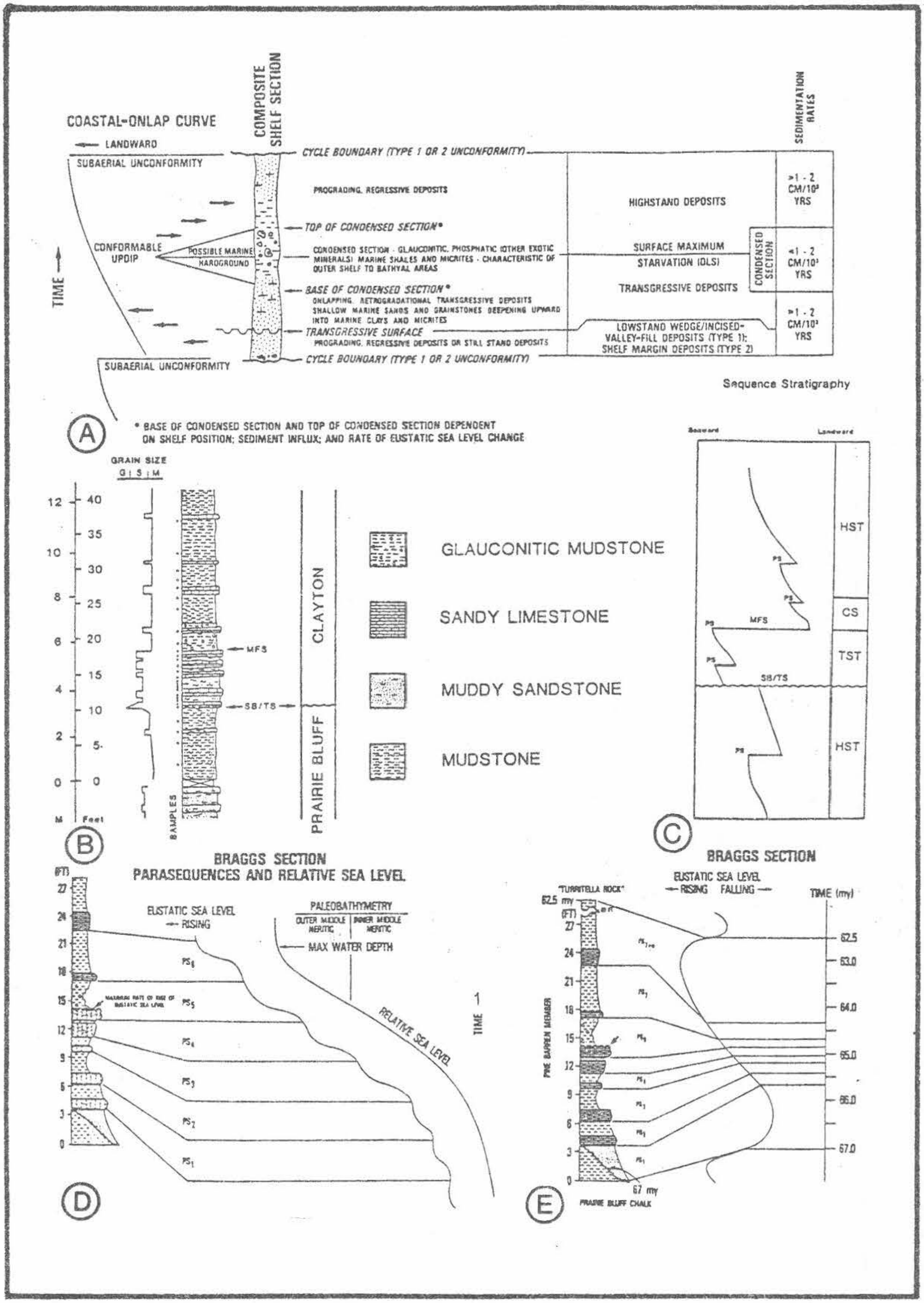
LITHOFACIES

- SUPRATIDAL
- PLATFORM
- PLATFORM-MARGIN GRAINSUPPORTSTONE/REEFS
- MEGABRECCIAS/SAND
- FORESLOPE
- TOE-OF-SLOPE/BASIN

Unconformity bound sedimentary packages have begun to be recognized in epicontinental sea deposits of the Paleozoic (e.g. Ross and Ross, 1985; 1988). Furthermore, a hierarchy of depositional cycles (1st to 6th order), apparently of eustatic origin, has been recognized particulary in Carboniferous deposits of the North American continental interior, but also elsewhere (Bush and Rollins, 1984; Heckel, 1986). Small scale (parasequence-like) features recorded as shallowing upward cycles or puntuated aggradational cycles (PACs) have been recognized in several portions of the stratigraphic column (Goodwin and Anderson, 1985).

How can these differing scales and types of cycles be reconciled with the sequence stratigraphy model of seismic stratigraphers? And how do sequences on epicontintental seas or foreland basins compare with those recognized on continental margins? Can the three primary surfaces of the idealized sequence, i.e. the basal erosion surface, transgressive surface, and surface of maximum starvation (downlap surface), be recognized in foreland basin/epicontinental sea areas? Answers to these questions require thorough re-examination of epicratonic strata. Most epicratonic successions are too thin to be resolved by seismic methods. Therefore application of sequence stratigraphic principles requires detailed field and subsurface correlation of strata in well-studied areas, as in the case study presented herein.

Figure 2. Stratigraphic profiles illustrating outcrop appearance of sequences. (A) Idealized model showing positions of three key surfaces; note that transgressive surface (TS) and cycle boundary tend to merge in shallow shelf areas; also note that condensed section spans late parts of the transgressive deposits and early highstand deposits; it includes the surface of maximum starvation (also referred to as downlap surface), which develops during times of maximum rate of sea level rise. From Baum et al. (1984). (B-E) Lithostratigraphy, sequence stratigraphy and inferred relative (D) and eustatic sea level fluctuations for Cretaceous-Tertiary boundary sediments exposed at Braggs, Alabama. Note that sequence boundary (SB) at Prairie Bluff/Clayton Contact) and first transgressive surface (TS) are merged. Also note condensed section (CS) overlying marine flooding surface (MFS), and separating transgressive (TST) and highstand (HST) systems tracts. PS1-PS6 are parasequences. Maximum relative and eustatic sea level stands occur above the surface of maximum starvation (MFS: noted with arrow) in early highstand; minimum relative sea level coincides with early transgressive deposits. Figures A, D and E from Baum and Vail (1988), figures B and C from Donovon et al. (1988).


COASTAL-ONLAP CURVE
LANDWARD
SUBAERIAL UNCONFORMITY
COMPOSITE SHELF SECTION
CYCLE BOUNDARY (TYPE 1 OR 2 UNCONFORMITY)
PROGRADING, REGRESSIVE DEPOSITS
TOP OF CONDENSED SECTION*
CONFORMABLE UPDIP
POSSIBLE MARINE HARDGROUND
CONDENSED SECTION - GLAUCONITIC, PHOSPHATIC (OTHER EXOTIC MINERALS) MARINE SHALES AND MICRITES - CHARACTERISTIC OF OUTER SHELF TO BATHYAL AREAS
BASE OF CONDENSED SECTION*
ONLAPPING, RETROGRADATIONAL TRANSGRESSIVE DEPOSITS SHALLOW MARINE SANDS AND GRAINSTONES DEEPENING UPWARD INTO MARINE CLAYS AND MICRITES
TRANSGRESSIVE SURFACE
PROGRADING, REGRESSIVE DEPOSITS OR STILL STAND DEPOSITS
CYCLE BOUNDARY (TYPE 1 OR 2 UNCONFORMITY)
SUBAERIAL UNCONFORMITY
TIME
HIGHSTAND DEPOSITS
SURFACE MAXIMUM
STARVATION (DLS)
TRANSGRESSIVE DEPOSITS
CONDENSED SECTION
LOWSTAND WEDGE/INCISED-VALLEY-FILL DEPOSITS (TYPE 1); SHELF MARGIN DEPOSITS (TYPE 2)
SEDIMENTATION RATES
>1 - 2 CM/10³ YRS
<1 - 2 CM/10³ YRS
>1 - 2 CM/10³ YRS
Sequence Stratigraphy
A
* BASE OF CONDENSED SECTION AND TOP OF CONDENSED SECTION DEPENDENT ON SHELF POSITION; SEDIMENT INFLUX; AND RATE OF EUSTATIC SEA LEVEL CHANGE
GRAIN SIZE
G S M
M Feet
SAMPLES
MFS
SB/TS
CLAYTON
PRAIRIE BLUFF
GLAUCONITIC MUDSTONE
SANDY LIMESTONE
MUDDY SANDSTONE
MUDSTONE
B
Seaward
Landward
HST
CS
TST
HST
PS
MFS
SB/TS
C
BRAGGS SECTION
BRAGGS SECTION
PARASEQUENCES AND RELATIVE SEA LEVEL
EUSTATIC SEA LEVEL
RISING
PALEOBATHYMETRY
OUTER MIDDLE NERITIC
INNER MIDDLE NERITIC
MAX WATER DEPTH
RELATIVE SEA LEVEL
TIME
PS1
PS2
PS3
PS4
PS5
PS6
D
EUSTATIC SEA LEVEL
RISING FALLING
"TURRITELLA ROCK"
62.5 my
TIME (my)
62.5
63.0
64.0
65.0
66.0
67.0
PINE BARREN MEMBER
67 my
PRAIRIE BLUFF CHALK
E

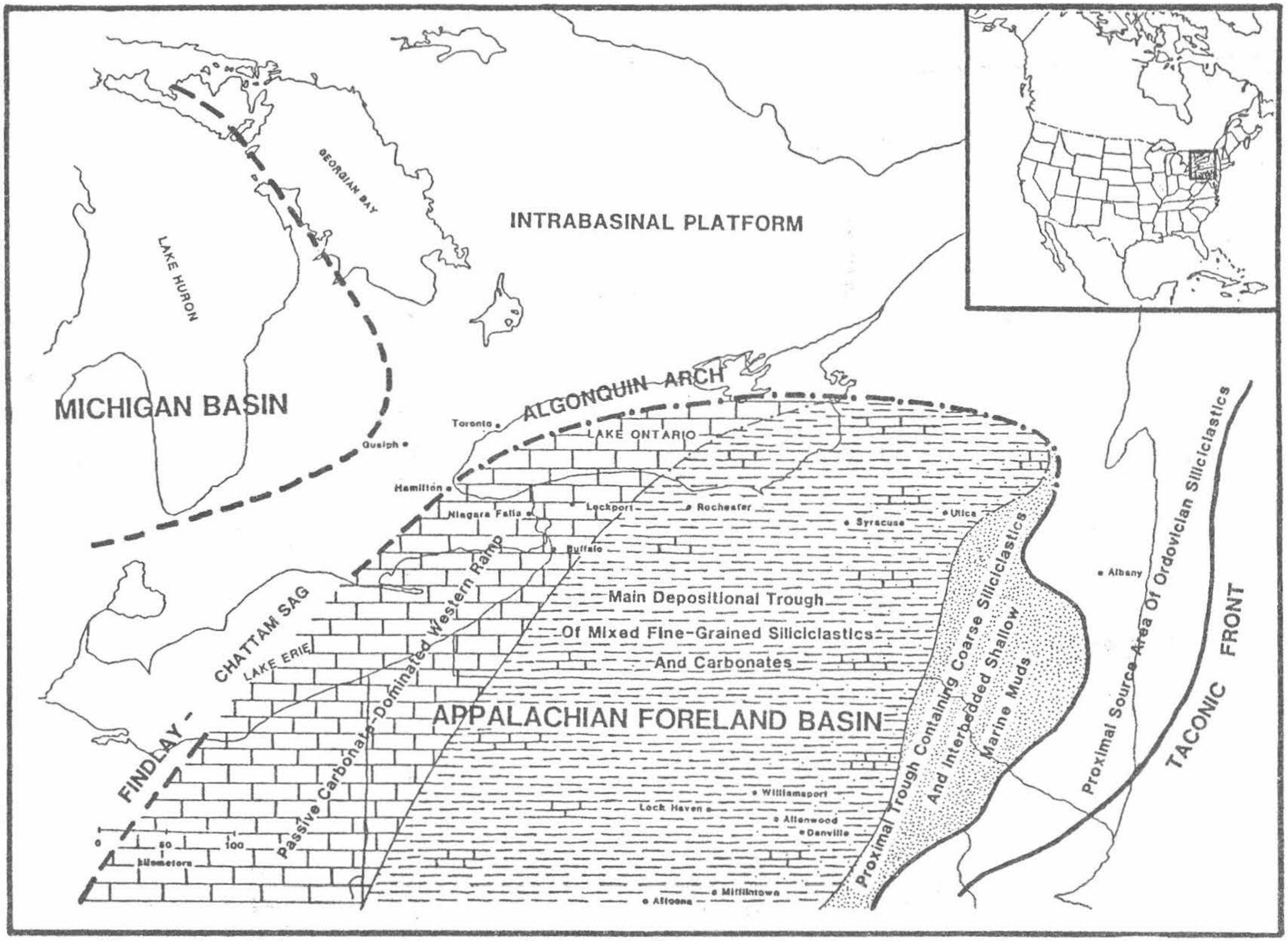


Figure 3. Paleogeographic map of the northern Appalachian foreland basin during medial Silurian time; inset map shows position of the study area within North America. Modified from Brett (1983).

Sat. C8

GENERAL GEOLOGICAL SETTING

Paleogeography and Sedimentology

The Medina, Clinton, and Lockport Groups of New York State and Ontario and their correlatives in Pennsylvania and Ohio were deposited in the northern end of the Appalachian Foreland Basin (Fig. 3). Along the southeastern margin, the basin was bordered by a linear belt of uplifted Middle to Upper Ordovician shales and sandstones and, beyond, by the Taconic orogenic belt from which Silurian siliciclastics were derived. The Appalachian Basin was bordered along its northwest margin by the northern extension of the Findlay Arch, locally referred to as the Algonquin Arch (Fig. 3).

During the Silurian Period, the northern reaches of the Appalachian Basin occupied a position approximately 20-25 degrees south latitude (Van der Voo, 1988). Given this position, it is likely that the basin was subjected to tropical storms following southerly paths (Middleton, 1987). Signatures of episodic sedimentation are common in the Llandoverian through Ludlovian Series of western New York.

These strata comprise a 130 meter thick, mixed carbonate-siliciclastic succession representing multiple, small to medium-scale depositional systems (Fig. 3). Near the southeastern strandline, the sequence is dominated by coarse-grained siliciclastics which were deposited on a storm-dominated shallow shelf and shoreface (Zenger, 1971; Cotter, 1983). The middle to outer regions of the eastern basin flank are represented by argillaceous limestones and sandy or silty shales containing coarser-grained, storm event beds with unidirectional (generally NW-directed) current indicators. Across the basin axis and up the western ramp, the succession becomes increasingly carbonate-dominated. Proximal to the Algonquin Arch, the Medina-equivalent Brassfield Formation of Ohio, and the Clinton and Lockport Groups of southwestern Ontario consist nearly entirely of carbonates, mainly dolomitic, crinoid and brachiopod grainstones, representing normal wave base environments.

Basin Tectonics

The Appalachian Foreland Basin developed from compression of a passive, carbonate-dominated, continental margin during collision with an island arc system during the Middle Ordovician. Early phases of the Taconic Orogeny are recorded by thick turbidite sequences of the Martinsburg (PA) and Snake Hill-Frankfort (NY) Formations. The waning phases of the main orogenic pulse during the Late Ordovician (Ashgillian) are recorded by the Bald Eagle-Oswego sandstone wedge and the overlying Juniata-Queenston red bed sequences.

In many parts of the Appalachian Basin there is evidence of a Late Ordovician to Early Silurian tectonic rejuvenation of the Taconic Front (Quinlan and Beaumont, 1984; Tankard, 1986). In New York State and Pennsylvania, evidence for a late Taconic pulse lies in the regionally extensive, low angle unconformity at the Ordovician-Silurian boundary (Cherokee Unconformity) and an overlying thick Early Silurian clastic wedge. Medial Silurian strata become finer grained indicating a period of quiescence between Taconic and Salinic orogenic pulses.

The Salinic Orogeny was a brief and relatively local uplift in eastern New York State which occurred during the Ludlovian Epoch. The signature of the Salinic Disturbance is an unconformity which ultimately separates the Pridolian Syracuse Formation and upper Ordovician strata. Lower and medial Silurian strata are missing completely east of Van Hornesville, New York.

Other evidence for the Salinic may be recorded in the Bloomsburg-Vernon red bed sequence, and, perhaps more definitively, in the overlying "upper Shawangunk Tongue" of southeastern New York and northeastern Pennsylvania (Prave et al., 1989). It appears that the upper Shawangunk Tongue consists of cannibalized middle Silurian sediments (e.g. Herkimer Sands) eroded from the northern Hudson Valley which were transported southward.

There is little sedimentological or stratigraphic evidence for major unconformities in latest Silurian strata. However, Salkind (1979) has argued for Latest Silurian tectonic activity based upon structural relationships in the Hudson Valley.

SILURIAN SEQUENCES AND BASIN HISTORY

The Llandoverian to Ludlovian (Medina through Lockport Group) succession in New York State, Ontario, Ohio, and Pennsylvania is divisible into at least six large-scale, unconformity-bounded stratal packages (Fig. 4) comparable in some respects to depositional sequences of seismic stratigraphers (e.g., Vail et al., 1977; Van Wagoner et al., 1988). For the most part, these sequences correspond to previously recognized group-level stratigraphic units, but in some cases, cut slightly across traditional subdivisions. The first sequence corresponds to the Medina Group, the second and third to the lower and middle portions of the Clinton Group, respectively, the fourth and fifth to parts of the upper Clinton Group, and the sixth to the Lockport Group and the Vernon Formation.

The sequences recognized herein are at least crudely divisible into systems tracts analogous to those within previously recognized sequences of seismic stratigraphers (see for example, Vail et al., 1977; Van Wagoner et al., 1988; Posamentier et al., 1988). In terms of temporal magnitude, Silurian <u>sequences</u>, like

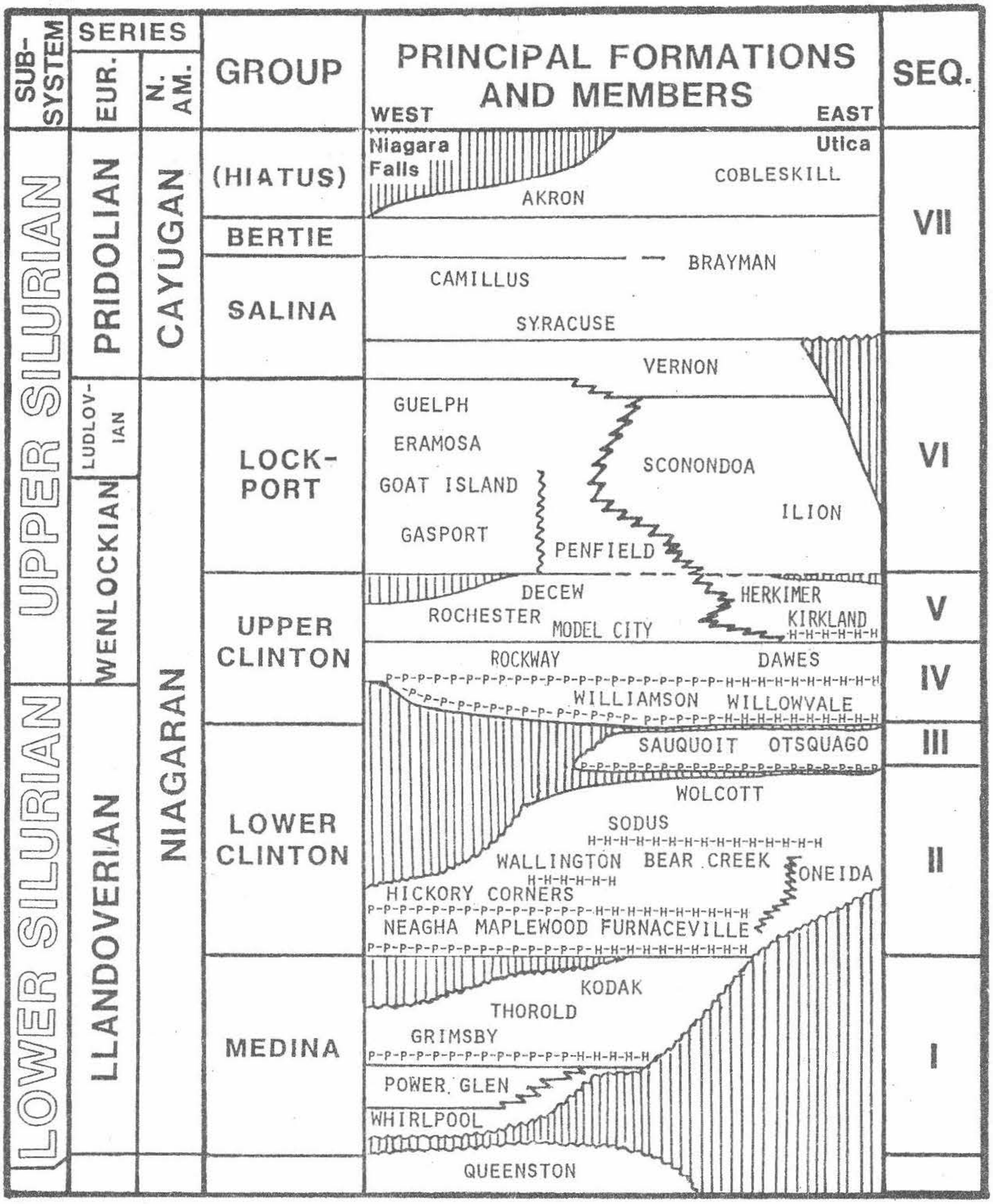


Figure 4. General time stratigraphic chart showing Silurian sequences of New York that are recognized on the basis of major, bounding unconformities.

those of the seismic stratigraphers, encompass about 1 to 5 million years and, therefore, can be classified as 3rd order cycles (Vail et al., 1977). The Silurian sequences are divisible internally into very prominent and basin-wide subsequences, which are of lesser temporal magnitude and display sharp, slightly erosive, disconformities. Each sequence contains 2-5 subsequences. We estimate, therefore, a subsequence recurrence interval on the order of 0.8 to 1.5 million years, coinciding with 4th order cycles or synthems (see Ramsbottom, 1979; Busch and Rollins, 1984). In a crude sense, the subsequences mark out lowstand, transgressive, early highstand, and late highstand systems tracts (see Van Wagoner et al., 1988). Sequences and subsequences are distinctive in having sharp bases at which typically shallower water deposits are juxtaposed over deeper water sediments of the stratal package. These boundaries may be nearly conformable or may be regionally angular unconformities.

In turn, the subsequences are made up of smaller-scale subdivisions that correspond approximately with submembers or even beds in the lithostratigraphic terminology. Each subsequence is divisible into two to three minor regressive, transgressive or shallowing, deepening cycles that may be nearly symmetrical to markedly asymmetrical nature. These parasequence sets, apparently represent 5th order cycles or cyclothems (Bush and Rollins, 1984; Heckel, 1986).

The smallest scale subdivisions of the 5th order cycles tend to be more numerous (typically three to five per cyclothem) and most commonly show an asymmetrical shallowing-upward pattern, in contrast to the larger scales of cyclicity. These correspond to PACs or 6th order cycles (Goodwin and Anderson, 1985).

In the following sections we describe the six major sequences and then discuss broader implications of the stratigraphic patterns and processes.

Sequence I: Medina Group

The first sequence nearly overlaps with traditional definitions of the Medina Group in New York State and parts of Ontario and Ohio, and of the Tuscarora Sandstone in Pennsylvania (Cotter, 1982; Duke, 1987; Figs. 4-8). Its lower boundary coincides with the major Hirnantian unconformity that marks the Ordovician-Silurian contact over much of eastern North America. Dennison and Head (1975) labeled this erosion surface the Cherokee Unconformity and used it to subdivide Sloss' (1963) Tippecanoe Sequence into a lower "Creek" (Middle to Upper Ordovician) and an upper "Tutelo" succession (Silurian-Lower Devonian).

The Cherokee Unconformity is a gently north-westward sloping, nearly planar surface (Middleton et al., 1987). In the central Appalachians, the unconformity is manifest by the basal contact of the Tuscarora Formation on the Martinsburg Shale or the Juniata Formation (Cotter, 1983; Fig. 5).

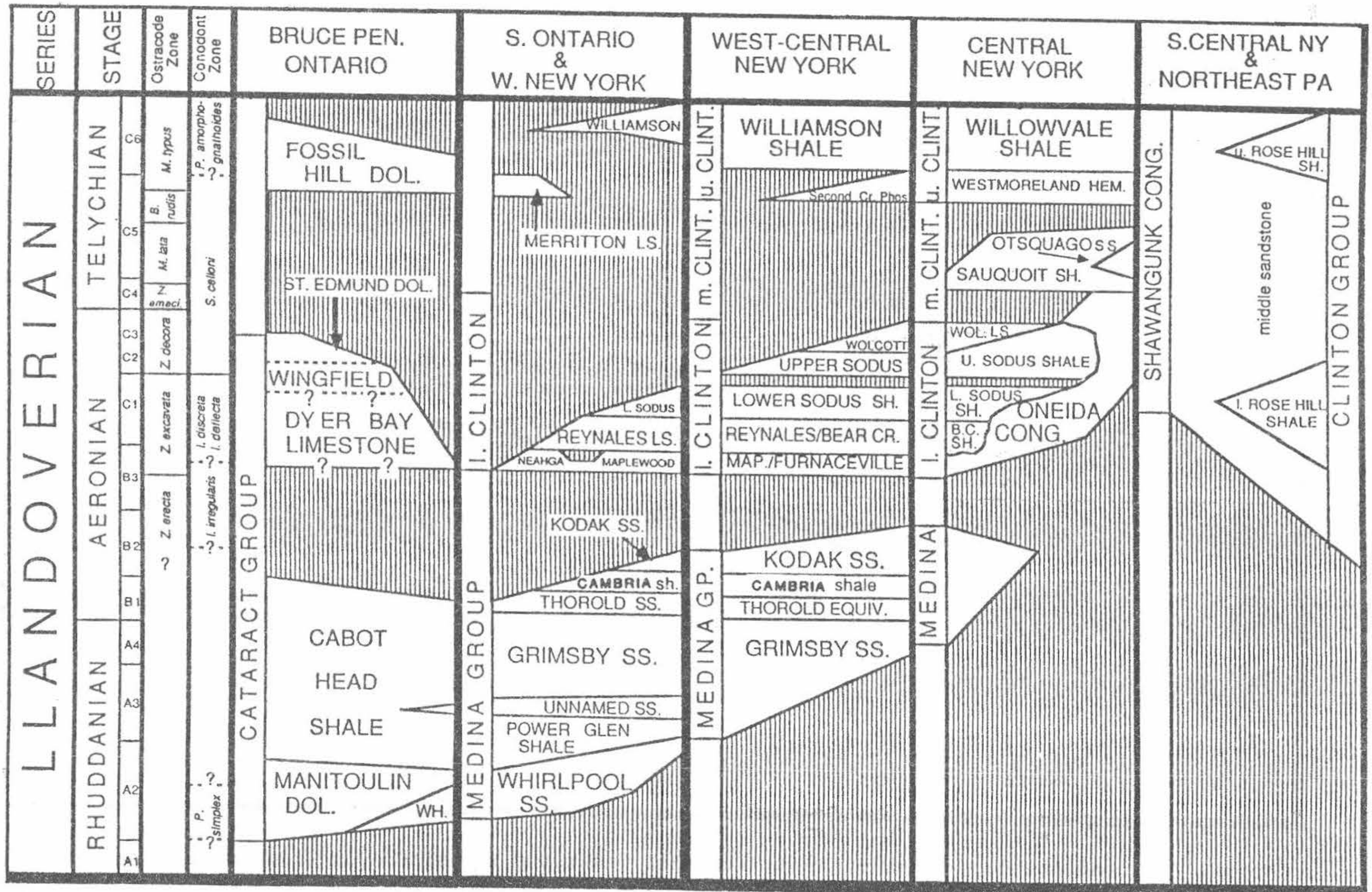


Figure 5. Chronostratigraphic correlations of Lower Silurian (Llandoverian) strata of sequences I and II in Ontario (Bruce Peninsula), western, central and eastern New York State. Vertical ruling indicates unconformities.

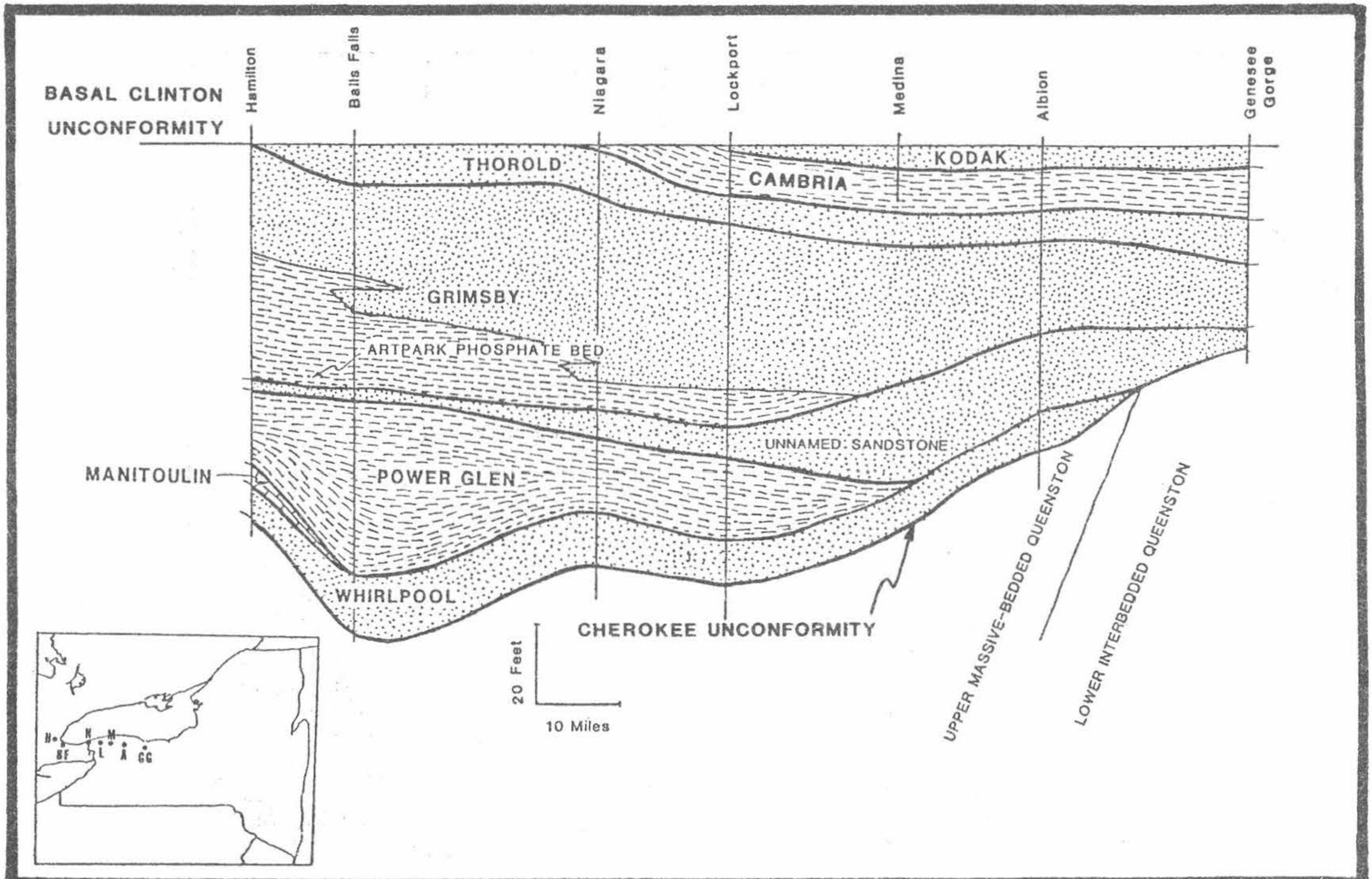


Figure 6. Regional cross-section of Medina Group (Sequence I) in western New York and southern Ontario. Note regional, eastward truncation of Ordovician units by basal Cherokee unconformity. Also note eastward pinching of lower Medina units (Whirlpool, Power Glen Formation) and somewhat complementary westward overstep of upper Medina units under regionally angular sequence I-II boundary unconformity.

In the central Appalachians the Cherokee Unconformity may not coincide precisely with the Ordovician-Silurian boundary. In New York State, the Cherokee Unconformity is a sharp surface under which occurs progressive eastward bevelling of Upper and Middle Ordovician strata (Queenston Shale downward to Frankfort or Schenectady Formation) and above which lie various Lower Silurian units.

The Cherokee Unconformity is nearly planar in New York and Ontario sections; i.e., a Type 2 unconformity (Van Wagoner et al, 1988). The unconformity surface may have been generated by the combined effects of a glacioeustatic regression and uplift along the tectonically active, southeastern basin margin. The Cherokee Unconformity has long been related to a major Late Ordovician sea-level lowstand caused by glaciation in North Africa (Dennsion and Head, 1975; Johnson, 1988). However, the geometry and magnitude the unconformity is not readily explained by sea-level change alone (Middleton et al., 1987). The angularity of the unconformity (Figs. 4-5) suggests that there must have been some tectonic rejuvenation of the Taconic Front in southeastern Pennsylvania late in the Ordovician or early in the Silurian as suggested by Quinlan and Beaumont (1984) and Tankard (1986).

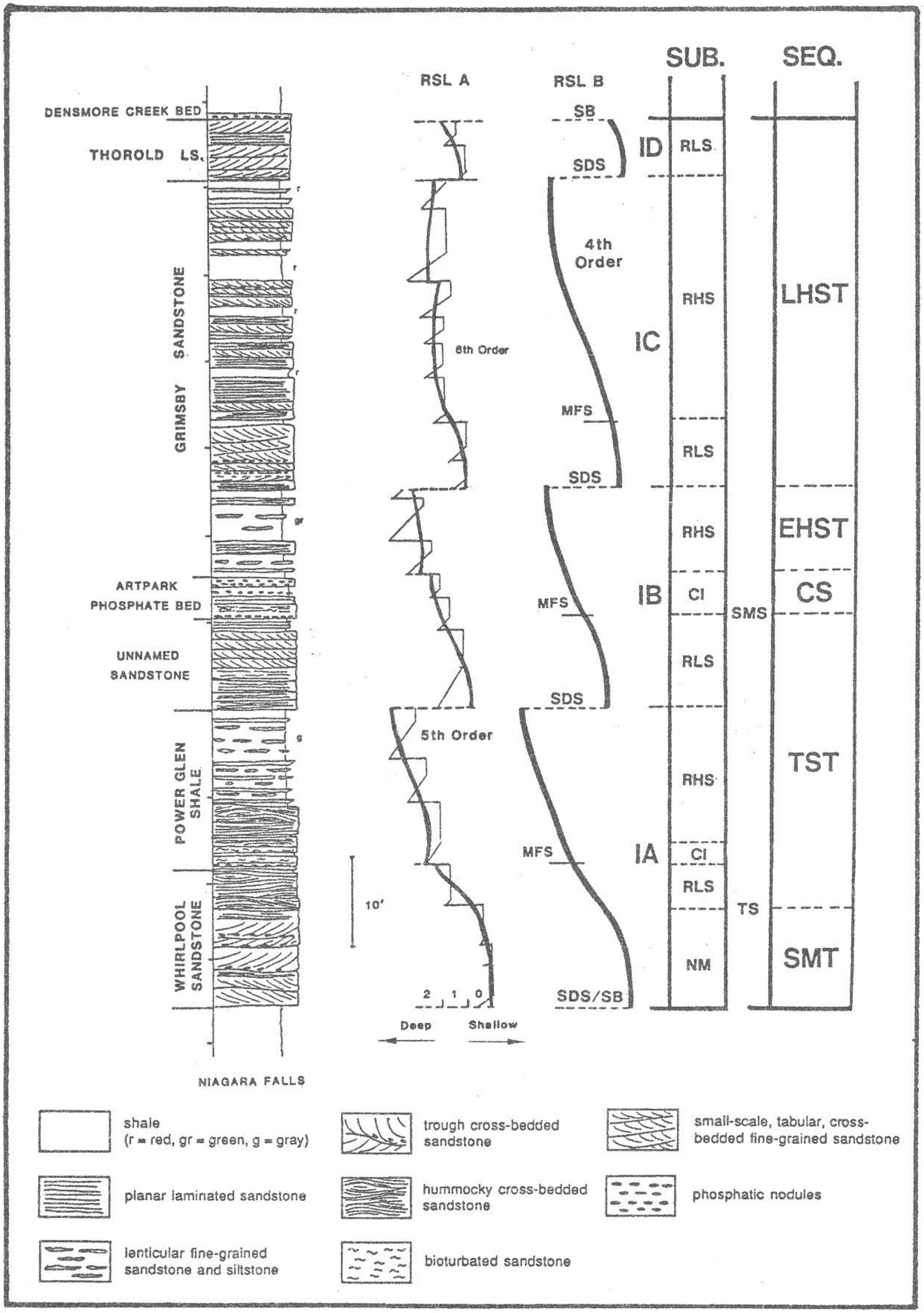
In western New York and the Niagara Peninsula of Ontario, the basal beds of Sequence I comprise the Whirlpool Sandstone, a relatively thin (5-8 meter) but widespread quartz arenite unit (Figs. 6, 7). Recent study of the Whirlpool by Middleton et al. (1987) demonstrates that the formation is subdivisible into two members; a lower unit of medium-grained, large-scale trough cross-bedded, white, quartzose sandstone which rests sharply on the underlying Late Ordovician Queenston red mudstones, and an upper lenticular bedded, fine-grained, white quartzose sandstone with green shale interbeds. Absence of marine indicators (fossil spores, but no acritarchs, A.J. Boucot, pers. comm., 1989) and consistent northwestward orientation of cross-beds led Middleton et al. (1987), to postulate a braided fluvial depositional setting for the lower member. This unit is inferred to represent a lowstand deposit (technically this is a "shelf margin systems tract" in the terminology of type 2 sequences; see Van Wagoner et al., 1988; but that term is entirely inappropriate here), that accumulated in a gently northeastward sloping braid plain prior to the major Early Silurian (Rhuddanian or early Aeronian (A-3 to B?) rise of sea-level (Johnson, 1988).

The upper 1-2 meters of the Whirlpool Sandstone display numerous indicators of deposition in shallow marine influenced environments. In some instances, green shale partings within trough cross-bedded sandstones near the top of the lower Whirlpool division contain Silurian marine acritarchs (M. Miller, unpubl. data) as well as reworked Ordovician palynomorphs. These occurrences suggest backfilling of relict tidal (?) channels during the initial sea-level rise. Macrofossils, including corals, lingulid brachiopods and echinoderms, and trace fossils have been noted in the upper Whirlpool beds of southwestern Ontario, In addition, the presence of hummocky cross-stratification (HCS) indicates deposition in shallow, storm-influenced shelf settings.

In eastern sections, the Whirlpool is separated from the heterolithic dark gray shale and tempestitic sandstone succession of the overlying lower Power Glen Formation, by a cryptic discontinuity (Figs. 7, 8). The discontinuity becomes more recognizable in a basinward (westward) direction. In the Niagara Peninsula the subtle discontinuity, marked by a thin (30 cm) calcareous, phosphatic, fossiliferous (crinoids, asteroids, brachiopods, bryozoans) sandstone, occurs at or near the top of the Whirlpool Formation. This bed appears to grade westward into the glauconitic top of the Manitoulin Dolomite, an argillaceous to arenaceous, fossiliferous carbonate with interbedded shales and hummocky cross-bedded siltstone/fine-grained sandstones. The Manitoulin contains a relatively diverse body and trace fossil assemblage, recording fully marine conditions. We interpret the Manitoulin and its correlative eastward condensed phosphatic beds as a signature of a relatively abrupt sea-level rise; the sea-level rise resulted in a brief period of siliciclastic sediment starvation during which the phosphatic pebble bed was generated on a marine-flooding surface (Fig. 7).

Figure 7. Lithostratigraphy, inferred relative sea level curve, and sequence terminology for the Medina Group (Sequence I) at Niagara Gorge, Lewiston, New York. RSL-A = relative sea level curve for higher lower order cycles; 6th order = small scale, shallowing upward cycles, thin line; 5th order, larger scale, subsymmetrical to upward deepening cycles of member scale; scale calibrated on fossil benthic assemblages. 0 = non-marine; 1 = benthic assemblage-1, lingulid and trace fossil community; 2 = benthic assemblage-2; fully marine shallow shelf fauna. RSL-B = relative sea level curve for large scale fourth order cycles, generally asymmetrical upward deepening cycles of subgroup scale.

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). Abbreviations 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.


DENSMORE CREEK BED
THOROLD LS.
GRIMSBY SANDSTONE
ARTPARK PHOSPHATE BED
UNNAMED SANDSTONE
POWER GLEN SHALE
WHIRLPOOL SANDSTONE
NIAGARA FALLS
RSL A
RSL B
SUB.
SEQ.
SB
SDS
ID
RLS
4th Order
6th Order
5th Order
RHS
LHST
IC
MFS
EHST
IB
CI
CS
SMS
TST
IA
TS
NM
SMT
SDS/SB
10'
2 1 0
Deep
Shallow
shale (r = red, gr = green, g = gray)
trough cross-bedded sandstone
small-scale, tabular, cross-bedded fine-grained sandstone
planar laminated sandstone
hummocky cross-bedded sandstone
phosphatic nodules
lenticular fine-grained sandstone and siltstone
bioturbated sandstone

The overlying beds (basal Power Glen dark shales, in New York and lower Cabot Head Shale in Ontario) reflect a time of maximum marine flooding and may represent the base of a highstand systems tract. The Power Glen contains offshore dysaerobic to aerobic marine muds with scarce to diverse faunas, and in the eastern, more proximal facies, thin siltstones, gutter casts and other indicators of storm deposition (Duke and Fawcett, 1987, in press; Duke et al., 1987).

The Power Glen-Lower Cabot Head succession is composed of at least two upward-coarsening, shale-sandstone cycles (parasequences). The top of the upper cycle is abruptly overlain by a prominent 2-3 m thick quartzose sandstone in western New York (= Devils Hole sandstone of Duke et al., in prep.). Locally, the upper portion of this sandstone is calcareous (dolomitic) and contains abundant phosphatized fossils (bryozoans, gastropods) and pebbles. At some localities, this bed is hematitic. This horizon, the Artpark Phosphate Bed, appears to represent a major marine flooding surface (Figs. 6-8). Although the basal 1-2 m of the Grimsby are green, marine shales resembling the Cabot Head, Grimsby beds are mainly maroon shales, and red, pink and white mottled sandstones. Duke and Brusse (1987) have recognized at least five to six parasequence-scale (1-3 m) upward coarsening cycles within the Grimsby succession of western New York. More complete successions from the base upward, consist of thin, fossil rich (typically lingulid and rhynchonellid brachiopods, bivalves and bryozoans) sometimes phosphatic beds, overlain by a succession of maroon shales with thin to medium bedded sandstones with hummocky cross stratification (HCS), and capped by almalgamated storm sandstones. In some parasequences the top beds are heavily bioturbated (Arthrophycus, Daedalus). At some localities, local channels filled with red sandstone may cut out upper parts of parasequences.

The Grimsby sequence is an overall shallowing succession considered to represent progradation of shallow, marine-tidal flat sands over marine muds (Martini, 1971; Duke and Fawcett, 1987). The shell rich phosphatic lags record minor marine flooding surfaces bounding parasequences. The shallowing trend of the Grimsby Formation culminates in a maximally regressive, laterally extensive sheet sandstone, the Thorold Formation. Thorold facies range from proximal, red, moderately to heavily bioturbated, tidal channel sandstones to fully subtidal interbedded white sandstones exhibiting ball and pillow structures and thin green silty shales.

The upper part (supra-Thorold) of the Medina Group is a slightly more marine influenced, 2-3 m thick red and green shale with abundant leperditian ostracodes; we refer to this shale, informally as the Cambria Member for the exposure along Lockport Junction Road on the eastern townline of Cambria, New York (STOP 2A; Fig. 8). The shale is capped by a second, widespread, heavily bioturbated quartzose sandstone, the Kodak Formation. The distal (westward) end of this sequence is truncated by the upper bounding

SUBSEQ.	DEPO. PHASE	ONTARIO	WESTERN NEW YORK	CENTRAL OHIO
IE		SEQUENCE I / II UNCONFORMITY		
	RLS		KODAK SS.	? ?
ID	RHS		unnamed shale	
	RLS	?	THOROLD SS.	"stray Clinton" Ss.
IC	RHS	u. CABOT HEAD SH. / SS.	upper GRIMSBY SS.	"red Clinton" Ss.
	RLS	unnamed sandstone	unnamed sandstone	
IB	RHS	u. CABOT HEAD SH.	lower GRIMSBY SH. / SS.	
			Art Park phosphate	? ?
	RLS	m. CABOT HEAD SH.	unnamed sandstone	"white Clinton" Ss.
IA	RHS	l. CABOT HEAD SH.	POWER GLEN SH.	CABOT HEAD SH.
	RLS	MANITOULIN DOL.	u. WHIRLPOOL SS.	BRASSFIELD LS.
	NM		l. WHIRLPOOL SS.	

Figure 8. Summary chart of subsequence interpretation of the Medina Group (sequence I) in Ontario, western New York and Ohio. Units IA-IE are considered to represent subsequences. Subsequence terminology, as in Figure 4. Vertical scale does not represent thickness but estimated temporal duration of units.

unconformity of Sequence I. However, the proximal facies are preserved in the Rochester, New York area, and record the easternmost incursion of marginal marine faunas in the Medina Group.

Sequence II: Lower Clinton Group

Although the Clinton Group has long been accepted as a useful stratigraphic subdivision of the medial Silurian in eastern North America, it is now evident that this unit is actually an amalgam of parts of four distinct sequences. Thus, we informally divide the Clinton into lower, middle, and upper Clinton units, approximately following the usage of Gillette (1947).

As defined herein, the lower Clinton constitutes a genetic, unconformity-bound sequence of strata ranging from the Maplewood-Neahga Shale (or equivalent Furnaceville Hematite) to the top of the Wolcott Limestone in the western and central New York sections (Fig. 4).

Sequence II is bounded at its base by a nearly planar (type 2), unconformity that is a very low angle regional truncation surface. Successively lower beds of the Medina Group are beveled in a westward direction, starting at least in the Rochester area, as described in the previous section. As such, this unconformity has the opposite sense of truncation from the major Cherokee Unconformity. Because only a few meters of Medina Group shales and sandstones are removed at this surface, the temporal magnitude of the sequence II bounding unconformity is thought to be relatively slight, no more than one or two ostracode zones. Nonetheless, this surface is considered to be significant in that it truncates strata and separates distinct, genetically related packages, the Lower Clinton above and the Medina below. In the Ontario and western New York sections, the upper boundary of sequence II is a more prominent truncation surface underlying the Llandoverian C-6 Williamson-Willowvale shales and their correlatives (see below).

The basal boundary of sequence II is clearly demarcated by a thin (1-10 cm) but regionally extensive phosphate pebble horizon designated the Densmore Creek Phosphate Bed (Brett et al., in prep.) for good exposures on Densmore Creek, near Rochester. The bed can be traced from Wayne County, New York, west to near St. Catharines, Ontario. Phosphatic steinkerns, including those of gastropods and brachiopods (Eocoelia, Hyattidina, and Leptaena), indicate a shallow marine, but sediment starved setting. Locally, this unit contains Trypanites bored clasts up to 10 cm across, of phosphatic, fossiliferous pelletal packstone and grainstone.

In Niagara County, the Densmore Creek Phosphate Bed forms the base of the Maplewood - Neahga Shale. The Neahga Shale, a 0.5 to 4 m dark gray, fissile shale, crops out from near St. Catharines, Ontario to Lockport, New York. The Neahga thins to near pinch out along a narrow NE-SW belt in Niagara County (Fig.

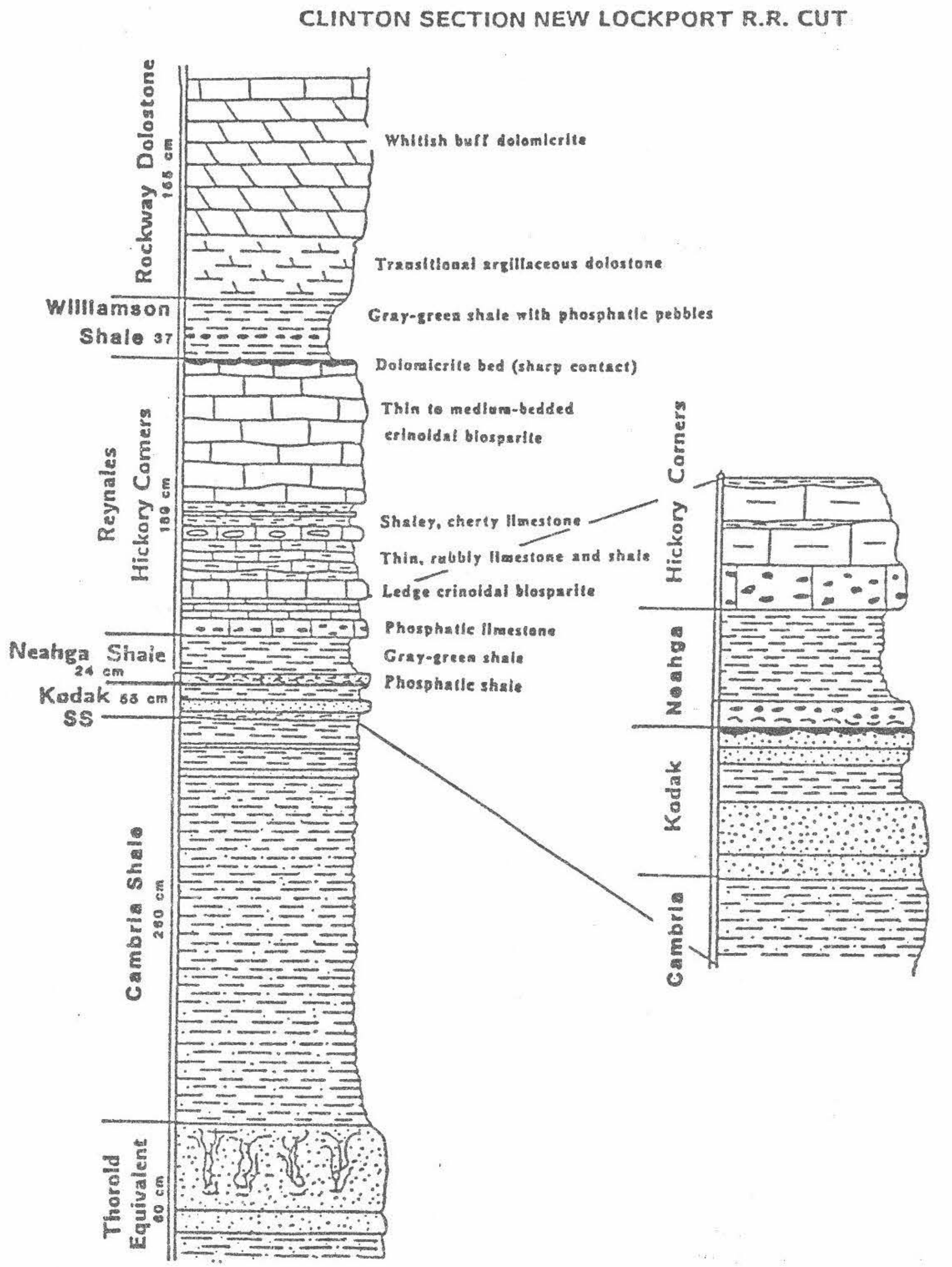


Figure 9. Stratigraphic section for uppermost Medina, lower Clinton, and base of upper Clinton.

10). Eastward of this region (approximately at Lockport, New York in the outcrop belt), the slightly lighter colored but equivalent beds are referred to as the Maplewood Shale. The Maplewood ranges up to 6 m in thickness near Rochester, but pinches out within 8 km in eastern Monroe county. This pinch-out, which trends approximately northeast to southwest, can be traced in the subsurface at least to Chatauqua County (Van Tyne, 1966). The Densmore Creek bed and a minor phosphatic limestone bed (Budd Road Bed, Brett et al., in prep.) above the top of the Maplewood merge as the Maplewood pinches out (Fig. 10). At this point, the composite phosphate bed becomes thicker (up to 30 cm), hematitic, and is referred to as the Furnaceville Iron Ore, a unit that contains multigenerational phosphatic conglomerates (LoDuca, 1988; LoDuca and Brett, 1990). Further east the Furnaceville merges into a thin quartz pebble conglomerate, the Oneida Formation.

The Densmore Creek Bed and the correlative Furnaceville Hematite are considered to represent sediment starved conditions associated with the initial transgression that initiates sequence II. The presence of an Eocoelia biofacies both within the shale and in the phosphate-hematite beds suggests a shallow subtidal (BA-2) setting.

Internally, sequence II is subdivided into two major (5 to 20m) sharply based stratal packages comprising shale to carbonate successions that appear to constitute upward-deepening fourth order cycles or subsequences, the Neahga (Maplewood Shale) to Reynales Limestone, and the Sodus Shale to Wolcott Limestone (Fig. 9). These stratal packages, in turn, are subdivisible into thin (1 to 6 m) but internally complex and condensed members that may be approximately analogous to fifth order cycles (Busch and Rollins, 1984); they also display deepening upward trends. These are, in ascending order: 1) the Maplewood/Neahga; 2) the Brewer Dock Member of the Reynales Formation, including the Seneca Park Hematite (Fig. 10); 3) the Wallington Member of the Reynales Formation(these three together forming the first fourth order cycle); 4) the Lower Sodus Shale; 5) the Upper Sodus Shale; 6) the Wolcott Limestone (together forming the second fourth order cycle). The base of each of these units is sharp, and, in some instances, demarcated by a very thin, phosphatic sandy horizon that appears to record an interval of sediment bypass and starvation associated with the onset of renewed transgression (LoDuca, 1988; LoDuca and Brett, 1990).

Only portions of the lower two formations (5th order cycles; parts of the 4th order subsequence III) are present in Niagara County, due to erosion at the top of sequence II (Figs. 9, 10). The basal unit, the Neahga Shale, consists of 0.5-2.0 m of dark gray, fissile shale dominated and contains a sparse Eocoelia biofacies. The Densmore Creek bed at the base of the Neahga shale also marks the transgressive surface of the Sequence II, as a whole.

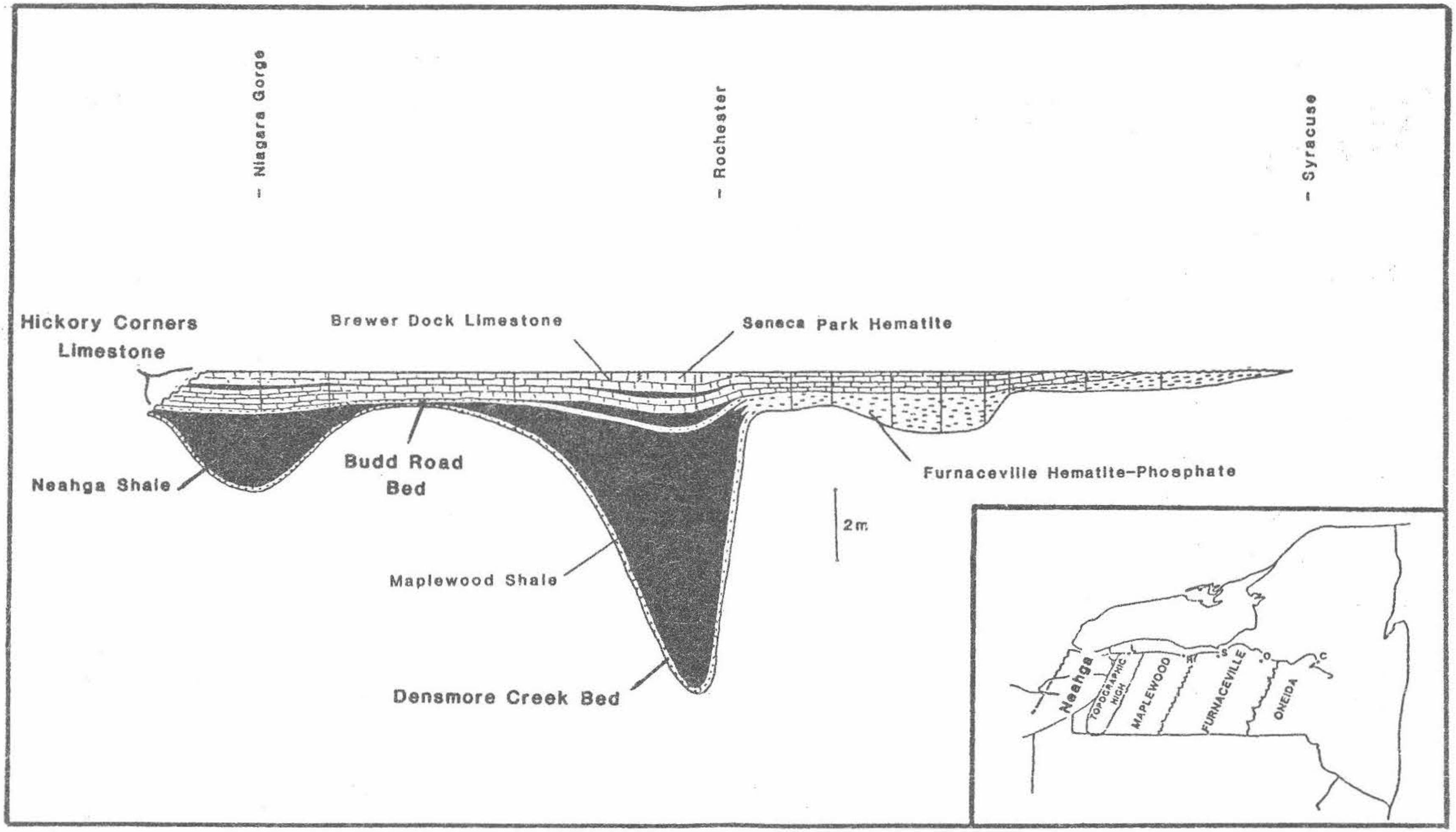


Figure 10. Regional east-west cross section of the basal units of Sequence II (lower Clinton Group) between Syracuse, New York and St. Catharines, Ontario. Note presence of two minor depocenters (for the Neahga and Maplewood shales, respectively) separated by a minor arch region; also note eastward passage of shales into highly condensed hematite and phosphate-bearing conglomerate (Furnaceville). Small inset map shows approximate orientation of extreme condensation and/or pinchout areas. West of Niagara County, the Neahga Shale is truncated by the major (sequence bounding) late Llandoverian unconformity.

The next 5th order cycle, as noted above, is demarcated at its base by a another thin phosphate pebble horizon designated the Budd Road Bed (Brett et al., in prep.). Overlying this bed are thin interbedded silty carbonates (greenish gray shale) of the Brewer Dock Member of the Reynales Formation. This cycle is capped by a thin (< 1m) crinoid/bryozoan rich, cross bedded unit which is locally hematitic (Fig. 10). A discontinuity may separate this cycle from the next based on conodont biostratigraphy (Kleffner, pers. comm., 1989), but does not appear to be substantiated by other zonations.

The third 5th order cycle, corresponding to the Wallington Limestone Member, begins abruptly with thin (< 0.5 m) green Eocoelia-bearing shales and passes upward into heterolithic argillaceous silty carbonates and shales. In western New York, the Wallington-equivalent beds are mainly crinoidal packstones and grainstones. In the Niagara region the Brewer Dock and truncated lower Wallington equivalents are difficult to distinguish and have, together, been termed the Hickory Corners Member (Kilgour, 1963). However, there still appears to be some justification for recognizing the two members.

It should be noted that, in most cases, only the eastern more proximal facies of lower Clinton sequence cycles are preserved. The western presumably deeper water facies have been removed in the New York - Ontario outcrop belt due to the major Late Llandoverian erosion surface that forms an upper boundary to sequence II. Sequence II is completely absent in the outcrop belt between Grimsby, Ontario and Owen Sound, (Fig. 11) Ontario where Sequence IV carbonates (Merriton - Fossil Hill) immediatly overlie the Medina Group (see below).

Sequence III: Middle Clinton Group

Sequence III is composed of the Sauquoit - Otsquago Formations in Central New York State, a 40 m-thick succession of greenish gray shales, with thin sandstone and conglomerate layers, which outcrops sporadically from near Cayuga to Madison County. No strata belonging to this sequence occur in western New York or Ontario.

Sequence IV: Basal Upper Clinton Group

The fourth major unconformity-bounded stratal package constitutes the upper portion of the Clinton Group as presently defined. In the axial portions of the Appalachian Basin, Sequence IV is a heterolithic, siliciclastic-dominated succession comprising the Williamson-Willowvale and Dawes-upper Rose Hill Formations (Figs. 4, 5, 11). This sequence contains the deepest water deposits in the Silurian succession of the Appalachian Basin. In the vicinity of the Algonquin-Cincinnati Arch, Sequence IV is dominantly dolomitic carbonates.

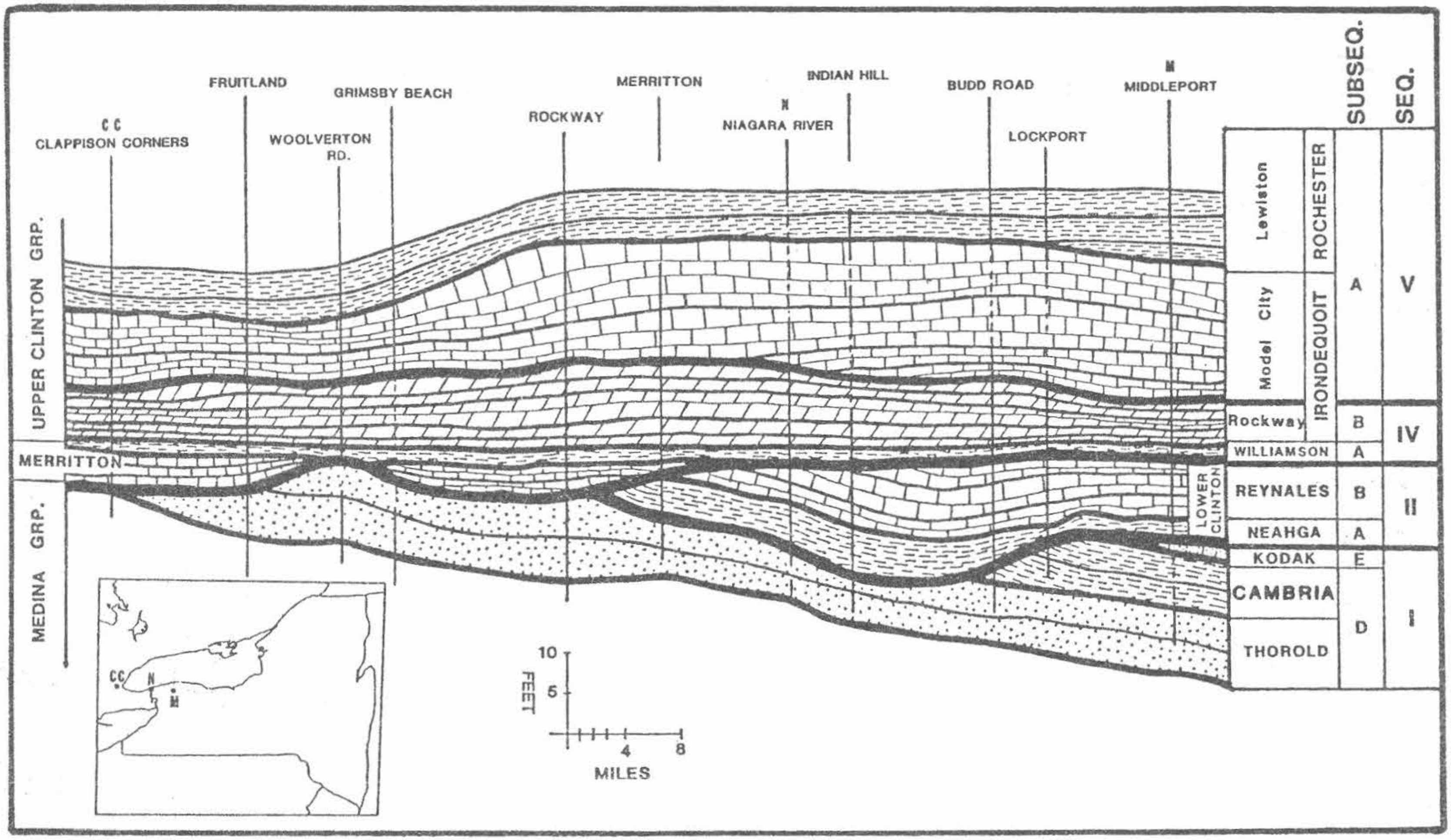


Figure 11. Regional cross section of upper Medina (sequence I), lower Clinton (sequence II) and upper Clinton (sequences IV and V) through western New York and Ontario outcrops (see inset map). Note convergence of two basal sequence boundaries (basal II and basal sequence IV unconformities) near the town of Merritton due to overstep of sequence II, and the insertion of a new unit (Merritton Limestone) apparently in the position of the Williamson Shale between the Niagara River and the Merritton outcrops. Also note beveling of upper Medina (sequence I) units beneath basal Clinton (sequence II) erosion surface east of Rockway and by combined sequence II - IV (and possibly III) erosion surfaces to the west. Modified from Kilgour (1963).

Sequence IV overlies the regionally angular, type 2 unconformity that truncates nearly the entirety of Sequences II and III between Syracuse, N.Y. and St. Catharines, Ontario. Details of this regional angular unconformity are described by Lin and Brett (1988). Subcropping units include (from east to west) the Sauquoit, Wolcott Furnace, Wolcott, Upper Sodus (Wayne County), Lower Sodus (near Rochester, N.Y.), Wallington, Brewer Dock (Niagara Gorge) and finally, the Neahga Shale (near St. Catharines, Ontario). The strike of the regional angular unconformity is approximately N30E (Fig. 12).

Between Grimsby and Owen Sound, Ontario, Sequence II is missing entirely. In this area, Sequence IV carbonates (Merritton-Fossil Hill) immediately overlie the Medina Group. In essence, the bounding unconformities of Sequence II, III and IV merge into a single unconformity atop the upper surface of the Thorold Sandstone and, north of Hamilton, Ontario, on the upper Cabot Head Formation (Grimsby equivalent) (Sequence I).

In the Bruce Penninsula, north of Owen Sound, Ontario, the two unconformities again splay apart as a wedge of Sequence II (Dyer Bay-Wingfield-St. Edmund Formations) strata is interposed. Hence, a southward or southeastward beveling of Sequence II units is evident in the Bruce Penninsula, approaching the Algonquin Arch. This is the crude "mirror image" counterpart of the north to northwest truncation surface observed at the Sequence II-IV boundary unconformity in western New York and Ontario. Hence, the geometry of the unconformity delineates the position of the Algonquin Arch which separated the subsiding Michigan and Appalachian basin.

In the New York outcrop belt, westward from the type Clinton, New York area, a 1-10 cm bed rich in phosphatic and quartz pebbles and limestone clasts, named the Second Creek Phosphate Bed (Lin and Brett, 1988) overlies the unconformity. This bed yields conodonts indicative of the amorphognathoides Zone and is traceable for some 240 km from near Syracuse at least to Niagara Gorge. In west-central New York this bed forms the base of the black to green Williamson Shale. About 10 km west of Niagara Gorge, the Second Creek bed overlies a distinct unit that appears immediately above the bounding unconformity. This unit is a thin (50 cm), glauconite-rich, dolomitic limestone termed the Merritton Limestone (Kilgour, 1963; Fig. 13). Although this bed has yielded few diagnostic fossils, the brachiopod Pentameroides, rare conodonts and Palaeocyclus corals suggest a late Llandovery C-5 to C-6 age (Kilgour, 1963; Berry and Boncot, 1970; M. Kleffner, pers. comm., 1990). The Merritton, (commonly, but erroneously, termed "Reynales Limestone" in Ontario) passes laterally into somewhat thicker Fossil Hill carbonates near Orangeville, Ontario. The Fossil Hill Dolostone of definite C-6 age, overlies the major unconformity that truncates the St. Edmund-Wingfield-Dyer Bay succession (probable Sequence II) on the Bruce Penninsula. A lithologically similar, also glauconite-rich limestone, the Dayton Limestone, overlies the unconformity in south-central Ohio.

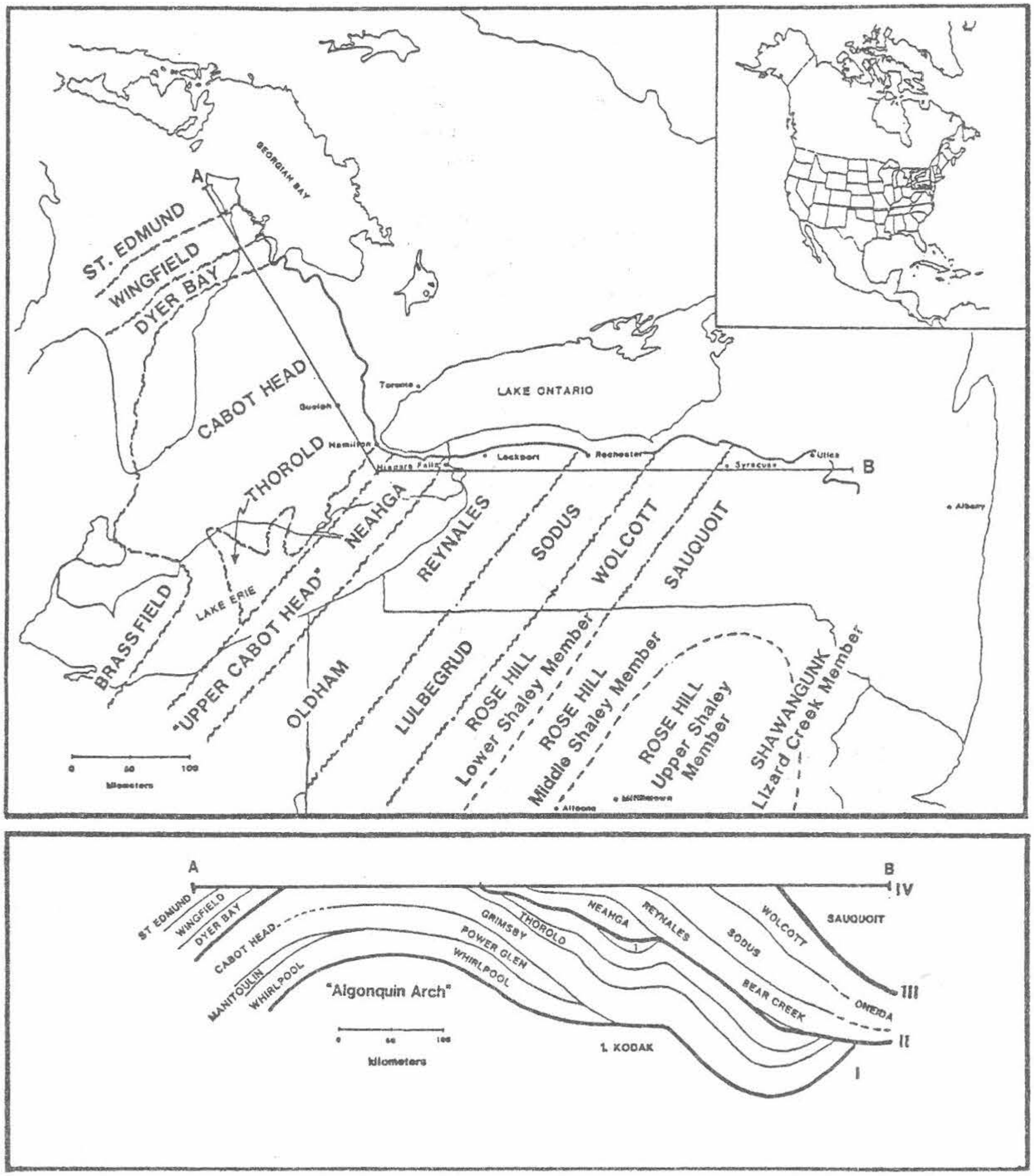


Figure 12. A) Subcrop map of beveled strata beneath Late Llandoverian (C6), Sequence IV unconformity. Line A-B shows location of cross sections shown in Fig. 12B. Subsurface data for Ontario from Winder and Sanford (1973); for New York and Ohio trends determined from data of Van Tyne (1966).

B) Schematic cross-sections along line A-B, indicated on map above. Note position of sequence I, II, and III basal bounding unconformities (labeled I, II, III): the II basal boundary displays beveling of underlying upper Medina units (Kodak, Thorold). Both sequence II and III boundaries are cross cut by the sequence IV basal erosion surface, shown as the horizontal upper line in cross section. Note regionally angular truncation surfaces on either side of region of maximum truncation - labeled "Algonquin Arch". Relative thicknesses of units on cross section are only approximate.

Where it is most typically developed in western New York and Ontario, the Sequence IV unconformity is the nearly planar but sharp contact between the Rockway Dolostone and the upper (Model City) Members of the Irondequoit Formation.

Overall, Sequence IV constitutes a large-scale (third order) symmetrical shallowing-deepening cycle. However, detailed investigation reveals that Sequence IV contains two very widespread (basin-wide) cycles (subsequences or 4th order cycles). The boundaries of these cycles correspond approximately to traditional formation and member contacts.

The first major cycle in sequence IV comprises the Williamson and eastern equivalent Willowvale shales; as noted above, this package commenses with the basal lag, condensed beds (Second Creek - Westmoreland) and passes upward into dark gray to black or greenish gray shale (Lin and Brett, 1988; Eckert and Brett, 1989). The entire Williamson-Willowvale interval remains rather uniformly thin over most of New York State but it pinches to a feather edge of green shale in Niagara County and adjacent Ontario. These muds accumulated during an early highstand when sea-level was near its highest point for the Silurian Appalachian Basin.

The overlying second subsequence is represented by the lower Irondequoit Formation (Rockway Member). The base of the Rockway is also marked by a widespread quartz granule-phosphate pebble bearing dolomitic wackestone, the Salmon Creek Bed (Lin and Brett, 1988) that appears to correlate westward with a diffuse zone of phosphatic nodules in the basal Rockway Dolostone of Ontario. This coarse-grained, condensed bed is, in many ways, analogous to the Second Creek horizon at the base of the Williamson Shale. In Ontario and western New York, the lower Rockway, is buff argillaceous dolostone with thin green shale; it contains abundant specimens of the large brachiopod <u>Costistricklandia gaspensis</u> and it thus belongs of an offshore (benthic assemblage 4) biofacies.

Sequence V: Upper Clinton Group

Sequence V comprises the bulk of the upper Clinton Group in the Niagara region (upper Irondequoit, Rochester Shale, DeCew Dolostone). This package is bounded at the base by the sharp lower contact of the upper Irondequoit Limestone (Model City Member). Although relatively little erosion occurred at this surface, the contact is sharp and displays an abrupt change from the relatively deep water shales and dolostones to crinoidal grainstone or sandstones. Furthermore, this is the local signature of a widespread low stand in sea level. Hence, we designate this surface as a sequence boundary.

The upper Clinton subsequences have recurrence intervals on the order of 1.0 m.y. and display a variety of internal motifs. There are three subsequences: 1) upper Irondequoit to Lewiston Mb.

Sat. C29

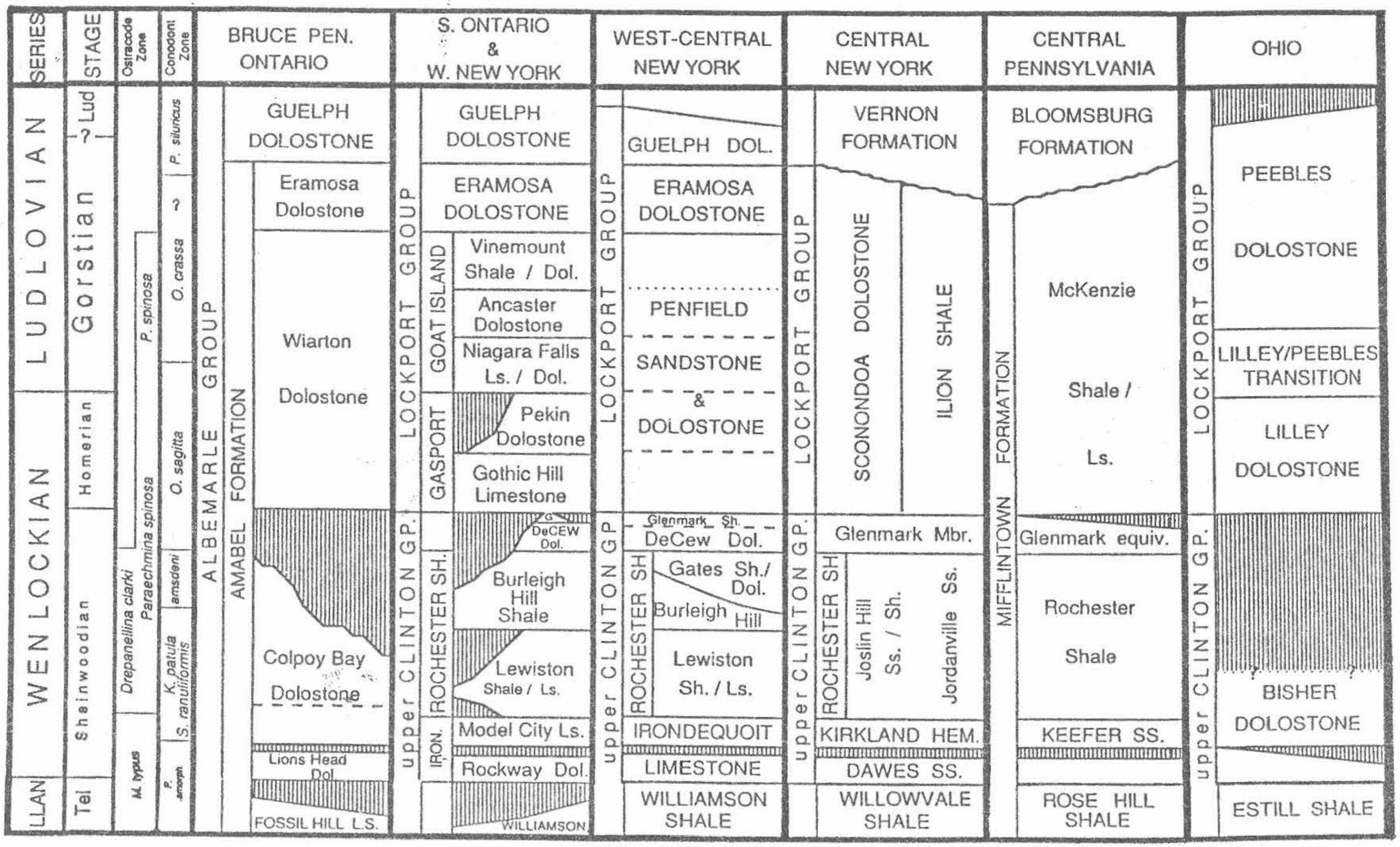


Figure 13. Chronostratigraphic correlation chart for units within the upper Clinton (sequences IV, V) and Lockport (sequences VI) groups, in Ontario, New York, Pennsylvania and Ohio. Formation names are listed in upper case, members in lower case.

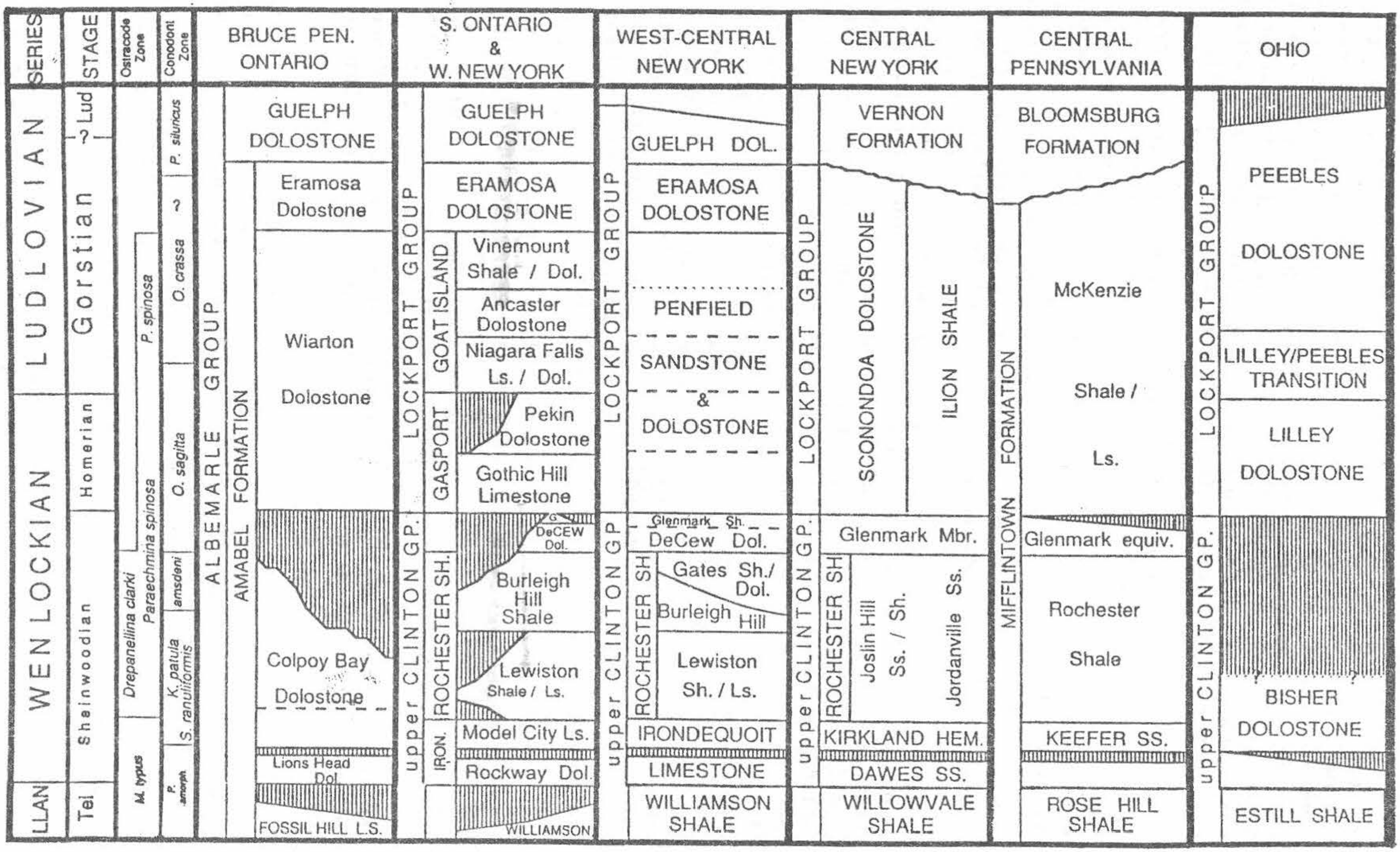


Figure 13. Chronostratigraphic correlation chart for units within the upper Clinton (sequences IV, V) and Lockport (sequences VI) groups, in Ontario, New York, Pennsylvania and Ohio. Formation names are listed in upper case, members in lower case.

of Rochester Shale; 2) upper Rochester Shale; 3) DeCew Dolostone - Glenmark Shale. In western New York and Ontario, each subsequence commenses rather abruptly with thin bryozoan and crinoid-rich pack and grainstones or calcisiltites that sharply overlie discontinuities on underlying mudstones. The crinoidal carbonates then pass upward into thin bedded to nodular wacke- and packstone carbonates and calcareous dark gray shales or mudstones (Fig. 15).

The upper Irondequoit displays facies that are everywhere shallower than those of the underlying Rockway beds and appears to mark a fairly major, albeit short-lived, sea-level lowstand of early Wenlockian age. The contact with the Rockway is sharp and is overlain by coarse crinoidal pack- to grainstone facies. Rip-up clasts, derived from the underlying fine-grained Rockway carbonates, are present locally within the basal 10-20 cm in western New York and Ontario. This bed is overlain by medium-bedded to massive, rarely cross-laminated crinoidal pack and grainstones. Near the basin center, at Rochester, New York, these beds grade upward into shaley brachiopod packstones and grainstones near the top of the Irondequoit which locally display small micritic bryozoan-algal bioherms. These structures, many of which are based on the top of a single coarse crinoidal grainstone horizon, may extend all the way through the upper Irondequoit and a short distance into the Rochester Shale. In Niagara County exposures, the entire Irondequoit is an amalgamated grainstone; this lower set of bioherms is absent, but comparable bioherms occur at or near the top of the Irondequoit in its transition to the Rochester Shale (Fig. 15). Evidently, upward growth of bioherms occurred in response to minor deepening events, a pattern also manifested in the higher Lockport section (= keep up carbonate pattern of Sarg, 1988; see below).

To the east, in the type Clinton area of New York, a similarly abrupt surface separates coarse grainstones of the Kirkland Hematite from the underlying hummocky cross-stratified calcareous sandstones of the Dawes Formation. Similarly, medium to coarse, hematitic, crinoidal, quartz arenites of the upper Keefer rest sharply on sandy shales of the Dawes equivalent (lower Keefer - upper Rose Hill) in Pennsylvania (Fig. 14). Hence, an episode of submarine non-deposition and minor erosion occurs circumbasinally in early Wenlockian time.

Crinoidal gravels and sands accumulated in shallow agitated skeletal shoals on the northwest rim of the basin, while clean quartz sands, possibly as a series of offshore sand ridges, accumulated on the southeastern margin. We suspect that these cannibalized shoreface sands were reworked during the early phases of sea level rise and the sharp surface beneath them may record a ravinement that modified a pre-existing surface of non-deposition and erosion.

In central New York State, the upward change from Irondequoit to Rochester Shale is gradational, but marked by a transitional, condensed, greenish, silty interval less than 1 m

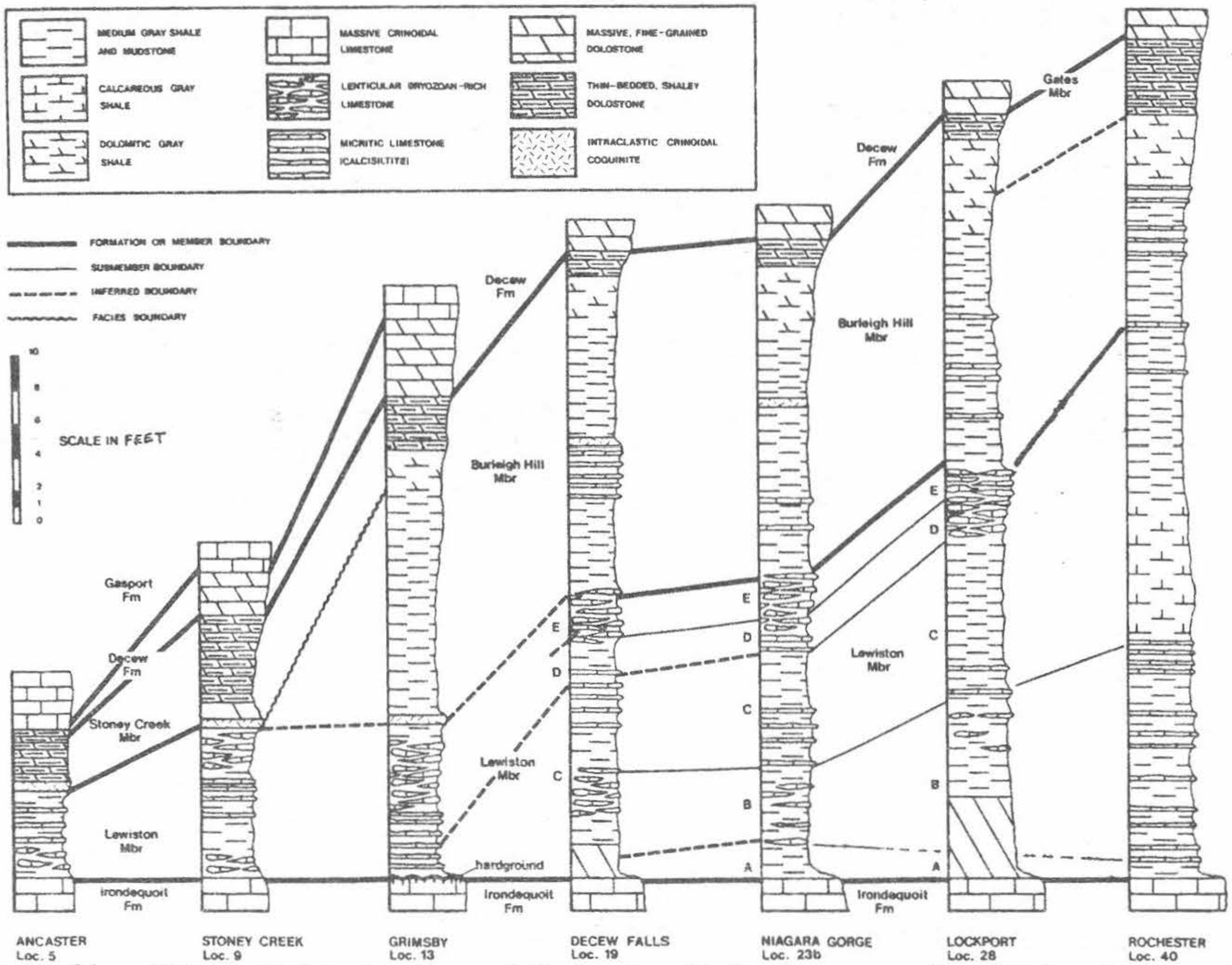


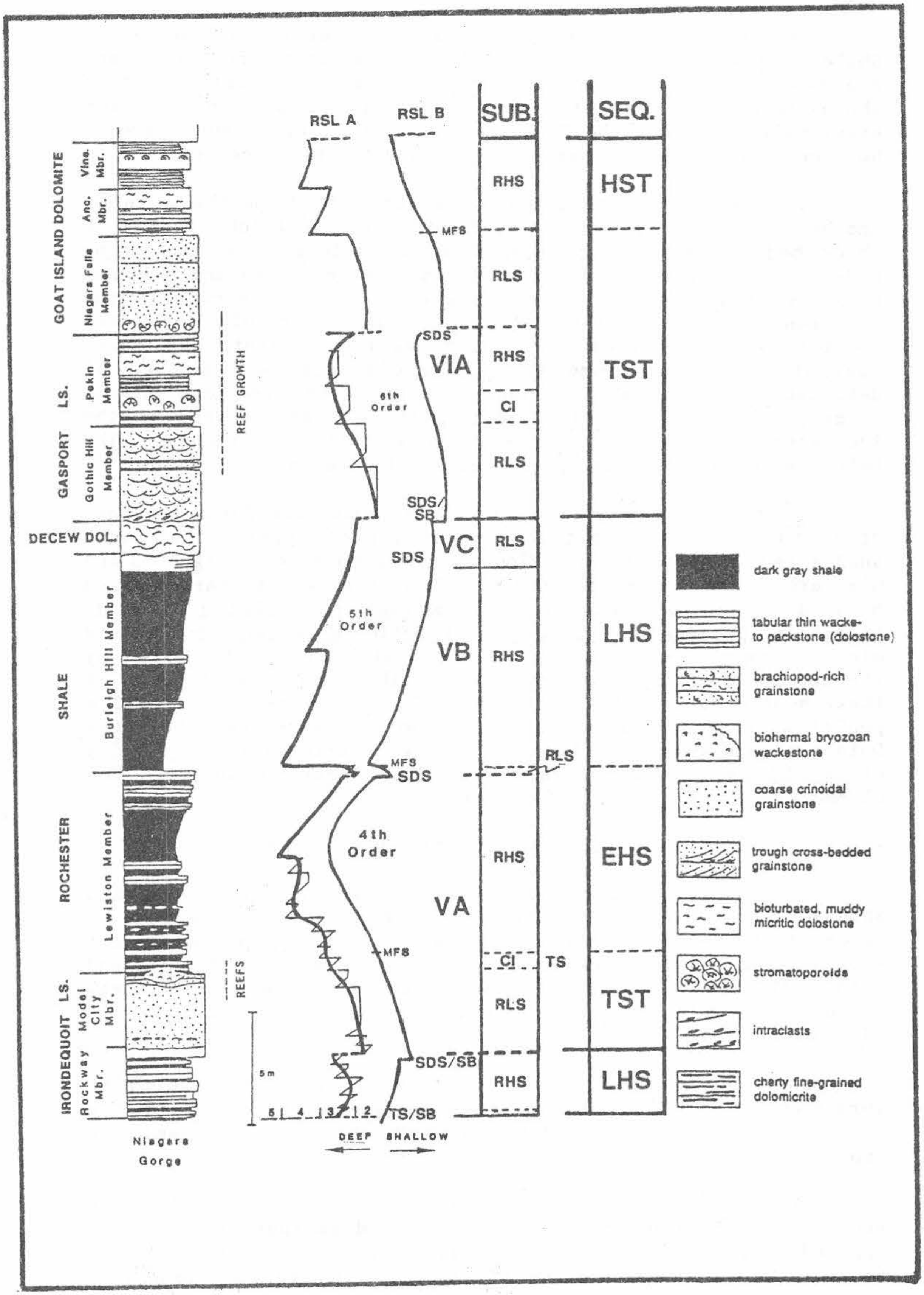
Figure 14. Stratigraphic section of Rochester Shale for seven localities in Ontario and western New York. Locations of numbered sections are shown in Figure 1. See Brett (1983) for descriptions of numbered localities.

thick. Bioherms from the Irondequoit commonly extend upward into this bed. Westward into Ontario, the top surface of the Irondequoit becomes a sharp and locally phosphate-impregnated hardground beneath the Rochester Shale. We infer that this sharp change upward to deeper water mudstones - siltstones of the Rochester Formation and the associated evidence for non-deposition and/or condensation records a marine flooding surface and probably also surface of maximum starvation.

Internally, the Rochester Shale is divisible into at least three cycles--corresponding to lower and upper submembers of the Lewiston Member and upper Burleigh Hill- Gates Members (Figs. 14, 15). The Lewiston Member displays two subsymmetrical cycles of upward decrease and increase in frequency of tempestites and in faunal diversity within the Lewiston Member; the central interval of nearly barren dark gray shale (Lewiston C) was inferred by Brett (1983) to represent deepest water conditions. The bundle of calcarenite beds at the top of the Lewiston Member (=E submember) represents a shallowing event. The sharply overlying Burleigh Hill -Gates Member records an abrupt return to dark mudstones comparable to Lewiston C and thus implies a second deepening event (Fig. 15).

The medial Rochester (Lewiston E) carbonates can be traced in the subsurface into eastern Ohio and western Pennsylvania. They occur, but are rather obscure, in the central Pennsylvania fold belt. In east central New York State, the top Lewiston carbonates appear to correlate with a westward extending tongue of coarse Herkimer Sandstone. Locally, this tongue displays a coarse phosphatic or hematitic lag on its upper surface. The sharpness of its basal contact with the lower Rochester in proximal sections and the hematitic character suggest that the Lewiston E marks the boundary of the upper Rochester subsequence.

Figure 15. Lithostratigraphy, inferred relative sea level, and sequence stratigraphic interpretation for upper Clinton and basal Lockport Groups (sequences IV, V, and basal VI) at Niagara River Gorge near Lewiston, New York. Note bases of three sequences labeled IV, V and VI, respectively; also note two zones of reef growth corresponding to early highstand (condensed) phases. Relative sea level curve calibrated to litho- and biofacies, as follows: 1 = pelmatozoan grainstone (inner BA-3); 3 = fossiliferous pack- and wackestone and calcareous (dolomitic) mudstone with diverse corals, bryozoans, pelmatozoans and brachiopods (outer BA-3 and BA-4); 4 = calcareous to dolomitic mudstone with diverse brachiopod (Striispirifer or Costistricklandia), bryozoan, pelmatozoan, and trilobite fauna (BA-4); 5 = nearly barren, dark mudstone or shale with low diversity brachiopod, trilobite or graptolite fauna (BA-5). For abbreviations, see Figure 7.



GOAT ISLAND DOLOMITE
Vine. Mbr.
Anc. Mbr.
Niagara Falls Member
GASPORT LS.
Pekin Member
Gothic Hill Member
DECEW DOL.
SHALE
Burleigh Hill Member
ROCHESTER
Lewiston Member
IRONDEQUOIT LS.
Model City Mbr.
Rockway Mbr.
Niagara Gorge
REEF GROWTH
REEFS
5m
5 4 3 2
DEEP
SHALLOW
RSL A
RSL B
MFS
SDS
6th Order
SDS/SB
SDS
5th Order
MFS
SDS
4th Order
MFS
SDS/SB
TS/SB
VIA
VC
VB
VA
SUB.
RHS
RLS
RHS
CI
RLS
RLS
RHS
RLS
RHS
CI
TS
RLS
RHS
SEQ.
HST
TST
LHS
EHS
TST
LHS
dark gray shale
tabular thin wacke- to packstone (dolostone)
brachiopod-rich grainstone
biohermal bryozoan wackestone
coarse crinoidal grainstone
trough cross-bedded grainstone
bioturbated, muddy micritic dolostone
stromatoporoids
intraclasts
cherty fine-grained dolomicrite

The overlying Burleigh Hill-Gates Member of the Rochester Shale displays an upward increase in tempestite frequency and coarse detrital carbonate (calcisiltite) indicative of a shallowing or progradational trend. The DeCew sharply and erosionally overlies the Rochester Shale in shallow shelf areas, but approaches relative conformity toward the basin center.

The DeCew Dolostone, which appears to form the base of another deepening upward cycle, is a 2 to 3 m thick, medium to thick bedded dolomititic pelsparite and calcisiltite containing beds of intraformational (storm rip-up) conglomerates and abundant hummocky cross-stratification. Its most notable feature is the occurrence of strong internal deformation, particularly in the lower beds. Overturned isoclinal folds (enterolithic structure) suggest submarine slumping. This distinctive 1-2m bed of deformation is traceable at least 150 km in the New York-Ontario outcrop belt and a similar calcisiltite bed at the top of the Rochester Shale in Pennsylvania displays similar deformation. The DeCew is tentatively interpreted as a seismite horizon.

In west-central New York the DeCew dolostone forms the base of a distinct 2 - 10 m-thick succession of uppermost Clinton shales and argillaceous brachiopod-rich carbonates, designated the Glenmark Member by Brett (1983b). This interval is characterized by a distinctive brachiopod assemblage (<u>Howellella</u> sp., <u>H. bicostata</u>, <u>Nucleospira pisiformis</u>, <u>Whitfieldella marylandica</u>) and minor thrombolitic horizons. Westward the unit is truncated by the sub-Lockport unconformity. A possible erosional remnant of these beds occurs at St. Catharines, Ontario where a local 2-3 m pod of shale and argillaceous limestone occurs between the DeCew Dolostone and overlying Lockport. The occurrence of <u>Howellella</u> cf. <u>H. bicostata</u> with corals suggests a correlation with the Glenmark Shales of central New York.

Sequence VI: Lockport Group-Vernon Shale

The sixth Silurian depositional sequence is an upward shallowing carbonate dominated succession, represented by heterolithic McKenzie Shales and limestone facies in the central Appalachians and the Lockport Dolostones in New York, Ontario and Ohio. The upper portion of the sequence comprises Bloomsburg-Vernon red and green siliciclastic mudstones (see Figs., 4, 16, 17).

Sequence VI spans the upper Wenlockian to Ludlovian interval. The exact position of the series boundary is the subject of ongoing study by LoDuca and Brett (in press); it apparently occurs low in the Lockport Group.

The lower boundary of sequence VI is demarcated by an erosion surface separating the Clinton and Lockport Groups, and located below the base of the Gasport Limestone in western New York. This flat, to gently undulatory, unconformity truncates part or all of the Glenmark, DeCew and Rochester Formations of the underlying Clinton Group (Sequence V). A correlative, but cryptic, discontinuity probably occurs within the Amabel Formation north of Hamilton, Ontario.

The top of Sequence VI is tentatively placed at a major truncation surface above the Vernon Shale in eastern New York State. This angular unconformity bevels Vernon-Ilion Shales (Lockport equivalent) and eventually all of Sequences V, IV and III in eastern New York. This is believed to represent an episode of uplift and erosion along the eastern basin margin associated with the Salinic Disturbance, as discussed under Geologic Setting.

The basal sequence VI erosion surface is overlain by a conglomeratic layer containing carbonate and mudstone clasts, frequently darkened by phosphatic or glauconitic impregnation, eroded from underlying beds (Fig. 16). These clasts occur at the base of thick-bedded to massive crinoidal grainstone belonging to the Gothic Hill Member of the Gasport Formation. The latter displays prominent trough and bidirectional cross-bedding indicating deposition on a tidally influenced shallow shelf. The unit locally varies in thickness form over 8m (25') to near zero within distances of 2-5 km and may have accumulated in a series of crinoidal ridges with intervening swales of minimal deposition. Near Rochester, New York, this unit is replaced by cross stratified, crinoid-bearing quartz arenites (Penfield dolomitic sandstone). Small tabulate coral-stromatoporoid bioherms occur within the carbonate grainstones in Niagara County. The lower Gasport grainstones grade upward into dark argillaceous dolostones (Pekin Member) with a sparse fauna. Locally bioherms up to 5 m high extend upward from the top of the grainstones into this upper argillaceous member. Hence, again, a deepening event is associated with upward growth of bioherms.

A second unit of pack- and grainstone, with a small stromatoporoid tabulate coral bioherm unit, occurs above the Gasport in the basal Niagara Falls Member of the Goat Island Formation. Again, there is a sharp contact on underlying Pekinargillaceous beds. The grainstones, like those of the Gothic Hill Member, are highly variable in thickness and pass upward into dark gray, cherty thin-bedded to argillaceous dolostones (Ancaster-Vinemount Members). At least two comparable grainstone-packstone based fining upward cycles occur in the succeeding Eramosa Formation. They have not yet been studied in detail.

The overlying Guelph displays cycles that include stromatolitic beds alternating with flaggy micritic beds. Together, these cycles define an upward shallowing sequence in the Lockport, although the component 4th order cycles are typically upward deepening. Each 4th order cycle appears to be marked at its base by a marine flooding surface. As in Clinton cycles, shallowing portions of cycles are poorly represented and are reflected primarily in minor hiatus surfaces below coarse, transgressive grainstone-packstone shoal deposits. Internally, the formation scale deepening upward cycles are composed of minor PAC-scale cycles that show the classic shallowing upward motifs.

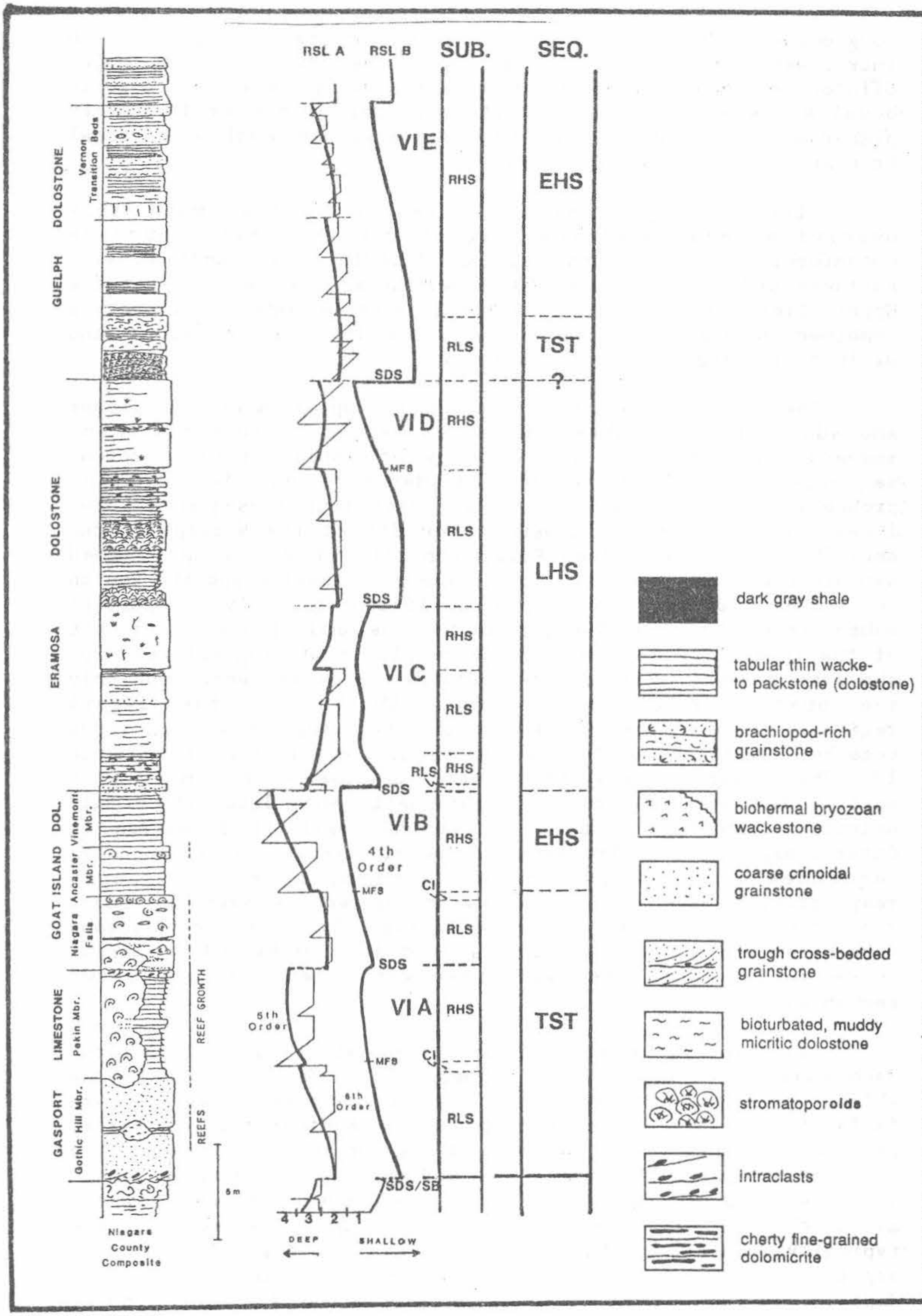
The upper Lockport grades upward into a thick (30 to 100 m) package of greenish gray to black shale, dolostone and red mudstone, the Vernon-Bloomsburg Formation. This package represents a major progradational wedge of marginal marine to non-marine muds.

SUMMARY and CONCLUSIONS

Recent study of outcrop and drill core sections in the Niagara Frontier region indicates that the classic Niagaran Series, ranging in age from Early Llandoverian to Late Ludlovian, is divisible into at least five major unconformity bounded sequences, designated sequences I, II, IV, V and VI (see Fig. 4). The sequences recognized herein are comparable in duration to those recognized by seismic stratigraphers in various portions of the Mesozoic and Cenozoic column. Each of the Silurian sequences is bounded at its base by a major erosional unconformity; two of these unconformities appear to be regionally angular surfaces which bevel strata of underlying sequences. Sequence III, the Middle Clinton Group of the type-Clinton area in central New York State, is completely missing due to erosional unconformities in western New York. The most pronounced of these major erosional surfaces is the basal Cherokee Unconformity which also coincides with a megasequence boundary of the Tutelo Holostome recognized by Dennison and Head (1975; see Dennison, 1989). The surface bevels older Upper Ordovician siliciclastic rocks of the Queenston clastic wedge in an eastwardly direction. The other sequence boundaries are less pronounced, but have been recognized at least as local unconformities by some workers. These are, in ascending order: the basal Clinton unconformity marked by the Densmore Creek Phosphate Bed; the upper Clinton angular unconformity marked by the Second Creek phosphate bed; the sub-Irondequoit unconformity marked by a zone of hardground and intraformational

---

Figure 16. Lithostratigraphy, inferred relative sea level, and sequence stratigraphic interpretation of Lockport Group (sequence VI), Niagara County, New York. Note intervals of upward reef (bioherm) growth corresponding to early highstand (condensed section) phases. Calibration of relative sea level curves based on litho- and biofacies, as follows: 2 = crinoidal grainstone with abundant stromatoporoids and tabulate corals (inner BA-3 to BA-2); 3 = argillaceous, reefy dolostones with diverse small rugose corals, tabulates, bryozoans and brachiopods (outer BA-3 to BA-4); 4 = dolomitic shales with scattered small rugose corals, brachiopods, including astrypids, rhynchonellids, and Leptaena, and trilobites (inner BA-4). For explanation of abbrieviations, see Figure 7.



RSL A
RSL B
SUB.
SEQ.
GUELPH DOLOSTONE
Vernon Transition Beds
ERAMOSA DOLOSTONE
GOAT ISLAND DOL.
Vinemont Mbr.
Ancaster Mbr.
Niagara Falls
GASPORT LIMESTONE
Pekin Mbr.
Gothic Hill Mbr.
REEF GROWTH
REEFS
5m
Niagara County Composite
VI E
VI D
VI C
VI B
VI A
4th Order
5th Order
6th Order
MFS
SDS
SDS/SB
CI
RHS
RLS
EHS
TST
?
LHS
4 3 2 1
DEEP
SHALLOW
dark gray shale
tabular thin wacke- to packstone (dolostone)
brachiopod-rich grainstone
biohermal bryozoan wackestone
coarse crinoidal grainstone
trough cross-bedded grainstone
bioturbated, muddy micritic dolostone
stromatoporoids
intraclasts
cherty fine-grained dolomicrite

conglomerate; the basal Lockport unconformity marked, again, by an intraclastic conglomerate. Of these, the lower Clinton, upper Clinton and basal Lockport are demonstrably regional, angular bevel surfaces. The sub-Irondequoit disconformity is nearly planar and shows only minor evidence for erosion without regional truncation of underlying strata.

Distinctive phosphatic, intraclastic beds immediately overlying unconformities (except for the basal Cherokee unconformity) mark both sequence bounding (lowstand) erosion surfaces and marine transgressive surfaces. In the case of the Upper Clinton unconformity, the basal sequence boundary reflects a combined lowstand erosion surface, initial transgression, and maximum flooding/ sediment starvation.

The Silurian sequences correspond approximately to groups and subgroups and subsequences correspond to formations and members. Systems tracts are variably developed within different sequences recognized herein. The Medina Group, for example, probably is subdividisible into at least four subsequences, the lowest corresponding to lowstand deposits of the Whirlpool, the second to the Power Glen Formation, the third to an unnamed sandstone and the overlying lower Grimsby Formation and the fourth to the Thorold Sandstone. Sequence II contains only one partial subsequence in Niagara County, due to erosional truncation of most of the Lower Clinton Group; Sequence III is missing entirely for the same reason. Sequence IV contains two subsequences but only the upper one is of any substantial thickness in the Niagara region. The Williamson Shale and the overlying Rockway Dolostone together constitute early and late highstand portions of sequence IV. The transgressive systems tracts of sequence IV is missing in most of the area, although the phosphatic and glauconitic, thin dolomitic limestone referred to as the Merritton Formation in Ontario may be a thin transgressive basal interval of sequence IV. Sequences V and VI (uppermost Clinton, Lockport Groups, respectively) have well developed transgressive systems tracts consisting of crinoidal grainstone facies. Overlying argillaceous carbonates and shales record early and late highstands. Hence, three subsequences are recognized within each of these two sequences.

Silurian subsequences display a consistent internal pattern. Each subsequence begins with a sharp to slightly erosional base which typically juxtaposes shallow water, coarse-grained winnowed facies (quartz arenites in the case of the Medina Group; crinoidal pack- and grainstones in the case of the uppermost Clinton and the Lockport Groups) onto deeper water mudstones. In sequences II and IV, these basal shallow water facies are absent. In subsequences, as in the larger sequences, the lower coarse-grained beds are typically abruptly set off from an upper finer grained, typically argillaceous portion, herein referred to as a relative highstand deposit, by a surface of sediment starvation (marine flooding surface) which is often marked by minor amounts of phosphate,

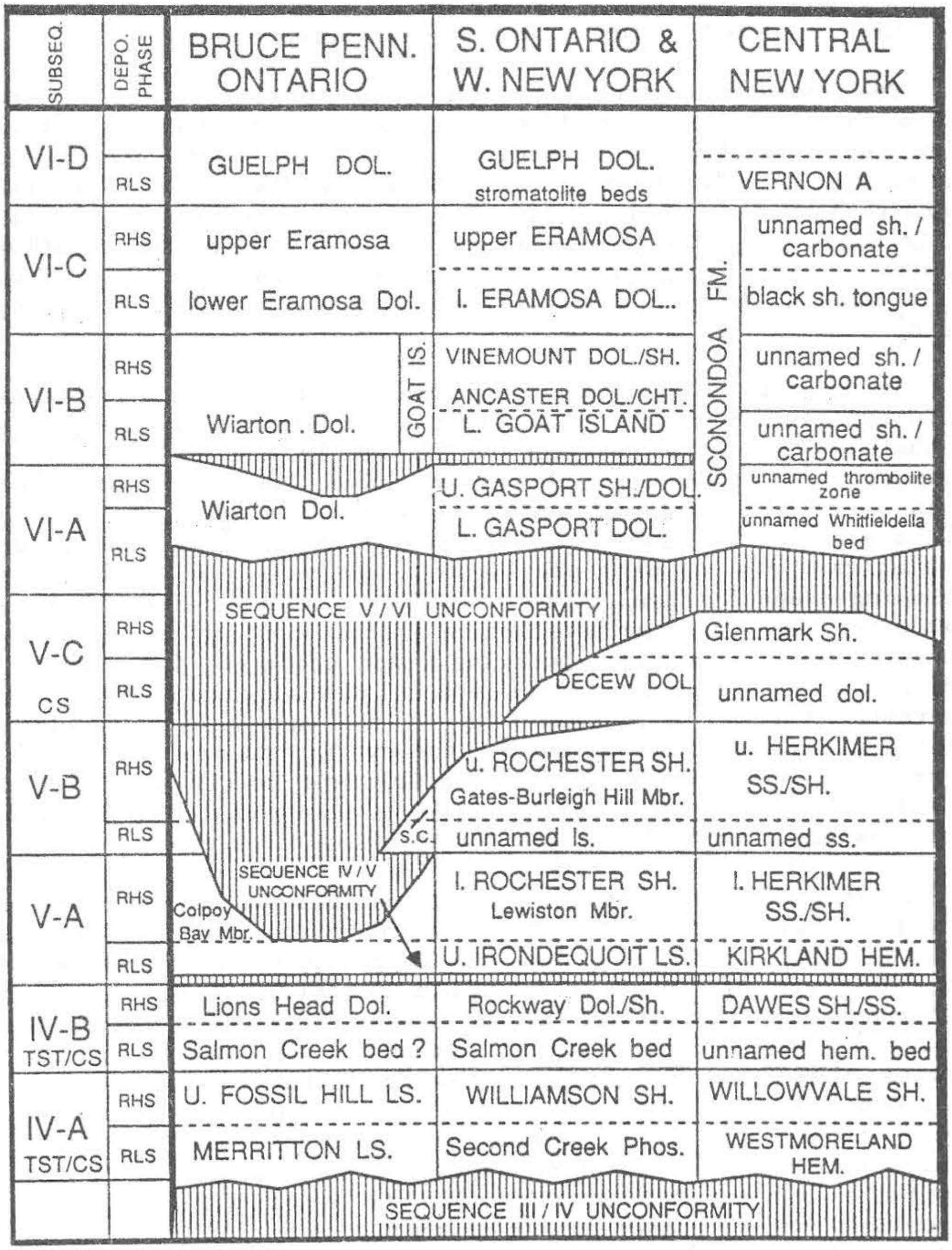


Figure 17. A) Summary chart showing subsequence division of correlative upper Clinton and Lockport Groups (sequence IV and V) in the Bruce Peninsula, southern Ontario to western New York, and central New York. Note major sequence boundaries. Abbreviations as in Figure 4. Units not scaled to thickness, but estimated relative time. In subsequence VB, SC = Stoney Creek (upper) Member of the Rochester Shale.

glauconite or hematite. Within each sequence, as a whole, one of the marine flooding surfaces of a smaller scale subsequence also corresponds to the surfaces of maximum starvation or downlap surface. These sharp internal surfaces within the subsequences are the result of maximally increased rates of relative sea level rise which produced maximal sediment starvation.

The overlying highstand deposits consist typically of shales, hummocky cross-bedded siltstones, sandstones, and/or carbonates. In the case of Medina cycles, the upper part of the subsequence or relative highstand phase may display a general upward shallowing or progradational trend. In the case of Clinton and Lockport sequences, the upward pattern within the relative highstand of the subsequences is less clear and may range from nearly static to even slightly retrogradational or deepening upward.

It is, of course, one thing to recognize cyclic patterns within sedimentary rocks, and another to understand the processes responsible for these patterns. The widely traceable Silurian sequences, subsequences, and parasequences certainly were not formed by local basin subsidence, because the cycles, even at a small scale, can be traced across depositional strike from areas of relatively low to high subsidence. In order to explain the shallowing-deepening cycles, basin tectonic models would have to account for their extreme persistence in the Appalachian region as a whole. Hence, the minor buckling which produced, for example, condensation and erosion of beds near the Algonquin Arch or, conversely, thickening of sections in the Appalachian foreland itself are too localized to explain the cyclic patterns. In fact, any tectonic model would have to incorporate nearly synchronous downward flexure and in some cases, rapid up-buckling of the entire eastern edge of North America to account for these cycles.

Conversely, at present it is impossible to substantiate a eustatic origin for cycles. Also, mechanisms for small scale cyclicity based upon old global sea level change are obscure at best. Certain of the larger cyclic patterns, sequences and subsequences in particular, appear to be traceable beyond the Appalachian Basin. For example, Johnson et al. (1985) provided evidence that Llandoverian shallowing-deepening cycles that correspond to our sequences and subsequences could be traced very broadly, if not globally. Johnson et al. (1985) provide evidence for worldwide relative lowstands during Llandoverian A-1 to B-2, C-3 to C-4 time, and in late C-5 to earliest C-6 time. Particularly strong pulses of sea level rise occurred about in the Llandovery A-2 to A-3 and in the C-1 to 2, and C-6. All of these signatures can be found in the strata of the Niagara region.

Recent work suggests that relative lows and highs of sea level in the Wenlockian and Ludlovian interval may be equally widely correlative. Early Wenlockian lowstands, corresponding to the Irondequoit limestone, and two closely spaced lowstands in the latest Wenlockian corresponding to the Gasport and lower Goat

Island Formations appear to be correlative into the British Silurian section. There, the Buildwas Beds of the Welch borderland and the Much Wenlock limestone correspond to the late Wenlockian sea level drop intervals. The highstands corresponding to the Rochester Shale and the upper Goat Island formations respectively, occur in the British successions as the Coalbrookdale beds (Wenlock Shale of older literature) and the basal Ludlow Elton beds (see Siveter et al., 1990a,b). Late-early to middle Ludlovian sea level lowering corresponding to the Eramosa and Guelph interval in New York and Ontario may occur in high Ludlow section (Whitcliff and Leintwardian beds) in the type Ludlow section of the Welch borderland.

Further work is required to corroborate these correlations and extend them into other geographic areas. There is certainly a strong suggestion that the major sequence and subsequence of sea level highs and lows may be eustatically produced. Since these cycles represent from 1 to 3 million year sea level fluctuation patterns, it may not be entirely unreasonable to propose a plate tectonic explanation for them. On the other hand, the smaller scale megacyclothems, cyclothems, and PACs are not so readily explained by plate tectonic eustatic processes. Either relatively poorly understood glacioeustatic processes or possibly geoidal fluctuations might account for these small scale oscillations in relative sea level.

ACKNOWLEDGEMENTS

We wish to thank several people who have contributed to this project with new field data or discussions including William Duke, James Eckert, Mark Kleffner, Dorothy Tepper and Denis Tetreault. Wendy Taylor provided invaluable assistance in the drafting of figures and Heidi Jacob patiently worked through the last several drafts. Ronald Cole, Curt Teichert and Tom Grasso provided critical reviews that aided in improving the manuscript. This research was supported by a grant from the donors to the Petroleum Research Fund (American Chemical Society), grants from the New York State Museum of Sigma Xi, and the Envirogas Corporation.

REFERENCES

Baird, G.C. and Brett, C.E., 1986, Erosion on an anaerobic sea floor: significance of reworked pyrite deposits from the Devonian of New York State. Palaeogeog., Palaeoclim., Palaeoecol., v. 57, p. 157-193.

Baird, G.C., Brett, C.E., and Kirchgasser, W.T., 1988, Genesis of black-shale roofed discontinuities in the Devonian Genesee Formation, western New York, p. 357-375. In McMillan, N.J., Embry, A.I., and Glass, D.J., eds., Devonian of the World, Can. Soc. Petrol. Geol. Mem. 14, v. 2.

Baum, G.R. and Vail, P.R., 1988, Sequence stratigraphic concepts applied to Paleogene outcrops, Gulf and Atlantic Basins. SEPM Spec. Pub. v. 42, p. 309-328.

Bayer, U. and Seilacher, A., eds., 1985, Sedimentary and Evolutionary Cycles. Lecture Notes in Earth Sciences 1, Springer-Verlag, Berlin, Heidelberg, New York, Tokyo.

Berry, W.B.N., and Boucot, A.J., 1970, Correlation of the North American Silurian rocks. Geol. Soc. Amer. Spec. Pap. 102, 289 p.

Bolton, T.E., 1957, Silurian stratigraphy and palaeontology of the Niagara Excarpment in Ontario. Geol. Survey of Canada Memoir 289, 145 p.

Brett, C.E., 1982, Stratigraphy and facies variation of hte Rochester Shale (Silurian: Clinton Group) along Niagara Gorge. N.Y. State Geol. Assoc. 54th Ann. Mtg. Fieldtrip Guidebook, p. 217-245.

Brett, C.E., 1983, Sedimentology, facies relations, and depositional environments of the Rochester Shale. Jour. Sed. Pet., v. 53, p. 947-972.

Brett, C.E., Goodman, W., LoDuca, S.T., Tepper, D.H., Eckert, B.Y. and Duke, W.L., in review, Revisions of Niagaran series stratigraphy for the type-section area in western New York. N.Y. State Mus. Bull.

Busch, R.G. and Rollins, H.B., 1984, Correlation of Carboniferous strata using a hierarchy of transgressive-regressive units. Geology v. 12, p. 471-474.

Cotter, E., 1982, Tuscarora Formation of Pennsylvania. SEPM (Eastern Section) Field Trip Guidebook, 105 p.

Cotter, E., 1983, Silurian depositional history. In Guidebook 48th Ann. Field Trip Conference of Pennsylvania Geologists, p. 3-27.

Cotter, E., 1988, Hierarchy of sea level cycles in the medial-Silurian siliciclastic succession of Pennsylvania. Geology v. 16, p. 242-245.

Cross, T.A., 1988, Controls on coal distribution in transgressive-regressive cycles, Upper Cretaceous, Western Interior, U.S.A. In Wilgus, C.K., et al., eds., Sea-level changes: an integrated approach. SEPM Spec. Pub. no. 42, p. 371-380.

Crowley, D.J., 1973, Middle silurian patch reefs in the Gasport Member (Lockport Formation). New York. Amer. Assoc. Petrol. Geol. Bull. v. 57, p. 283-300.

Dennison, J.M., ed., 1989, Paleozoic sea-level changes in the Appalachian Basin. 28th Intl. Geol. Congress Field Trip Guidebook T354, 56 p., Washington, D.C.

Dennison, J.M. and Head, J.W., 1975, Sea level variations interpreted from the Appalachian Basin Silurian and Devonian: Amer. Jour. Sci. v. 275, p. 1089-1120.

Donovan, A.D., et al., 1988, Sequence stratigraphic setting of the Cretaceous-Tertiary Boundary in central Alabama. In Wilgus, C.K., et al., eds., Sea-level changes: an integrated approach. SEPM Spec. Pub. No. 42, p. 299-307.

Duke, W.L., 1987, Revised internal stratigraphy of the Medina Formation in outcrop: an illustration of the inadequacy of color variation as a criterion for lithostratigraphic correlation. In Duke, W.L., ed., Sedimentology, stratigraphy and ichnology of the Lower Silurian Medina Formation in New York and Ontario, SEPM Northeastern Section Field Trip Guidebook, p. 16-30.

Duke, W.L., et al., 1987, Sedimentology, stratigraphy and ichnology of the Lower Silurian Medina Formation in New York and Ontario. In Duke, W.L., ed., SEPM. Northeastern Section Field Trip Guidebook, 183 p.

Duke, W.L. and Brusse, W.C., 1987, Cyclicity and channels in the upper members of the Medina Formation in the Niagara Gorge. In Duke, W.L., ed., Sedimentology, Stratigraphy, and Ichnology of the Lower Silurian Medina Formation in New York and Ontario. SEPM Northeast Section, Field Trip Guidebook.

Duke, W.L., and Fawcett, P.J., 1987, Depositional environments and regional sedimentary patterns in the upper members of the Medina Formation. In Duke, W.L., ed., Sedimentology, Stratigraphy, and Ichnology of the Lower Silurian Medina Formation in New York and Ontario. SEPM Northeast Section, Field Trip Guidebook, p. 81-95.

Eckert, J.D., 1988, Systematics, evolution, and biogeography of Late Ordovician and Early Silurian crinoids. Unpubl. Ph.D. diss., Univ. of Rochester, 408 p.

Eckert, B.Y. and Brett, C.E., 1989, Bathymetry and paleoecology of Silurian benthic assemblages, Late Llandoverian, New York State: Palaeogeog. Palaeoclim. Palaeoecol. v. 74, p. 297-326.

Ettensohn, F.R., 1987, Rates of relative plate motion during the Acadian orogeny based on spatial distribution of black shales. Jour. Geol. v. 95, p. 572-582.

Fischer, A.G., 1963, The Lofer cyclothems of the Alpine Triassic. Kansas Geol. Surv. Bull. 169, p. 107-149.

Fischer, A.G., Herbert, T. and Premoli Silva, I., 1985, Carbonate bedding cycles in Cretaceous pelagic and hemipelagic sequences, p. 1-10. In Pratt, L.M., Kauffman, E.G., and Selt, F., eds., Fine Grained Deposits and Biofacies of the Cretaceous western Interior Sea Way: Evidence for Cyclic Sedimentary Processes. SEPM Guidebook 4.

Galloway, W., 1989, Genetic stratigraphic sequences in basin analysis I: architecture and genesis of flooding-surface bounded depositional units. Am. Assoc. Petrol. Geol. Bull., v. 73, p. 125-142, p. 143-154.

Gillette, T., 1947, The Clinton of western and central New York. New York State Mus. Bull., v. 341, 191 p.

Goodwin, P.W. and Anderson, E.J., 1985, Punctuated Aggradational Cycles: a general hypothesis of episodic stratigraphic accumulation. Jour. Geol. v. 93, p. 515-533.

Hall, J., 1839, Third annual report of the Fourth Geological District of the State of New York. New York State Geol. Surv. Ann. Rpt., No. 3, p. 287-339.

Hall, J., 1852, Containing descriptions of hte organic remains of the lower middle division of the New York System: Paleontology of New York, v. 2, 362 p.

Heckel, P.H., 1986, Sea-level curve for Pennsylvanian eustatic marine T-R depositional cycles along the mid-continental outcrop belt, North America. Geology, v. 14, p. 330-334.

House, M.R., 1985, A new approach to an absolute timescale from measurements of orbital cycles and sedimentary microrhythms, Nature, v. 316, p. 721-725.

Johnson, M.E., 1987, Extent and bathymetry of North American platform seas in the Early Silurian, Paleoceanography, v. 2, no. 2, p. 185-211.

Johnson, M.E., Rong, J.Y., and Yang, X.C., 1985, Intercontinental correlation by sea level events in the Early Silurian of North America and China (Yangtze Platform). Geol. Soc. America Bull. v. 96, p. 1384-1397.

Kilgour, W.J., 1963, Lower Clinton (Silurian) relationships in western New York and Ontario. Geol. Soc. Amer. Bull. v. 74, p. 1127-1141.

Kleffner, M.A., 1989, A cononont-based Silurian chronostratigraphy. Geol. Soc. Amer. Bull. v. 101, p. 904-912.

Lin, B.Y. and Brett, C.E., 1988, Stratigraphy and disconformable contacts of the Williamson - Willowvale interval: revised correlations of the Late Llandoverian (Silurian) in New York State. Northeastern Geology v. 10, p. 241-253.

Lukasik, M., 1988, Lithostratigraphy of Silurian rocks in southern Ohio and adjacent Kentucky and West Virginia. Unpubl. Ph.D. Dissertation, Univ. Cincinnati, 313p.

LoDuca, S.T., 1988, Lower Clinton hematites: Implications for stratigraphic correlations. Abstracts for the Central Canadian Geological Conference, London, Ontario, 62 p.

LoDuca, S.T. and Brett, C.E., 1990, Stratigraphic relations of Lower Clinton Hematites. Geol. Soc. Amer. Abs. with programs v. 22, p. 31.

LoDuca, S.T., and Brett, C.E., in press, Placement of the Wenlockian/Ludlovian boundary in New York State: Lethaia.

Martini, I.P., 1971, Regional analysis of sedimentology of the Medina Formation (Silurian) in Ontario and New York. Amer. Assoc. Petrol. Geol. Bull., v. 55, p. 1249-1261.

Middleton, G.V., 1987, Geologic setting of the northern Appalachian Basin during the Early Silurian. In Duke, W.L., ed., Sedimentology, Stratigraphy, and Ichnology of the Lower Silurian Medina Formation. Soc. Econ. Paleont. Mineral., Northeastern Section, Fieldtrip Guidebook, p. 1-15.

Middleton, G.V., Rutka, M., and Salas, C.J., 1987, Depositional environments in the Whirlpool Sandstone Member of the Medina Formation. In Duke, W.L., ed., Sedimentology, Stratigraphy, and Ichnology of the Lower Silurian Medina Formation. SEPM Northeastern Section, Field Trip Guidebook, p. 1-15.

Miller, M. and Eames, L.E., 1982, Palynomorphs from the Silurian Medina Group (Lower Llandovery) of the Niagara Gorge, Lewiston, New York. Palynology v. 6, p. 221-254.

Nummedal, D. and Swift, D.J.P., 1987, Transgressive stratigraphy at sequence-bounding unconformities: some principles derived from Holocene and Cretaceous examples. SEPM Spec. Pub. 41, p. 241-260.

Posamentier, H.W., Jervey, M.T., and Vail, P.R., 1988a, Eustatic controls on clastic deposition - conceptual framework. In Wilgus, C.K. et al, eds., Sea-level changes: an integrated approach. SEPM 42, p. 109-124.

Posamentier, H.W. and Vail, P.R., 1988b, Eustatic controls on clastic deposition II: sequence and systems tract models. In Wilgus, C.K., et al., eds., Sea-level Changes: An Integrated Approach. SEPM Spec. Pub., no. 42, p. 125-154.

Prave, A.R., Alcala, M.L., and Epstein, J.B., 1989, Stratigraphy and sedimentology of Middle and Upper Silurian rocks and an enigmatic diamictite, Southeastern New York. In N.Y. State Geol. Assoc. Field Trip Guidebook, p. 121-140.

Quinlan, G.M. and Beaumont, C., 1984, Appalachian thrusting, lithospheric flexure, and the Paleozoic stratigraphy of the eastern interior of North America. Canadian Jour. Earth Sci. v. 21, p. 973-996.

Ramsbottom, W.H.C., 1979, Rates of transgression and regression in the Carboniferous of Northwestern Europe. Jour. Geol. Soc. London v. 136, p. 147-153.

Rickard, L.V., 1975, Corelation of the Silurian and Devonian rocks of New York State. New York State Map and Chart Series.

Ross, C.A. and Ross, J.R.P., 1985, Late Paleozoic depositional sequences are synchronous and worldwide. Geology, v. 13, p. 194-197.

Ross, C.A. and Ross, J.R., 1988, Late Paleozoic transgressive-regressive deposition. In Wilgus, C.K., et al., eds., Sea-level Changes: An Integrated Approach. SEPM Spec. Pub., no. 42, p. 226-247.

Sarg, J.F., 1988, Carbonate sequence stratigraphy. In Wilgus, C.K., et al., eds., Sea-level Changes: An Integrated Approach. SEPM Spec. Pub., no. 42, p. 155-181.

Siveter, D.J., Owens, R.M., and Thomas, A.T., 1989a, The Northern Wenlock Edge Area: Shelf muds and carbonates on the midland Platform. In Bassett, M.G., ed., Silurian Field Excursions: A Geotraverse Across Wales and the Welsh Borderlands. Geol. Ser. No. 10, Natl. Mus. of Wales, 133 p. Cardiff, Wales.

Siveter, D.J., Owens, R.M. and Thomas, A.T., 1989b. The Ludlow Anticline and Contignous areas: a shelf marine to non-marine transition. In Bassett, M.G., ed., Silurian Field Excursions: A Geotraverse Across Wales and the Welsh Borderlands. Geol. Ser. No. 10, Natl. Mus. of Wales, 133 p. Cardiff, Wales.

Sloss, L.L., 1963, Sequences in the cratonic interior of North America. Geol. Soc. Amer. Bull. v. 74, p. 93-113.

Swartz, C.K., 1923, Stratigraphic and paleontologic relations of the Silurian strata of Maryland, Maryland Geological Survey, Silurian, p. 25-50.

Tankard, A.J., 1986, On the depositional response to thrusting and lithospheric flexure: examples from the Appalachian and Rocky Mountain basins. Spec. Publs. Int. Assoc. Sediment. v. 8, p. 369-392.

Vail, P.R. and Mitchum, R.M., Jr., 1977, Seismic stratigraphy and global changes of sea-level, part I: overview, In Payton, C.E., ed., Seismic Stratigraphy -- applications to hydrocarbon exploration. Amer. Assoc. Petrol. Geol. Soc. America. Bull. v. 100, p. 311-324.

Vail, P.R., Mitchum, R.M., Jr., and Thompson, S., et al., 1977, Seismic stratigrpahy and global changes in sea level. Parts 2-4. In Payton, C.E., ed., Seismic stratigraphy-applications to hydrocarbon exploration. Am. Assoc. Petrol. Geol. Mem. 26, p. 83-97.

Van der Voo, R., 1988, Paleozoic paleogeography of North America, Gondwana and intervening displaced terranes: Comparisons of paleomagnetism with paleoclimatology and biogeographical patterns. Geol. Soc. America Bull. v. 100, p. 311-324.

Van Tyne, A.M., 1966, Progress Report - subsurface stratigraphy of the pre-Rochester Silurian rocks of New York: Proceedings of the Symposium, Petroleum Geology of the Appalachian Basin, Pennsylvania State University, p. 97-116.

Van Wagoner, J.C. Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F., Loutit, T.S. and Hardenbol. J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions. In Wilgus, C.K. et al., eds., Sea-level changes: an integrated approach. SEPM Spec. Pub. v. 42, p. 39-46.

Wheeler, H.E., 1963, Post-Sauk and pre-Absaroka Paleozoic stratigraphic patterns in North America. Amer. Assoc. Petrol. Geol. Bull. v. 47, p. 1497-1526.

Wilgus, C.K., Hastings, B.S., Kendall, C.G., Posamentier, H.W., Ross, C.A., and Van Wagoner, J.C., eds., 1988, Sea-Level Changes: An Integrated Approach. SEPM Spec. Pub. 42, 407 p.

Woodrow, D.L., 1985, Paleogeography, paleoclimate, and sedimentary processes of the late Devonian Catskill Delta. In Woodrow, D.L. and Sevon, W.D., eds., The Catskill Delta, Geol. Soc. Amer. Spec. Pap., no. 201, p. 51-63.

Zenger, D.H., 1965, Stratigraphy of the Lockport Formation (Middle Silurian) in New York State. New York State Museum and Science Service Bull., v. 404, p. 210.

Zenger, D.H., 1971, Uppermost Clinton (Middle Silurian) stratigraphy and petrology, east-central New York. New York State Museum and Science Service Bull., v. 417, 58 p.

ROAD LOG FOR SILURIAN SEQUENCES

Note: From Fredonia take I-90 (NY State Thruway to junction of I-290 (Youngman Highway). Take I-290 westbound (toward Niagara). Road log begins at junction of I-90 and I-290.

CUMULATIVE MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Junction I-90 (North)/I-290 west; exit onto I-290 west.
0.4	0.4	Main Street Exits;
0.5	0.1	Note road cuts in cherty Onondaga Limestone (Middle Devonian). This is the old Vogelsanger Quarry.
1.4	0.9	Exit at Route 324/240.
3.0	1.6	Exits for Millersport Highway
3.8	0.8	Cuts on left at underpass are in Upper Silurian Camillus Shale (and gypsum).
3.9	0.1	Exit 4 for I-990 North; take exit to right onto I-990.
4.7	0.8	Exit 1 for SUNY Buffalo
5.7	1.0	Exit 2 for Sweet Home Road
6.5	0.8	Exit 3 for Audubon Parkway
8.0	1.5	Exit 4 for North French Road (end I-990).
8.2	0.2	Junction North French Road; turn left (west).
8.6	0.4	Junction NY Rt. 270; turn right (north).
11.0	2.4	Bridge over Tonawanda Creek, enter Niagara County.
12.8	1.8	Junction Bear Ridge Road, Town of Pendleton.
14.0	1.2	Junction Fiegel Road.

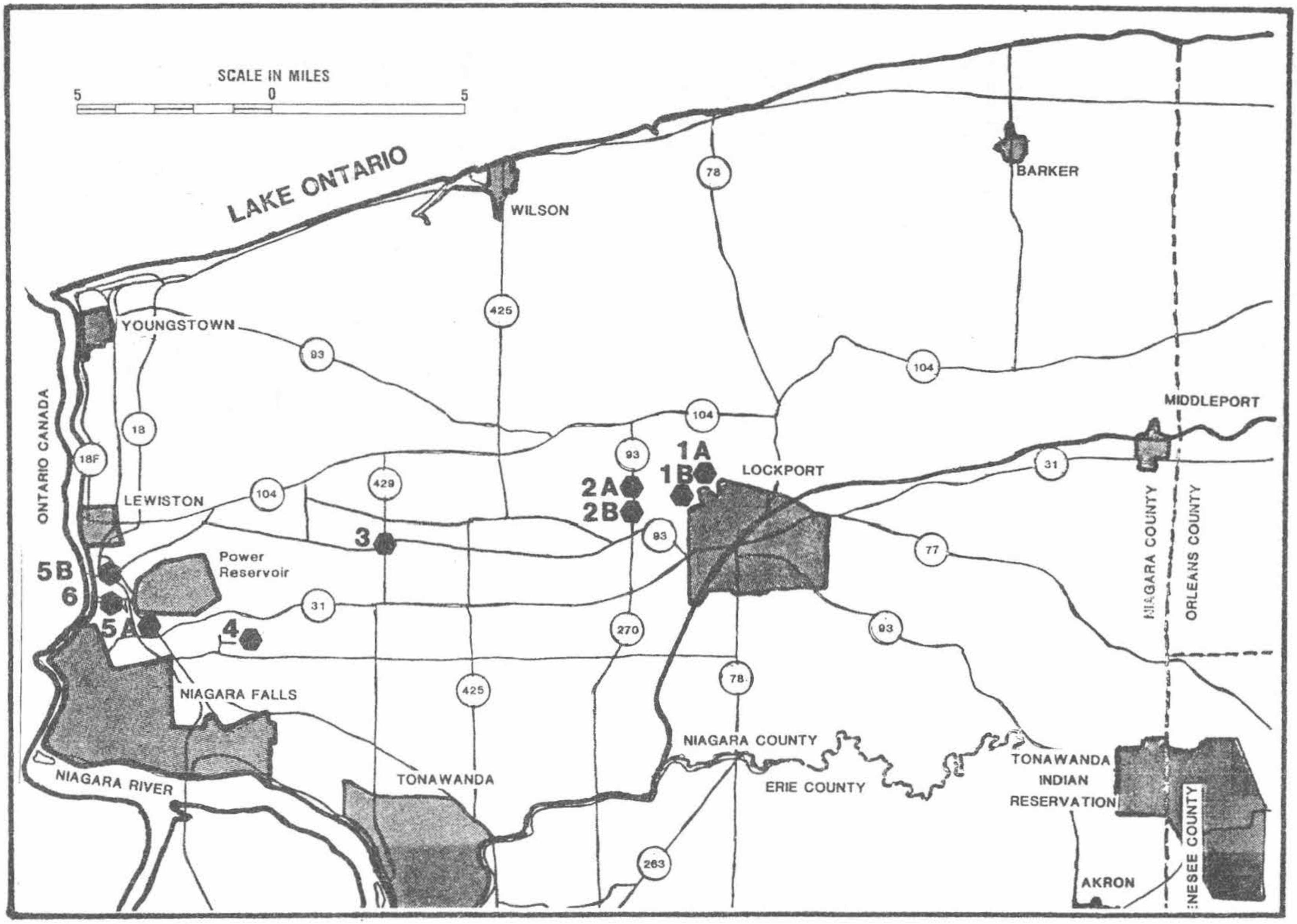


Figure 1b. Map of Niagara County, New York showing location of field trip stops.

15.0	1.0	Junction Mapleton Road.
15.8	0.8	Junction Lockport Road; Route 270 becomes Lockport/Cambria Townline Road.
17.0	1.2	Junction Saunders Settlement Road (Route 31).
18.5	1.5	Junction Upper Mountain Road; Route 270 ends. Turn right (east) onto Upper Mountain Road (Route 93 east). Town of Lockport, NY.
19.0	0.5	Junction Gothic Hill Road (type section of lower Gasport Fm. grainstone is cut on this road). Turn right (north).
19.5	0.5	Junction Niagara street (lower Mountain Road). Turn right (east).
20.15	0.6	Junction Sunset Road.
20.4	0.25	Junction West Jackson Street; turn left (northeast).
21.2	0.8	Small cut in red Grimsby Sandstone.
21.5	0.3	Note sewage treatment plant on left.
21.7	0.2	Junction Plank Road; turn right and pull into parking area near base of railroad viaduct.

STOP 1: ROAD CUT ON W. JACKSON STREET BELOW SOMERSET RAILROAD VIADUCT (QUEENSTON SHALE - MEDINA GROUP).

This excellent and relatively new outcrop has been described previously in considerable detail (see Friedman, 1982; Duke et al., 1987). It provides an outstanding exposure of the basal Silurian Cherokee Unconformity and a good opportunity to study the lower units of the Medina Group as well as the uppermost beds of the Queenston Shale.

About 6 m (20') of upper red mudstones and siltstones of the Queenston Formation (Upper Ordovician, Ashgillian) are exposed at this locality. The Queenston has been interpreted as either very shallow marginal marine or non-marine red beds. The Cherokee unconformity in this area is the basal surface of Whirlpool Sandstone which is nearly planar. It is also a megasequence boundary separating the early or Creek phase of Sloss' Tippecanoe "megasequence" from the later Tutelo phase (see Dennison, 1989).

The Medina Group (Sequence I of the Silurian System), consists of an Early Silurian (early Llandovery, A1-B), siliciclastic wedge derived from tectonic source areas to the southeast (Figs. 1, 2). The lowest Silurian unit, white Whirlpool sandstone is about 3.5 m (11.5') thick at this location. Basal beds of the Whirlpool Sandstone are quartz arenites with northwest dipping cross strata which have been interpreted recently as non-marine, braided stream deposits (Middleton et al., 1987). Large-scale channel-like structures occur in, or at least at the top of, these sands. Shale drapes within such channels at Lockport have yielded marine acritarchs (M. Miller, unpublished data) indicating that the channels were backfilled by very shallow marine sands and minor muds during a rise of sea level. Hence the irregular channeled surface that separates lower Whirlpool braided fluvial facies from upper Whirlpool hummocky cross-stratified, sparsely fossiliferous beds is a transgressive surface. The Whirlpool Sandstone thus is interpreted to contain both a lowstand (or shelf margin) systems tract and a transgressive deposit. A thin bed containing phosphatic pebbles and fossil grains occurs the Whirlpool-Power Glen contact. This phosphatic pebble bed may mark a marine flooding surface, or surface of maximum starvation associated with relatively increased rates of sea level rise. This surface marks the change from shallow shelf sands of the upper Whirlpool into deeper shelf muds and storm sands of the Power Glen Formation, herein interpreted as relative highstand deposits (subsequence IA). The Power Glen exhibits two small-scale parasequences.

This outcrop is one of the easternmost exposures of the Power Glen Shale. At this locality the Power Glen Shale comprises of about 5 m (16') of greenish gray shale with thin tempestitic siltstone and sandstone beds. The basal meter-thick transitional zone consists of thin (2-10 cm) muddy sandstones with interbedded sandy shales. Sandstones in the Power Glen Shale feature small-scale hummocky lamination and gutter casts suggestive of shallow, storm influenced shelf deposition. Small burrows (Planolites) are common, but body fossils are rare.

Greenish to reddish sandy shales and reddish sandstones occur near the top of the Power Glen suggesting a minor upward shallowing trend. However, the top of the unit (as defined herein) is sharply demarcated at the base of a massive white to pink mottled sublitharenitic sandstone about 2.5 meters (7.7') thick. This unnamed unit corresponds with lenticular sandstone beds ("Devil's Hole Sandstone" of Duke et al., 1987) seen at Niagara Gorge (Stop 6). The basal and upper beds of the sandstone contain lingulid brachiopods and probable Lingula burrows.

The white sandstone appears to record a relative sea-level drop during which sands were distributed widely into the basin. The unit has some characteristics in common with the upper member of the Whirlpool Sandstone and, by analogy, is considered to be a relative lowstand to transgressive deposit.

The unnamed sandstone, in turn, is overlain by about 2.0 meters of brick red shales and interbedded sandstones assignable to the lower Grimsby Formation. These beds are ferruginous and exceedingly rich in fragments of lingulids with rare nautiloids and bryozoans. Thin spastolithic (oolitic) hematite stringers occur near the top of the unnamed sandstone and probably reflect reworking of sediments in shallow marine environments during an interval of sediment starvation. Hence, these shell-rich ferruginous sediments represent a condensed interval at the base of the Grimsby highstand deposits.

The reddish marine shales near the base of the Grimsby Formation pass upward into red and white-mottled sandstones and thin sandy shales. These beds are exposed high in the cut and are not readily accessible. This upper interval will be seen to better advantage at Niagara Gorge (Stops 4, 5, 6).

Return to vehicles and reverse direction turning left at West Jackson Street.

23.1	1.4	Junction Niagara Street
		If time permits, we may make an optional stop at the upper railroad cut. In such case we will turn left proceede for 0.5 miles up the Niagara Escarpment, and pull off into parking area just before Sommerset Railroad crossing. Then proceed on foot northeast along the railroad for about 0.4 mile to outcrops.

OPTIONAL STOP 1B: CUTS ALONG SOMERSET RAILROAD, NORTHEAST OF NIAGARA STREET CROSSING, LOCKPORT, NY (UPPER MEDINA AND CLINTON GROUPS)

Cut banks along both sides of the railroad track provide an excellent exposures of the upper beds of the Medina and Lower Clinton groups (Fig. 9). Lowest exposed units are reddish and white, mottled sandstones and sandy shales of the upper Grimsby Formation. A prominent 2.2 meter thick, pale pinkish sandstone unit (Thorold-equivalent), the overlying 3.0 meters (9') of red to greenish silty shale (Cambria Member), and about 0.5 m of pale greenish to white sandstone and sandy shale, referred to as the Kodak Formation, complete the Medina Group.

The Lower Clinton Group (Sequence II) rests unconformably on the Medina Group with the contact marked by a thin phosphatic, calcareous sandstone. Only 24 cm (17") of Neahga Shale (Fig. 9) overlie the unconformity, followed by 1.9 m of crinoidal grainstone belonging to the Hickory Corners Member of the Reynales Formation. These units and the underlying Medina strata are comparable to the section seen at Lockport Junction Road (Stop 2A), and are described in greater detail for that section.

The upper contact of the Reynales Limestone (upper sequence II boundary) is well exposed along the top of the cut on the southeast side of the railroad tracks. Here, the uppermost bed of Reynales crinoidal packstone crops out from beneath the soil cover. The surface is an irregularly sculpted and bored hardground with black phosphatic staining and what appear to be phosphatized stromatolites. This is a local manifestation of the major Upper Clinton unconformity.

Southwest of the main railroad cut are small, partially overgrown exposures along the southeast side of the tracks. Here a thin remnant of the upper Clinton Group (sequence IV and parts of V) can be observed. In the ditch, southwest of the end of the main outcrop are exposed about 37 cm of dark gray shale rich in phosphatic nodules. The unit, which yielded an acritarch assemblage identical to that of the Williamson Shale of the Rochester area, lies between the top of Reynales unconformity and the basal phosphatic bed of the Rockway Dolostone. The pale buff-weathering Rockway Member (0.5 meters thick) and its contact with the overlying pinkish gray grainstones of the Irondequoit Formation are visible in the small outcrop.

Rochester Shale is poorly exposed in slumped banks between this outcrop and the Niagara Street crossing. Well-preserved bryozoans, brachiopods, and even cystoid calyces are occasionally found in patches of weathered shale along the southwest side of the tracks.

Return to vehicles and reverse directions, returning to the junction of Niagara Street and West Jackson Street.

23.1	0.4	Junction Niagara Street and West Jackson Street. Turn or continue west on Niagara Street.
23.5	0.4	Junction Sunset Road
23.9	0.6	Junction Gothic Hill Road

24.3	0.4	Junction Leete Road
24.5	0.2	Overpass over Lockport Junction Road (Route 93).
24.55	0.22	Access road to Route 470 on left; turn left onto road.
24.6	0.05	Pull off along berm on left side of road and park. Proceed on foot directly down embankment along Route 98 to road cut.

STOP 2A: LOCKPORT JUNCTION ROAD CUT (LOWER) (UPPER MEDINA AND LOWER CLINTON GROUPS)

This newly widened (1986) cut along Lockport Junction Road exposes a similar section to that seen along the the Somerset Railroad cut (optional stop 1B). However, this cut does not show the upper Clinton beds.

The lowest beds are exposed beneath and just north of the overpass of Lower Mountain Road over Route 93. The basal units seen here are red shales near the top of the Grimsby Formation. These shales are overlain by a 1.0 to 1.2 meter thick, blocky, pinkish gray sandstone that displays color mottling due to bioturbation. Swirly spreiten of the trace fossil Daedalus occur sporadically near the top of the sandstone ledge. Detailed regional correlation by Duke and Fawcett (1987) indicates that this unit is the equivalent of the Thorold Sandstone at the Niagara Gorge. The upper contact of the sandstone is marked by a thin red silty bed containing a "hash" of lingulid shell fragments.

The Thorold, in turn, is overlain by 1.7 to 1.8 m of dominantly red silty shale of the (Cambria Member) uppermost Medina Group. This shale bears distinctive fine, white mottling due to bioturbation. Although previously assigned to the Grimsby Formation, this is a distinctive, ostracode-bearing shaley unit that overlies the Thorold Sandstone and is traceable regionally at least as far east as the Rochester area. We informally identify this unit as the Cambria Shale Member of the Thorold Formation, using this section as the type locality.

The upper portion of the Cambria Member is pale purplish to greenish gray sandstone and sandy shale that was formerly termed Thorold Sandstone. In actuality, this is simply a leached sandy zone in the Cambria Shale. The Kodak Sandstone, observed at Stop 1B, and about 1 m of Cambria Shale, have been removed here at the sequence I/II unconformity. Sandstones contain small Skolithos burrows and intercalated green shale beds, especially the topmost layer, contain prolific leperditiid ostracodes. The greenish color extends down about 30 to 50 centimeters below the upper contact where sandstones are mottled pale purple and green. This discoloration is probably associated with the top unconformity and deposition of overlying reducing sediments (Duke et al., 1987).

Here, as at the Lockport railroad cut, the top of the Medina Sandstone is an erosion surface overlain by a thin (3 to 5 centimeter) dark gray, phosphatic sandy limestone (Densmore Creek Bed) with prolific Hyattadina brachiopod valves. A thin laminated siltstone rests on the bed at the contact with the greenish gray Neahga Shale, which at this locality is about 1.0 meter thick--substantially thicker than at Lockport (STOP 1B).

At the base of the Hickory Corners Limestone (Reynales) is a thin (3 to 5 centimeter) pyritic, sandy limestone packed with black phosphatic pebbles and shell fragments (Budd Road phosphatic Bed). It is overlain by about 40 centimeters of alternating greenish gray shales and thin limestones capped by a 60 centimeter thick ledge of nodular crinoidal pack- and wackestone at the top of the roadcut. These beds contain a prolific fauna including corals, brachiopods (Hyattidina, Dalejina, Platystrophia) and pentameric crinoid stems belonging to a newly described inadunate. The upper surface at this locality is a glacially striated pavement and the post-Reynales erosion surface cannot be observed.

Reboard vehicles and continue on exit lane.

24.7	0.1	Junction with Lockport Junction (Townline) Road (Route 93); turn right (south) and proceed up escarpment.
25.0	0.3	Pull into small parking area at gap in guard rail (for driveway) to left.

STOP 2B: LOCKPORT JUNCTION ROADCUT (UPPER) (UPPER ROCHESTER, DECEW, GASPORT FORMATIONS)

This large roadcut displays the upper part of the Rochester Formation (about 5 meters), DeCew Dolostone (2.5 meters), and Gasport Limestone (over 7 meters); it was described in detail as Stop 1 of Brett's (1982) NYSGA Guidebook article. However, since that time the exposure has been freshly blasted to widen Route 93.

A subsequence boundary occurs between the dolomitic calcisiltites and shales of the upper Rochester Shale and the overlying DeCew Dolostone. The latter is a buff-weathering, silty dolostone with thin layers of intraclasts and contorted bedding. The basal contact is sharp and locally channeled, but nearly conformable. The upper Rochester and DeCew are burrowed in some levels; body fossils are rare but this exposure has produced long columns of crinoids in the lower DeCew.

The sharp and undulatory upper contact of the DeCew Dolostone with the Gasport Formation forms the boundary between the Clinton and Lockport Groups, and is interpreted as a sequence bounding unconformity (between sequences V and VI). The surface represents an abrupt lowering of relative sea-level and beveling of older strata. The basal Gasport bed is a greenish gray brachiopod-rich, crinoidal grainstone conglomerate with dolostone clasts eroded from the DeCew. Missing at this contact is the newly recognized Glenmark Shale, a fossiliferous gray shale with litho- and biofacies resembling the Rochester Shale.

Only the lower (Gothic Hill) member of the Gasport is present here. At this locality and at cuts along adjacent Gothic Hill Road this unit is exceptionally thick (7-8 meters) and composed of well to poorly sorted pelmatozoan pack- and grainstones. This facies is interpreted as a high energy crinoidal bank. Brachiopods are common in an argillaceous, thin-bedded unit near the top of the unit. A small bioherm composed of algal and bryozoan boundstone (micrite) occurs within the Gothic Hill Member. The overlying units will be examined at the next stop.

Reboard vehicles and turning left out of the driveway, continuing south along Route 93.

25.3	0.2	Upper end of roadcut
25.5	0.3	Junction Upper Mountain Road (Rt. 270/93). Turn right (west).
26.2	0.7	Junction Thrall Road on right. Exposures of fossiliferous Rochester Shale along Thrall Road were described as Stop 2 in Brett (1982).
28.0	1.6	Junction Blackman Road.
28.9	0.9	Junction Cambria Road (North).
30.1	1.2	Junction Route 425 (South).
30.2	0.1	Junction Route 425 (North). Shawnee Road.
30.9	0.7	Junction Baer Road.

32.2	1.3	Pekin Village; fire department on left.
32.3	0.1	Junction Old Pekin Road (one way). Turn right and proceed down hill.
32.6	0.3	Park on shoulder before stop sign at junction with Route 429 and walk to outcrop along Route 429 turning left (up hill) at intersection.

STOP 3: PEKIN HILL (ROUTE 429) ROAD CUT AT UNDERPASS BENEATH UPPER MOUNTAIN ROAD. (GASPORT-GOAT ISLAND FORMATIONS).

This classic exposure, described in detail by Crowley and Poore (1974), displays biohermal structures characteristic of the lower Lockport Group. Although the exposure is becoming overgrown, it has been restudied recently as part of our stratigraphic revision.

The lowest stratum, barely visible in the northernmost end of the outcrop (east side of road) is a crinoidal grainstone which marks the top of the Gothic Hill Member (seen at the last stop). The crinoidal grainstone is overlain on the east side of the road cut by approximately 4 to 5 meters of thin-bedded, bioturbated argillaceous dolostone which Brett et al. (in prep) refer to as the Pekin Member of the Gasport Formation. A small bioherm, "rooted" in the underlying Gothic Hill Member, protrudes upward into the sparsely fosssiliferous Pekin Member on the west side of the road cut.

Of particular importance is the sharp upper contact of the Pekin Member with pink, fine-grained crinoidal and cladoporid coral-bearing dolomitic grainstones of the basal (Niagara Falls Member) of the Goat Island Formation on the east side of the road. This contact can be traced to near the concrete wall for the Upper Mountain Road overpass. Here, it is seen to drop in elevation, apparently defining a channel-like feature that underlies stromatoporoid-rich biohermal dolomicrites. The latter appear to laterally replace the pink grainstone facies of the Goat Island.

On the west side of the road, the main mass of the bioherm described by Crowley and Poore (1974) can be seen. It is evidently a continuation of the biohermal mass seen near the bridge footing on the east side. This bioherm appears to occur above the sharp and distinctly irregular contact of the Goat Island Formation. Again, this contact is channelized into the dark, argillaceous dolostones of the underlying Pekin Member. Hence the so-called "Gasport bioherm" at this locality appears to be a biohermal mass within the lower Goat Island Formation.

The bioherm consists of dolomicrite containing very abundant, large stromatoporoid heads, many of which are re-oriented or overturned and hence, are not preserved in situ. Near the center of the biohermal mass, the dolomicrite facies passes abruptly into pale pinkish gray fine-grained grainstone, closely resembling the non-biohermal basal Goat Island facies on the east side of the road. The pale gray "reefy" dolomicrite facies appears both north and south of the grainstone pocket which appears to have filled the center of the large channel-like feature cut into the top of the Pekin Member. Also note two or more sharply defined surfaces within the biohermal masses that may represent minor erosion surfaces. Overall, we interpret the bioherm here as a possible bank of algally (?) bound micrite and stromatoporoids (and a few tabulates) that lined a (storm surge?) channel cut into the muddy carbonates of the Pekin Member. The center of the channel was occupied by fine carbonate sand but was eventually "overrun" by the biohermal facies developing inward from the two channel margins. The channel itself was apparently oriented NW/SE, roughly perpendicular to regional facies strike. The erosion of the Pekin Member mudstones may have occurred during a relative sea-level lowstand that marks the boundary between Gasport and Goat Island depositional subsequences. (For a radically different interpretation of the stromatoporoid facies as a late successional stage of a Gasport bioherm, see Crowley, 1973, and Crowley and Poore, 1974).

The outcrop continues south of the Upper Mountain Road bridge. Here the upper portion of the stromatoporoid biohermal mass is abruptly overlain (contact now obscured) by thin-bedded, fine-grained, buff-weathering dolomicrite with bands of white chert nodules. We refer to this facies as the middle or Ancaster Member of the Goat Island Formation. Note a distinctive marker bed of silicified _Whitfieldella_ brachiopods. The cherty facies records a general deepening trend in the upper Goat Island. It is somewhat comparable to the Pekin Member of the Gasport, but the latter is not cherty.

Return to vehicles and drive to intersection with Route 429. Turn _left_ (south) at Route 429 (as you did on foot) and proceed up through the cut.

32.6	0.1	Upper end of road cut.
32.8	0.2	Junction Grove Street (access to Upper Mountain Road).
34.4	0.6	Junction Route 31 (Saunders Settlement Road). _Turn right (west)_.

35.1	0.7	Junction Bridgeman Road.
36.2	1.1	Junction Chew Road.
36.9	0.7	Junction Walmore Road. Turn left (south).
37.5	0.6	Jog in Walmore Road.
38.5	1.0	Junction Lockport Road at 84 Lumber. Turn right (west).
39.0	0.5	Niagara Airforce Base on left.
39.6	0.6	Tuscarora Road (on right).
40.0	0.4	Road forks; proceed straight ahead on Lockport Road.
40.5	0.5	Railroad overpass.
40.7	0.2	Junction Miller Road. Turn right (north).
40.75	0.05	Junction Quarry Road. Turn right (east).
41.3	0.55	Office headquarters of Niagara Stone Quarry. Check in then proceed by vehicles into quarry.

STOP 4: NIAGARA STONE QUARRY (LOCKPORT GROUP)

This active quarry provides the most complete section of the Lockport Group in the Niagara region. Aspects of the jointing and hydrogeology of the Lockport at this locality are discussed in detail in the contribution by Tepper et al., this guidebook). We will briefly examine the overall stratigraphy focusing on the Goat Island Formation. The section extends from the top of the DeCew Dolostone and its unconformable contact with the overlying Lockport Group (sequence V/VI boundary) to beds near the base of the Guelph Formation at the quarry rim. The lower Gasport limestones appear to fall in the *sagitta* conodont zone making them late Wenlockian in age. Furthermore, recent graptolite discoveries indicate that the Wenlockian/Ludlovian boundary lies close to the Gasport/Goat Island contact (LoDuca and Brett, in press). Kleffner (1989) obtained *siluricus* zone conodonts from the upper Eramosa beds here.

The DeCew Dolostone is observable in the deepest part (lowest lift) of the quarry where it exhibits a contorted upper surface and internal hummocky cross-stratification. The sequence bounding unconformity between the DeCew and the Gasport Formation of the Lockport Group is marked by a thin, greenish (glauconitic) grainstone containing rip-up clasts of DeCew Dolostone up to 10 centimeters across. The lower or Gothic Hill Member of the Gasport is a massive, 5 meter thick interval of light pinkish gray, commonly orange weathering, crinoidal grainstone. The overlying Pekin Member is a 5 meter-thick interval of thin to medium-bedded, medium gray argillaceous dolostone. A bioherm extends up through the interval in the northwest corner of the deep pit.

The lower Goat Island (Niagara Falls Member) can be examined along the top of the ramp road leading to the lowest pit of the quarry, although the basal contact with the Pekin Member is obscure and inacessible. The upper 1.5 meters of the Niagara Falls Member is a distinctive interval (Niagara Falls C unit) consisting of very dark gray, vuggy dolostone with distinct, white weathering stromatoporoids and corals. These and immediately underlying beds reflect a relative sea-level lowstand corresponding to the biohermal bed of Pekin cut. Beds overlying the Niagara Falls Member are assigned presently to the Ancaster and Vinemount Members of the Goat Island Formation. Both of these units have their type sections near Hamilton, Ontario.

The middle Goat Island interval is well developed in the walls above the second lift. The platform below is bulldozed off on a surface within the Ancaster Member. Thin flaggy beds of pale buff-weathering, fine-grained dolostone in the lower half meter of contain the chert nodules that characterize the Ancaster Member in its type area. However, the overlying interval is medium to thick-bedded, buff colored, medium-grained and generally non-cherty dolostone. A distinctive layer of silicified and valves of <u>Whitfieldella</u> brachiopods occurs about 1.5 meters above the second lift. This layer may correlate with the <u>Whitfieldella</u> bed seen at the Pekin road cut (Stop 3).

Ancaster facies are interpreted to represent a similar offshore, low energy shelf environment. The abundance of chert in the Ancaster is unique among the Silurian units in the Niagara region; the cherts may reflect the availability of biogenic silica from the abundant sponges which occupied the seafloor at this time, although this is probably not a complete explanation.

The uppermost half meter of the Ancaster Member is marked by numerous vugs that appear to represent small corals and stromatoporoids. This bed is abruptly overlain by a layer of dark gray dolomitic shale that marks the base of a 3 m-thick bituminous argillaceous unit. Detailed outcrop and subsurface tracing of the unit indicates that it is coextensive with the Vinemount shale beds of the Hamilton, Ontario region. These argillaceous beds appear to grade laterally and vertically into Ancaster cherty dolostone beds and hence both are assigned member status in the Goat Island Formation. The Vinemount here comprises about 3 meters of dark gray, bituminous and argillaceous dolostone with some shale partings. Fossils are rare and poorly preserved, but include rhynchonellid and atrypid brachiopods, small rugose and favositid corals and the large nautiloid _Dawsonoceras_.

Together, the three members of the Goat Island Formation comprise an upward deepening (retrogradational) subsequence. It is marked at the base by an erosion surface between the Pekin Member of the Gasport Formation and the Niagara Falls Member of the Goat Island. The top of the Niagara Falls Member constitutes a marine flooding surface. Cherty beds of the basal Ancaster reflect relatively slow deposition during the early highstand.

The upper quarry walls above the Vinemount display the entirety of the Eramosa Formation (formerly called Oak Orchard Member; see Zenger, 1965). Recent work (Brett et al., in press) indicates that "Oak Orchard" is an invalid term since the highest unit is exposed on Oak Orchard Creek is equivalent to the Vinemount beds. Furthermore, tracing of key units into Ontario indicates their correlation with the Eramosa and lower Guelph Formations in the type areas of Canada.

A number of informal units are recognizable within the Eramosa and are visible in the wall. Lowest is a 1.8 meter thick, massive, biostromal bed containing white gypsum filled vugs. This is succeded by a rather nondescript, fine-grained facies 3.5 m which, in turn, is overlain by biostromal or thrombolitic dolomicrite. Notable are the two flaggy-weathering intervals at the top and base of this zone that may represent minor discontinutiies. Toward the top of the quarry are 6 m Eramosa Formation, containing large, lenticular stromatolitic bioherms observable in the south wall.

Overlying units, at or near, the quarry rim are about 1.5 to 2.0 m of biostromal brownish gray dolostones with very abundant _Favosites_ corals some of which are infilled by purple fluorite. The topmost bed in parts of the quarry is a light gray-weathering, stromatolitic (LLH) dolostone which forms a regional marker.

The Eramosa interval appears to represent an upward shallowing succession (subsequence) reflecting late highstand conditions.

Reboard to vehicles and return to quarry entrance.

41.9	0.6 (approx.)	Miller Road. _Turn left_.
41.95	0.05	Junction Lockport Road. _Turn right (west)_.

42.15	1.1	Junction Military Road (Route 265). <u>Turn</u> <u>right</u> <u>(north)</u>.
43.75	0.8	Junction Route 31 (Saunders Settlement Road). Go straight.
44.65	0.9	Junction back access road (marked for <u>Deliveries</u>) for Robert Moses Power Plant. <u>Turn</u> <u>left</u> <u>(west)</u>.
44.60	0.15	Underpass beneath Route I-190. Outcrop of stromatolites (Stop 5).
45.0	0.2	Turn right into parking area near chain link fence overlooking Robert Moses forebay. Examine upper Lockport strata in forebay and proceed on foot to roadcut Stop 5A.

STOP 5A: ROBERT MOSES ACCESS ROAD AND FOREBAY (UPPER LOCKPORT GROUP)

Higher units of the Lockport Group are visible along the fore bay and access road to the Robert Moses Power Plant, just west of Military Road and 0.9 miles north of Route 31. Here we will briefly examine, from a distance, exposures in the forebay canal of the upper Eramosa (new usage in New York) and lower Guelph Formations. Note the large algal bioherms which characterize the uppermost units of the Eramosa Formation in Niagara County (seen at Stop 4). The highest units in the forebay canal are more or less tabular stromatolitic dolostone which are, at present, assigned to the basal beds of the Guelph Formation (Fig. 16). Exposures in the small road cut at the underpass of the access road beneath the lanes of I-190 contain exceptionally large stromatolites. These strata represent a very distinctive marker horizon in the basal Guelph beds in western New York. The stromatolite heads are approximately 2 meters across but superimposed on these are small digitate upward growths of stromatolitic boundstone. The stromatolitic horizon is thought to be traceable at least to Hamilton where it comprises the basal transition bed between the Eramosa and Guelph.

Reboard vehicles and return to Military Road.

45.3	0.3	Junction Military Road. <u>Turn</u> <u>left</u> <u>(north)</u>.

45.7	0.4 0.5	Pass by forebay area (on left) and Robert Moses Pump Generating Station (on right).
46.5	0.8	Junction Upper Mountain Road (right) and entrance road to Route I-190, Route 104, and Robert Moses Parkway (left). <u>Turn left onto entrance road</u>.
46.55	.05	Exit for Route I-190 north onto Lewiston-Queenston International Bridge.
46.75	0.2	Exit for Route I-190 south.
46.75	0.2	Exit for Route 104 east (north) to Lewiston.
47.10	0.15	Exit for Route 104 west (south) to Niagara Falls. <u>Bear right onto exit ramp</u>.
47.2	0.1	Merge onto Route 104.
47.4	0.2	Stop light at Irving Drive. Turn <u>left</u> (east).
47.45	0.05	Old Lewiston Road. <u>Turn right</u>. Park vehicles near small building on right. Proceed on foot through railroad underpass (under Route 104). At far end turn left, walk up slight embankment then down other side to exit lanes onto Robert Moses Parkway. Turn Turn left and walk across exit lane and carefully cross two lanes of Rober Moses Parkway. At 55 mph sign step step over guard rail and down a slope bearing right (north) to cliffs along Niagara Gorge. (Note: permission should be obtained from Niagara Parking Commission for access to this stop.)

STOP 5B: NORTH END, NIAGARA GORGE BETWEEN LEWISTON-QUEENSTON BRIDGE AND ROBERT MOSES POWER PLANT (CLINTON GROUP)

The Rochester Shale in this cliff exposure along the east face of Niagara Gorge was described in detail by Brett (1982). It provides an excellent section of the entire Clinton and lower Lockport units. It also provides excellent fossil collecting in the Rochester Shale.

Proceed down to a small ditch in the gorge nearly opposite the Irondequoit-Rochester contact in the main cliff exposure. At its base, this gully exposes orange-weathered Thorold Sandstone, overlain by a thin (about 0.5 m) maroon and pale green sandy shale situated beneath the Neahga Shale. This is the last vestige of the Cambria Shale. To the south and west it has been lost to pre-Neahga erosion. A thin, Hyattidina-bearing, calcareous sandstone (Densmore Creek Bed equivalent) occurs at the base of the Neahga Shale. The latter is about 2 meters thick, dark gray, fissile gray shale. The overlying Reynales Limestone with a thin basal phosphate bed (Budd Road Bed) is nearly identical to that seen at Lockport Junction Roadcut (Stop 2). Note thick, nodular, coral-bearing packstone at the top.

Exposures in the main wall of the gorge show the basal contact of the overlying Rockway Dolostone. A 10-15 cm phosphatic, black shale resting on the Reynales represents a thin tongue of the Williamson Shale. This contact is sharp and represents a major unconformity at which most of the lower and middle Clinton units have been beveled (compare STOP 1B).

The overlying Rockway Member (of the Irondequoit Formation, Kilgour, 1963) is a buff-weathering, argillaceous dolostone with thin shales. The Rockway shows prominent bands, ranging from a few centimeters to half meter thick, of argillaceous dolomite, which are interbedded with thin green shales. These two interbedded facies define a small scale cyclicity which can be traced on a regional basis. Rockway beds apparently record relatively sediment starved, off-shore marine lime and siliclastic muds. Extensive dolomitization has obscured the fossil content of this unit. However, large brachiopods (Costistricklandia) are locally abundant, particularly in the middle thick bedded unit of the Rockway. Dendroid graptolites, rare favositid corals and nautiloids fossils are occasionally found. On the basis of conodont biostratigraphy, the Rockway belongs to the upper part of the amorphognathoides zone (latest Llandovery to earliest Wenlock Series).

It should be noted that the rock, here termed Rockway Member of the Irondequoit Formation, whose type section is in Canada (15 Mile Creek at Rockway, Ontario), is commonly referred to as the upper beds of "Reynales" limestone by Canadian geologists. However, the Sequence IV boundary unconformity, cuts out the true Reynales Limestone just west of the Niagara Gorge. Hence, the true Reynales is absent in Canada, except along the Niagara River.

A sharp contact separates the Rockway from the overlying massive, crinoidal grainstones of the upper (Model City) member of the Irondequoit. Clasts of fine-grained dolostone derived from the underlying Rockway occur in the basal thin bed of the Model City.

The upper member of the Irondequoit Formation (Model City Member of Brett et al., in press), consists of massive, pinkish-gray, crinoidal limestone and dolostone which rests with the sharp and slightly disconformable surface on the upper shaley, beds of the Rockway Member (Subsequence IVB). In places a thin phosphate or pyrite crusted surface delimits the upper surface of this 6 inch band. This pyritic crust probably marks a hardground. The Model City member shows a minor trend of upward deepening toward the top and a conformable, although abrupt, contact with the overlying Rochester Shale. Faunal density is interpreted as representing a shallow, highly agitated (near wave base) crinoidal shoal environment; brachiopod assemblages suggest a benthic assemblage-(BA) 3 paleobathymetry. Conodonts indicate an earliest Wenlockian age.

The Model City in turn, is interpreted as a transgressive systems tract of subsequence VA (Fig. 15). Its sharp upper contact with the Rochester Shale, which is locally glauconitic, is considered to be a surface of maximum starvation (marine flooding surface). In the cliff opposite the small ditch (which exposes the Thorold-Reynales) a small micritic (algal?) bryozoan bioherms protrude from the top of the Model City into the overlying Rochester Shale.

The overlying Rochester Shale, which is about 18 m thick, is a gray calcareous to dolomitic mudstone with numerous thin carbonates representing shell-rich and carbonate-silt storm deposits (Fig. 15). It can be examined by walking south along the base of the cliff toward the Robert Moses Power Plant. Though the Rochester initially appears homogeneous, it shows a distinct internal cyclicity which is traceable very widely. Most notable is the medial band of bryozoan-rich carbonates that occurs about 9 m above the base of the Rochester Shale. The sharp top of this unit has been used to demarcate the boundary between the Lewiston and Burliegh Hill Members of the Rochester Shale (Brett, 1983).

The Lewiston Member itself displays an alternation of fossiliferous limestones and calcareous shale intervals with nearly barren shale. Bryozoan rich beds occur up to about 10 feet above the base (Lewiston B unit). This fossiliferous unit is followed by a barren shale zone containing only thin calcisiltites (Lewiston C). This barren unit, in turn, gives way upward to bryozoan rich mudstones and finally, bryozoan pack- and grainstones (Lewiston D and E; Fig. 14). These facies alternations are interpreted as defining small scale cycles (parasequences).

The overlying Lewiston E limestone appears to be a weak signature of a relative sea-level fall and subsequent transgression. In short, a relative sea level drop followed by a rather rapid sea level rise is envisioned for this portion of the section (Fig. 2). Although no erosion or even sharp surfaces are seen at the base of the Lewiston E bed at Niagara Gorge, as this bed is traced westward onto the Algonquin Arch, it shows an increased sharpness to its base and, eventually, an erosion surface beneath the Lewiston E truncates most of the beds of the lower Lewiston. The upper surface of the Lewiston is sharp and locally glauconitic, and shows an abrupt change to the barren dark gray Burleigh Hill shales. This contact, again, is interpreted as a surface of relative sediment starvation, or marine flooding surface, associated with increased rates of sea level rise.

The Burleigh Hill shales are largely barren of fossils and show relatively few interbeds, except in the upper third, where the unit becomes silty and contains some thin fossil stringers (Fig. 7). The member on the whole is considered to be an upward shallowing or regressive cycle (highstand deposit). As such, it can be interpreted as representing highstand conditions with minor progradation of carbonate silt toward the end of Rochester Shale deposition. The contact with the overlying DeCew Dolostone may locally appear sharp or gradational. A minor diastem is likely and subsequence boundary is placed at the bottom of the DeCew dolostone, but no major sequence bounding unconformity occurs at this surface.

The DeCew is an interesting 9 m thick, blocky dolostone which shows spectacular soft sediment deformation. The deformation may be interpreted as a response to single event, possibly a seismite associated with early pulses of tectonic activity during the onset of the Salinic Orogeny in eastern New York. Shaley beds of the DeCew are present locally along Niagara Gorge but have been removed in most areas seen on this trip.

The base of the Gasport crinoidal grainstone also seen at STOP 2B, is similar to the base of the Model City grainstone in that it contains rip-up clasts of dolomitic, argillaceous dolostone derived from the underlying, finer grained unit. To the west, this erosion surface removes the DeCew entirely in the vicinity of Hamilton, Ontario and further cuts out the upper beds of the Rochester Shale northwest of Hamilton.

The lower Gasport Gothic Hill Member shows spectacular small scale trough and planar cross stratification. Some of this bedding appears to be bipolar, possibly indicating the influence of tidal currents. Like the similar Irondequoit grainstones, the lower Gasport grainstone is interpreted as an amalgamated crinoidal shoal deposit consisting of multiply reworked pelmatozoan sand and gravel. No Pekin Member is present here. Instead, a 2 meter thick vuggy, stromatoporoid-bearing dolostone (basal Niagara Falls Member A) of the Goat Island Formation rests directly on the Gothic Hill (lower Gasport) grainstone. Overlying beds, largely inaccessible, consist of bluff dolostone and tan cherty dolostones (Ancaster Member).

Retrace route back to vehicles. Turn vehicles around, proceed to intersection of Old Lewiston Road and Irving Drive. Turn left.

47.5	0.5	Light at intersection of Route 104. Turn left (south).
47.85	0.35	Pass by forebay area at tops of conduits for water inflow to generator turbines.
48.2	0.35	Robert Moses Power Vista.
48.5	0.3	Junction Hyde Park Blvd. (Rt. 62) at entrance to Niagara University. Turn left off Route 104 on Rt. 62.
48.75	0.25	Entrance to south haul road access to Robert Moses Power Plant and fishermen's access to Niagara River. Turn right, cross RR track.
48.75	0.25	Parking area for fisherman on left. Pull in and park. We will proceed on foot down the south haul access road for about 0.3 miles to the fishing area near the bottom of the gorge. **Please note: access to this section must be obtained from the Power Authority; hard hats are required and collecting is not permitted.**

STOP 6: SOUTH HAUL ROAD TO ROBERT MOSES POWER PLANT

This outstanding outcrop, arguably the most complete Silurian section in New York State, provides an opportunity to summarize and review all of the points of the preceding trip.

The section begins with about 8 m of the upper Ordovician Queenston Shale. Its sharp contact with the overlying Whirlpool Sandstone is the Silurian sequence I boundary. The Whirlpool, here as at Lockport (Stop 1), is represented by channel facies which record a nonmarine to marine transition. Minor phosphatic material occurs at the upper contact (starvation surface, or marine flooding surface) with the Power Glen.

The Power Glen is both thicker (8 m) and shalier at this locality than the section seen at stop 1. It is entirely gray shale, lacking the upper reddish sandy zone seen at Lockport.

As at Lockport, the Power Glen is overlain by a 2 meter thick massive white quartzose sandstone (Devil's Hole Member of Duke et al., in press), the base of subsequence IB of the Medina Group. Overlying (lower Grimsby) shales here are green and fossiliferous, in contrast to the red, ferruginous, lingulid shales seen at Lockport. The locality represents a more offshore section.

At the north end of Niagara Gorge a distinct, meter-thick phosphatic sandy dolostone, the Artpark Phosphate Bed, occurs in this position. Here, only minor phosphatic sandstone overlies the "Devil's Hole" Sandstone. These beds and the ferruginous zone at Lockport constitute a condensed section associated with rapid sea-level rise.

The main body of the Grimsby Formation (about 15 m thick) is red shale with thin sandstones. Three or four small-scale shallowing upward shale to sandstone cycles have been recognized by Duke et al. (1987) in this interval. This section is one of the most shale-rich exposures of the Grimsby in Niagara County. Sections both north and south of this locality contain more channel sandstones.

The upper beds of the Medina Group comprise about 2 meters of white, crossbedded Thorold Sandstone. The Thorold has a sharp erosive base which marks the base of the next Medina subsequence. Overlying Cambria Shale, seen at stops 2 and 5, is completely missing here. The sandy phosphatic Densmore Creek Bed rests sharply on the Thorold, marking the sequence I/II boundary.

The Neahga Shale is poorly exposed but is about the same thickness as at stop 5 on the north side of the power plant. The two phosphatic beds of the basal Reynales Limestone and the nodular capping bed are nearly the same as in the other sections seen in Niagara County. A prominent black phosphatic staining occurs on the upper surface of the Reynales Limestone (see near one of the protective cages on the outcrop). This stained surface is the sequence II/IV contact.

The Rockway/Model City (sequence IV/V boundary) is also well displayed. No bioherms occur here at the upper contact of the Irondequoit Limestone with the Rochester Shale, but a 30 centimeter thick shell bed marks the condensed zone.

The lower Rochester Shale is partly obscured by a protective cage (toxic chemical seep?), but upper limestone beds of the Lewiston Member are well exposed. Overall the Lewiston Member is somewhat less fossiliferous and contains fewer limestone beds than at stop 5, just on the north side of the power plant. This loss of interbeds reflects the beginning of a major facies change into sparsely fossiliferous Rochester Shale southward along Niagara Gorge (see Brett, 1982).

The upper Rochester displays a sharp contact with the enterolithic DeCew Dolostone beds, herein interpreted as the basal unit of a partially eroded upper Clinton subsequence. The sequence V/VI boundary at the DeCew/Gasport contact is well exposed near the entrance to the "tunnel" beneath the Robert Moses Parkway at the top of the cut. Weathered surfaces of the Gothic Hill grainstones display probable bipolar cross-stratification. In contrast to the cut on the north side of the power plant (stop 5) some 2.5 meters of argillaceous upper Gasport (Pekin Member) strata are present here. These beds display a sharp undulatory contact with overlying vuggy dolostones of the Niagara Falls Member of the Goat Island Formation. The base of the Niagara Falls Member is a subsequence boundary which erosionally bevels the 2.5 meters of Pekin Member strata between the two sides of the power plant.

As a whole, the Gasport Formation comprises a complete subsequence. The basal crinoidal grainstones of the Gothic Hill Member represent transgressive systems tract deposits, the contact with the Pekin Member is a marine flooding surface and the Pekin muddy carbonates represent relative highstand deposits (Fig. 15). This subsequence is analogous to the Model City - lower Rochester Shale subsequence.

The dark, argillaceous dolostones of the upper Gasport Formation are sharply and erosionally overlain by basal medium-bedded dolostones of the Goat Island Formation (Fig. 16). This erosionl contact is the same one observed at Pekin Roadcut (Stop 3). These dolostones, in turn, are overlain by massive, dolomitic, crinoidal grainstones appear at the type section of the Goat Island Formation at the brink of Niagara Falls. The south haul road section is a reference section for the lower portion of Goat Island Formation (Zenger, 1965). The unit is crinoidal and contains scattered <u>Cladopora</u> corals and stromatoporoids. Upper beds of the Niagara Falls Member display abundant vugs which appear to be solution cavities in a stromatoporoid-rich zone. This layer is locally highly biostromal, as noted at Niagara Stone quarry (STOP 4).

The Ancaster Member is poorly developed compared to that seen at Niagara Stone quarry (STOP 4), thin (~2.5 m) and only sparingly cherty. However, in the cliffs immediately north of the Robert Moses Power Plant, this member is developed in its typical thick cherty phase. There appears to be a complimentary relationship, in terms of thickness, between the Niagara Falls Member and the overlying Ancaster Member; i.e. where the Niagara Falls Member is thick, the Ancaster is thin to nonexistent and vice versa. Argillaceous gray dolostones of the Vinemount Member form the uppermost unit on the access road.

The Goat Island Formation as a whole here appears to be a "twin cycle" to the Gasport. The lower or Niagara Falls grainstone member constitutes the transgressive portion; its sharp boundary with the overlying Ancaster Member is a surface of starvation, sometimes recorded in the relatively glauconitic cherty beds of the lower Ancaster (Fig. 16). Ancaster and Vinemount Members represent highstand deposits.

Reboard vehicles and retrace route to road.

49.25	0.25	Junction Hyde Park Blvd. Turn left (north).
45.5	0.25	Junction Route 104. Turn right (north).
50.5	1.0	Exit for Route I-190. South to Buffalo; turn right and proceed south on I-190 to the Thruway and return to Fredonia.

END FIELD TRIP

EURYPTERID BIOFACIES OF THE SILURIAN-DEVONIAN EVAPORITE SEQUENCE: NIAGARA PENINSULA, ONTARIO, CANADA AND NEW YORK

Samuel J. Ciurca, Jr.
Rochester, New York 14621

For the past 28 years, I have been studying eurypterids, their stratigraphic occurrence, distribution, and associated flora and fauna, particularly in northeastern United States and Ontario, Canada. In New York the best known eurypterid assemblages are: 1) the *Eurypterus remipes* Fauna of the Fiddlers Green Formation that includes the earliest discovery of an eurypterid on this planet (by Dr. Mitchell in 1818); 2) the younger *Eurypterus lacustris* Fauna of the Williamsville Formation of western New York and Ontario, Canada; and 3) the eurypterid fauna of the Shawangunk Formation southeastern New York.

The search for new eurypterid faunas and the study of their geographic distribution in the Siluro-Devonian rocks of the region is important for several reasons. Foremost is the fact that the eurypterid faunas are punctuated within the known stratigraphic sequences. New genera appear suddenly within the sequences, e.g. *Waeringopterus Zone*, *Erieopterus Zone*, etc. When an eurypterid species disappears from the record completely, it is easily understood and we often assume extinction. The sudden appearance (or ecological replacement) of a species by an unrelated forms, (such as the Silurian *Eurypterus* being replaced by *Erieopterus* in the Early Devonian, is much more difficult to decipher. Only continued search for intervening occurrences, stratigraphically and geographically, will provide the data needed to understand the evolution of the eurypterids and their assemblages.

A second factor of significance concerns the eurypterid assemblages. Generally, where eurypterids are found there is a conspicuous absence of other fossils, in number and especially in diversity. Hence, the stratigraphic sequence is often considered to be unfossiliferous. The evaporite-bearing Salina Group is notable of this designation. Precise correlation, regionally and inter-regionally is needed to achieve a more complete understanding of these unique invertebrates. See Figure 1 for a map showing areas having exposures of Late Silurian-Early Devonian strata.

LOCKPORT GROUP

Eurypterids have only rarely been reported from the Lockport Group. Several reports in the literature are misidentifications, usually of the hard parts of other organisms, particularly phyllocarids.

Tylopterella boylei, reportedly from the Guelph Formation near Elora, Ontario, is the only fully known eurypterid from the Lockport Group of Canada. See Clarke and Ruedemann (1912, p. 216) for a full description.

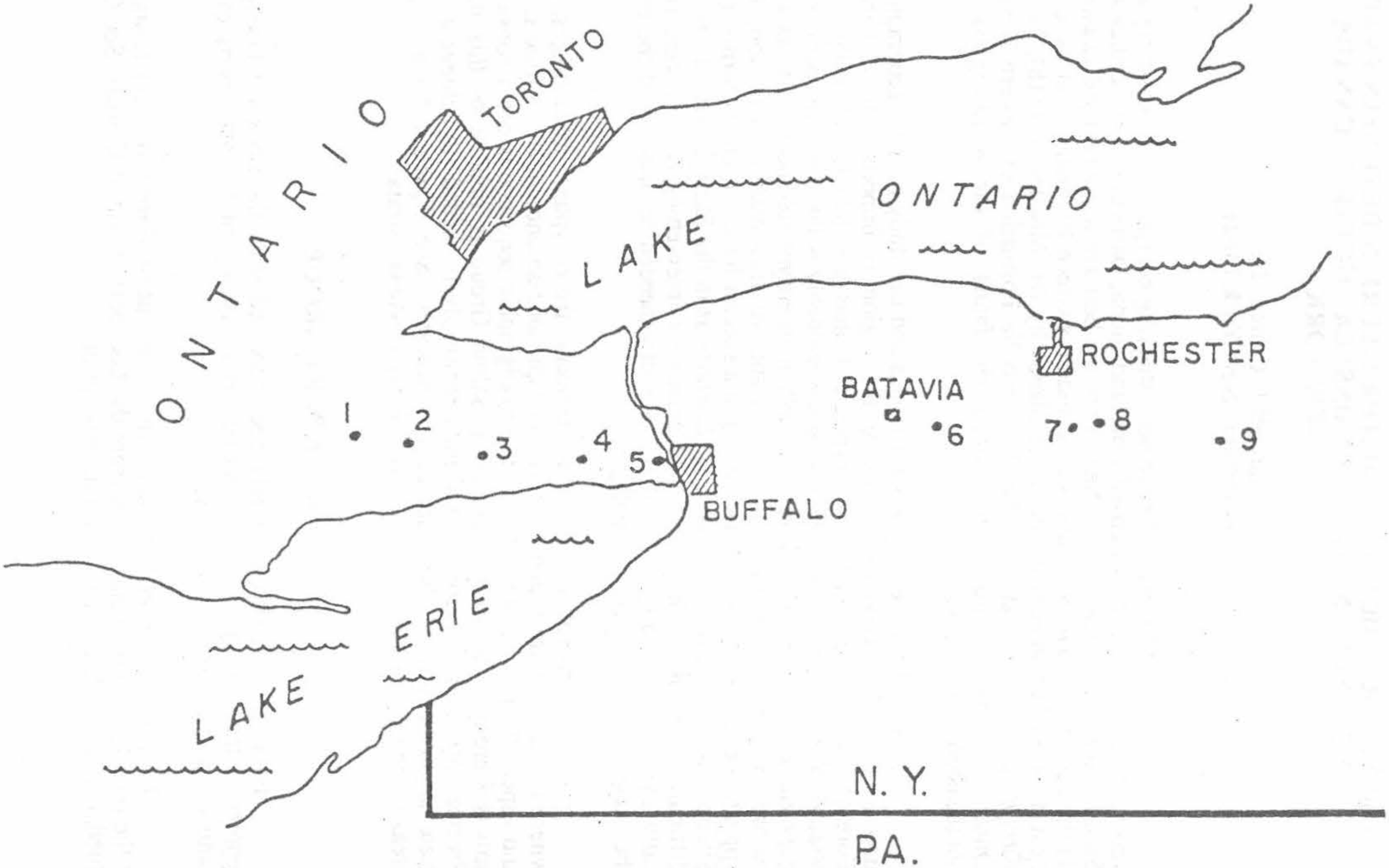


Figure 1. Areas (1-9) have exposures of Late Silurian-Early Devonian strata:
1) Haggersville 2) Cayuga 3) Dunnville 4) Port Colburne 5) Fort Erie
6) LeRoy 7) Honeoye Falls 8) East Victor 9) Phelps

(From Ciurca, 1982)

Eurypterus logani (Williams, 1915) is not an eurypterid. An examination of his plates reveals that all of the structures illustrated pertain to a phyllocarid (Ciurca, 1989). The specimens are tentatively referred to *?Ceratiocaris logani*. This is further supported by the discovery of many additional phyllocarid remains at numerous sites from the Niagara River on the Ontario side, westward to the Eramosa River.

It is not surprising that few eurypterid remains have been reported from typical Lockport facies. The lower Lockport Group could be termed a coral-stromatoporoid phase (Decew/Gasport/Goat Island) and eurypterids are not usually found in the various facies associated with these forms.

Much of the Lockport Group, however, belongs to the stromatolite phase and it is expected that eurypterid remains would be associated with some of the facies accompanying stromatolite development. This interpretation is based on occurrences of eurypterid faunas in the Bertie and Manlius Groups (Ciurca, 1978; 1990, this paper).

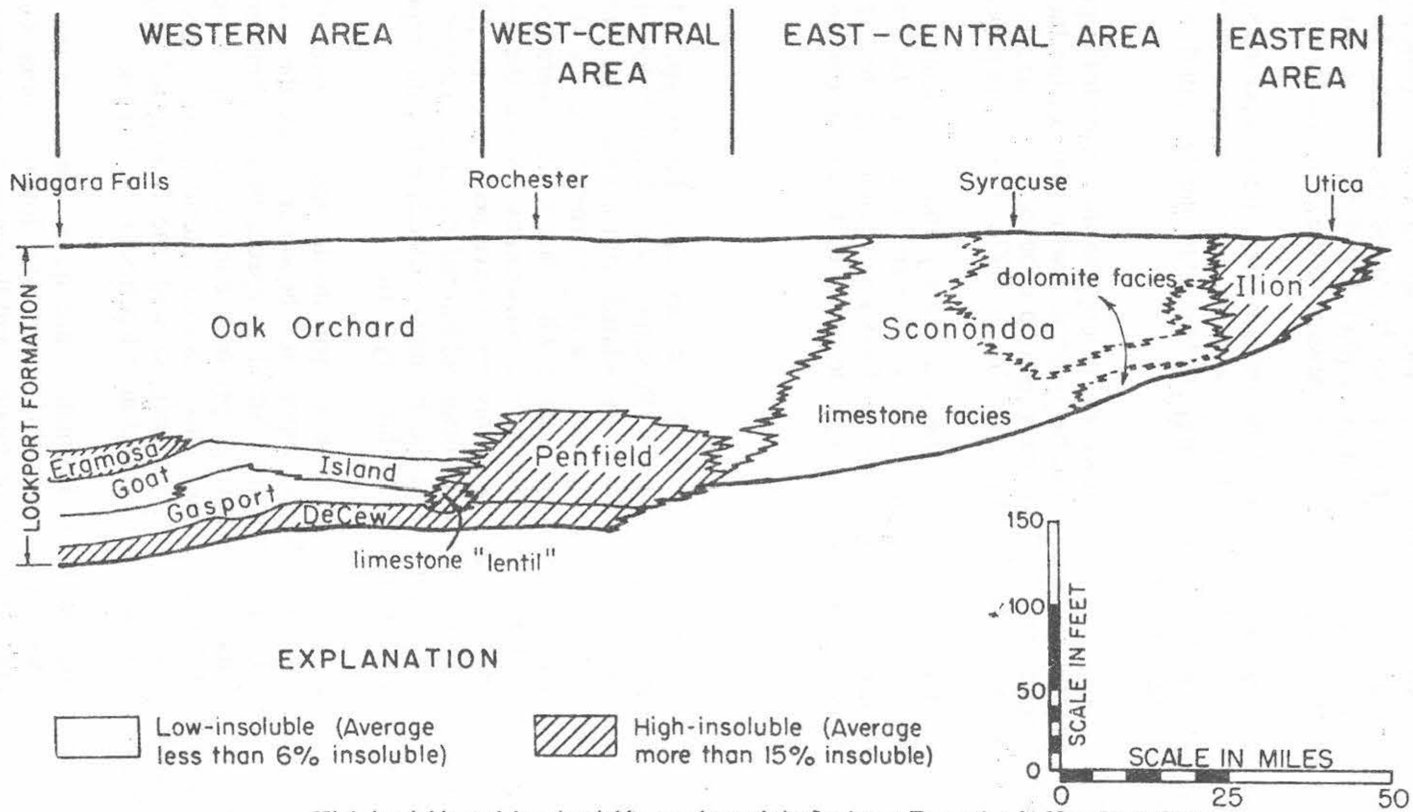
The following discussion utilizes the stratigraphy and nomenclature of Zenger (1965) shown in Figure 2. Work in progress by Brett and associates (at University of Rochester) will undoubtedly lead to a revision of Lockport nomenclature and the eurypterid occurrences discussed below will have to be modified to correspond with the new stratigraphic interpretations.

ILLION FORMATION

Clarke and Ruedemann (1912, p. 420) described a peculiar eurypterid from eastern New York, *Rhinocarcinosoma vaningeni*. Specimens were found below the red Vernon Shale in a large "concretionary block" associated with *Lingula* and *Orbiculoidea*. The "concretionary block" is now recognized as a large stromatolite and it is obvious that *Rhinocarcinosoma* lived in an environment in which black muds were accumulating among and behind stromatolitic or thrombolitic biostromes. Since the original discovery, no further specimens pertaining to this genus have been reported. Dale (1953) made an extensive search, but failed to locate additional material. Zenger (1965, p. 104) failed to find eurypterid remains at the Farmers Mills Site but did report telsons and scalelike markings that may pertain to this eurypterid.

I have located *Rhinocarcinosoma* in a stromatolite bed at Moyer Creek, but remains are exceedingly scarce. More important, however, was the discovery of another eurypterid in the Illion Formation. Small carapaces of a hughmilleriid, probably *Parahughmilleria* (Kjellesvig-Waering, personal communication), were found at Farmers Mills and also at Moyer Creek. This small eurypterid is easily recognized by its semicircular carapace and cannot be confused with the "common" *Hughmilleria socialis* so well known from the Pittsford Bed near Rochester, New York.

The dark shales of the Illion Formation are quite similar to the "Pittsford Facies" originally described by Sarle (1903) from the lower Salina Group near Rochester, New York. Large stromatolitic mounds and thrombolite horizons are intercalated with the otherwise mudstone and shale that constitute most of the Illion Formation.



High-insoluble and low-insoluble members of the Lockport Formation in New York State.

Figure 2. Stratigraphy of the Lockport Group (from Zenger, 1965)

Spectacularly, the upper Illion Formation, transitional to the overlying Vernon Formation, harbors a rich eurypterid fauna identical to the lithofacies and biofacies described for the eurypterid beds in the type region of the Pittsford Bed (see Sarle, 1903 and Ciurca, 1985a).

Ciurca (1986a) has termed the new eurypterid bed the Farmers Mills Bed for the occurrence along Oriskany Creek at Farmers Mills, ironically the locality which provided the first *Rhinocarcinosoma* remains described by Clarke and Ruedemann (1912), and has transferred the unit to the Salina Group based on lithological characteristics. This reassignment may help us to realize that the eurypterid-bearing black shales of the Salina Group are facies of portions of the Lockport Group. Zenger (1965) has already suggested a similar relationship for the Clinton and Lockport Groups in the region.

It may be that the entire Illion Formation should be included in the Salina Group. It is known that the Illion Formation interfingers with the Sconondoa Formation to the west, supporting the thesis that the eurypterid beds of the east are correlative with the beds westward that are lithologically part of the Lockport Group. There is also paleontological support for this. (see discussion of the Sconondoa and Oak Orchard Formations).

SCONONDOA FORMATION

Zenger (1962; 1965, p. 88-89) described the Sconondoa Formation of central-eastern New York as "brownish grey to dark grey, aphanitic, thin to medium-bedded and bituminous" limestone. Zenger noted that stromatolites and associated edgewise conglomerates are abundant east of Baldwinsville, especially at Sconondoa Creek in the Oneida Quadrangle.

At the Sodus Quarry, about 70 miles west of Sconondoa Creek, a similar facies in the floor of the north quarry was discovered (Ciurca and Domagala, 1988). This horizon apparently was not exposed at the time Zenger studied this section (Zenger's Locality No. 17).

A stromatolite horizon occurs near the base of the section, with over 60 feet of "Oak Orchard" above. The stromatolite zone may actually consist of four separate horizons with intercalated ripplemarked black shales and dolomitic beds.

Below the stromatolite beds are variable beds which could best be described as the *Chondrites* facies. The beds consist of fine-grained limestone with conchoidal fracture, apparently argillaceous limestone with sparry interlayers bearing an interesting fauna of brachiopods, gastropods (some unusually large), cephalopods, ostracods and trilobites. The most exciting find was an eurypterid which corresponds to the *Rhinocarcinosoma vaningeni* of the eastern facies, ie. Illion Formation. If this analysis is correct, the occurrence at the Sodus Quarry extends the known geographic distribution of *Rhinocarcinosoma* substantially. This unusual eurypterid was also recently discovered in the McKenzie Formation in Pennsylvania (Ciurca, 1978) in a facies not unlike that of the Sconondoa and Illion Formations of central-eastern New York.

Currently, the exact relationship of the Sodus Quarry eurypterid/stromatolite horizons to those farther east is unknown. I suspect we will find that stromatolitic beds occur still lower in the section in an eastward direction.

OAK ORCHARD FORMATION

Most of the Lockport Group in western New York consists of massive dolostones, some biostromal and vuggy (mineraliferous) units of thin and medium bedded, fine to coarse-grained dolostone and much of this is described as bituminous. Crystalline vugs are common and different horizons or beds carry distinct mineral suites (see Ciurca, 1989; 1990).

The uppermost Lockport Group at Rochester, eg. the beds at Allens Creek at New York Route 65, consists of fine-grained thin and medium-bedded dolostone containing a small fauna; *Howellella*, pelecypods and eurypterids, chiefly *Eurypterus pittsfordensis*. Interbeds consist of low stromatolitic mounds and oolitic beds with occasional *Favosites*. Intraformational conglomerates are present and are undoubtedly associated with the algal phase. Such conglomeratic horizons are common in the Illion Formation to the east and in the algal phase in the Bertie and Manlius Groups.

Re-excavation of uppermost Lockport rocks during July of 1989 at the eurypterid site at North Chili (NY 33 and 259) has provided additional data on the occurrence of *Eurypterus pittsfordensis*. The lithologies present are more indicative of the Lockport Group than of the Salina Group, but the sequence is still best described as transitional. There is little indication of the greenish shales so typical of parts of the Salina Group at the eurypterid sites near Pittsford, New York and no red shale is present. Much of the rock is dolomitic, some resistant and more typical of Lockport lithologies, and some is dolomitic mudstone that weathers more readily. *Eurypterus pittsfordensis* is common along with *Lingula* and pelecypods. *Chondrites* is very important here and, like the occurrence at the Sodus Quarry (see Sconondoa Fm.), this section can best be described as belonging to the *Chondrites* facies. Three large algal mounds were observed and two of these were recovered (one has been presented to the Buffalo Museum of Science). The North Chili Site represents the most westward occurrence of *Eurypterus pittsfordensis* currently known (Ciurca, 1990, in press).

Along the Niagara River a portion of an eurypterid was discovered in uppermost Lockport rocks or transitional Salina beds (Ciurca, 1989). The most likely source is the interval described by Zenger (1965, Loc. 7, p. 167) at the Intake Towers of the Niagara Power Project. Unfortunately, the uppermost Lockport Group is not exposed anywhere in the region and only future excavations and cores can provide data for this interval.

As increasingly observed for the facies, the Niagara Falls eurypterid is associated with a sequence bearing numerous stromatolite horizons and possible salt hoppers. The large-scale stromatolites described by Zenger (1965, p. 114, Fig. 22) are still visible at several points along the Niagara River Gorge. An analogous occurrence in Ohio is worth mentioning. At the Maumee Quarry east of Toledo, favositid "reefs" are overlain by stromatolites before giving way to the overlying Greenfield Formation. Within this stromatolitic sequence occurs *Paracarcinosoma*, a genus prominently known from the Williamsville Waterlime at Buffalo, New York. This is the first eurypterid to be reported from this quarry (Ciurca, in preparation).

SALINA GROUP

The Salina Group represents an extraordinary sequence of red beds, fine-grained limestone and dolostones and extensive beds of evaporites, mostly anhydrite/gypsum and halite. Though generally unfossiliferous, certain beds within the group have provided faunas rich in eurypterids and other peculiar forms. Recent efforts have been directed at identifying more precisely the known eurypterid horizons and at locating additional horizons not only within the Salina Group of New York, but also within the wonderful stratigraphic sequences displayed in Ontario, Canada, Pennsylvania and other parts of the Appalachian Basin.

The following discussion is limited to recent observations of the stratigraphic position and geographic distribution of eurypterid horizons in the Salina facies. Much of the Salina Group in New York is detrital, and the eurypterid biofacies possibly reflect, at least in part, faunas differing from those typical of the Bertie Group that overlies this sequence.

VERNON FORMATION

The type section for the Vernon Formation is at Vernon, New York. (See Alling (1928, p. 23) for a vivid description of the Vernon and its relation to the salt beds. Most of the Vernon is red shale and mudstone. However, several other lithofacies are present: green and black shales, resistant waterlimes, thin dolostone beds, evaporites, and even sandstones (presumably tongues of the Bloomsburg Formation of Pennsylvania or the Shawangunk Formation of southeastern New York.

In thin (0.25m-1.0m) green, black and greenish-black beds, mostly within the lower Vernon Formation, occur a variety of eurypterids, other arthropods (*Pseudoniscus*, phyllocarids) and usually *Lingula*, cephalopods, pelecypods, and, rarely, brachiopods. Most of the eurypterid-bearing units currently known are located in the Rochester, New York area, the type area for the"Pittsford Shale" of Sarle (1903). The following beds have recently been proposed but are not described below in stratigraphic order.

Pittsford Bed

Type Locality: Covered sequence behind and near the Spring House, Monroe Avenue, Pittsford, New York.

When Sarle described the "Pittsford Shale" and its fauna, the unit (reasonably) included the eurypterid beds and all the strata below to the contact with the Lockport dolostones. Fisher (1960) and Rickard (1975) thought that the Pittsford was simply a local facies of the Vernon Formation and suggested the unit be abandoned. Ciurca (1985a), in contrast, suggested that, even though of possible limited extent (not much is known about the Lockport/Vernon contact throughout its extent), the "Pittsford Shale" was a lithologic unit bearing an exceptional eurypterid-bearing interval and that the unit be termed the Pittsford Bed. Supporting this was the recognition, in the region, of other distinct eurypterid-bearing horizons separate from the type Pittsford Bed by other lithofacies, mostly red and green shales and mudstones. The Pittsford Bed, thoroughly described by Sarle, bears *Hughmilleria socialis* in enormous numbers. Also present are *Eurypterus pittsfordensis*, *Pterygotus monroensis* and many other species.

My observations of the Pittsford Bed, exposed during excavations for an enlarged Wegmans/Chase-Pitkins plaza and other nearby projects, indicate that the depositional site dried up (subjected to subaerial exposure) leaving a mudcracked interval before the deposition of overlying waterlimes, shales and mudstones. The Pittsford Bed provides the best impression of eurypterids being trapped in an evaporating body of water. Nothing, of course, is known about their potential for surviving dry spells analogously to many modern forms, especially fish, that survive by burying themselves in the muds until the next inundation. While salinity may have been high during deposition of the Pittsford Bed, the evidence has not been observed in this unit as it has in the stratigraphically higher Barge Canal Bed (see Ciurca, 1985b).

Barge Canal Bed

Type Section: Barge Canal just west of village of Pittsford, N.Y.

The Barge Canal Bed (Ciurca, 1985b) was named for the black shale described by Ruedemann (1919) and containing predominately the *Eurypterus pittsfordensis* Fauna. The outcrop in the Barge Canal was rediscovered by the author during exploration for available exposures of the Vernon Formation in the region. The Barge Canal Bed is still exposed when the canal is drained during winter, and consists of black shale grading upward into more greenish to bluish-weathering shale. In this unit, well developed salt hoppers were observed for the first time this low in the Salina Group, and suggest that all of the eurypterid horizons experienced salinity crises.

This stratigraphic position in the area is easily established, but the exact thickness of Salina beds occurring between the Pittsford Bed below, and the Barge Canal Bed above, is unknown. About 20 feet of strata were observed overlying the Pittsford Bed at the I-590 Site with no evidence of the Barge Canal Bed. Thus the intervening beds are over 20 feet in thickness and probably less than 60 feet. The most important faunal element of the Barge Canal Bed is *Eurypterus pittsfordensis*. *Hughmilleria socialis* appears to be excluded. Rare *Mixopterus* remains have been found, however the geographic extent of the Barge Canal Bed is unknown. I have suggested that the Gananda Site (Hamell, 1978) probably is equivalent (Ciurca, 1984).

Monroeav Bed

Type locality: I-590 just south of Monroe Avenue, Rochester, New York.

The discovery of an even lower eurypterid bed, i.e. relative to the Pittsford Bed, prompted the redefinition of Sarle's "Pittsford Shale", and the naming of newly discovered and previously described eurypterid-bearing units (Ciurca, 1984).

About 15 feet below the Pittsford Bed occurs about 7 feet of greenish dolomitic shale and mudstone replete with *Lingula*. The interval corresponds to the 7 feet of section described in Clarke and Ruedemann (1912, p. 103). To this shaly interval I have given the name Monroeav Bed (Monroe Avenue). While the lower contact was not observed, it seems likely that the Monroeav Bed rests directly upon the Lockport Group.

Eurypterus pittsfordensis is the only eurypterid obtained from this unit. No correlatives are known, but the North Chili Site may be a possible correlative, even though the facies at North Chili appears to be more Lockport-like in dolomitic character. The Monroeav, Pittsford, and Barge Canal Beds are all located within a small geographic area at Pittsford, New York, and their relative position is easily determined. See description of Harris Hill Bed that follows for additional comments.

Harris Hill Bed

Type Section: Wegmans parking lot, south side of NY 441, south of Harris Hill, East Penfield.

North and east of the type Pittsford Bed, another eurypterid occurrence was found that appears not to be related to the previously described units. Temporary excavation for construction of Wegmans Supermarket revealed a thin section of fine-grained dolostones, red, green, and black shales. For the thin (about 0.3m) black shale I have proposed the name Harris Hill Bed (1985b). The Harris Hill Bed contains abundant *Hughmilleria socialis* and rare *Pterygotus*, and *Mixopterus*.

The Harris Hill Bed is the thinnest of the beds described and it is possible that it is simply the thinned equivalent of one of the previously described beds. However, if the inclination of the bedrock in the area is relatively normal, calculations indicate the possibility the Harris Hill Bed is even lower, stratigraphically, than the Monroeav Bed. The Vernon red beds thicken dramatically eastward (about 600 feet at Syracuse) and it is possible that some of this increase in thickness is picked up at its base (see Figure 3). Still higher eurypterid occurrences within the Vernon Formation compound the problem of age relationships between the various eurypterid horizons.

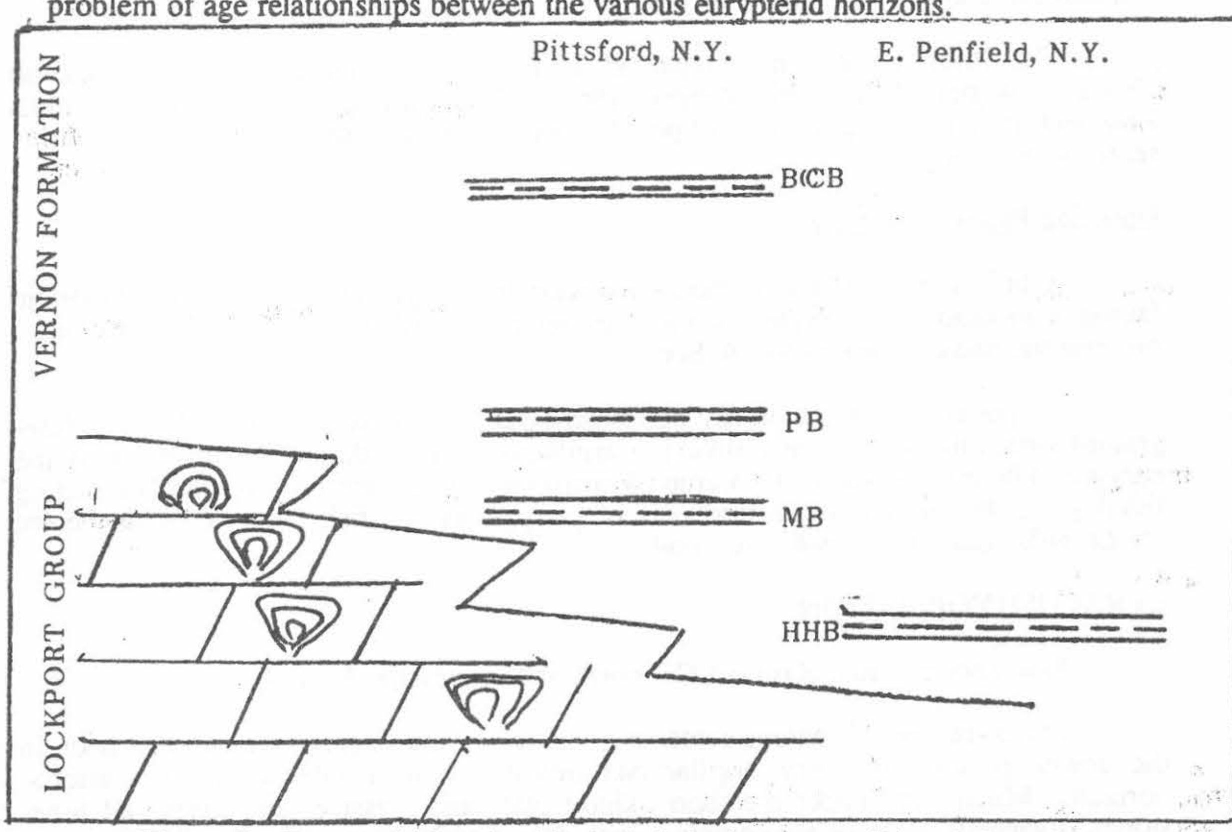


Figure 3. Eurypertid-bearing black shale horizons: possible relationship to the Lockport Group Below. HHB = Harris Hill Bed, MB = Monroeav Bed, PB = Pittsford Bed, BCB = Barge Canal Bed.

Farmers Mills Bed

Type Section: Oriskany Creek and tributary at Farmers Mills.

In eastern New York the Illion Shale underlies the Vernon Formation. Transitional shales have been described as Farmers Mills Bed by Ciurca (1986a). This portion of the Illion Formation has been assigned to the Salina Group on lithological grounds. The Farmers Mills Bed is about one meter thick and consists of greenish and black fissile shale lithofacies so well-known from the lower Salina beds at Pittsford, New York.

Thus far, *Rhinocarcinosoma vaningeni*, described from the Illion Shale proper, has not been recognized in the Farmers Mills Bed. Other eurypterids, however, are abundant as they are in the type Pittsford Bed of western New York. The following animals have been found in the type Farmers Mills Bed: *Hughmilleria*, *Pterygotus*, *Parahughmilleria*, *Mixopterus*(?), and *Lingula*.

Correlation of the Farmers Mills Bed is currently impossible on the basis of the eurypterids alone. Microfossils, however, may provide a basis for comparison with the eurypterid horizons of western New York. The Farmers Mills Bed is currently recognized in the area between Clinton and Frankfort, New York. The occurrence at the latter locality (Moyer Creek) represents the easternmost point from which eurypterid remains are known from the Salina Group.

The Illion Formation is regarded as the eastern equivalent of the Lockport Group of western New York (Zenger, 1965). The stratigraphy of the Farmers Mills Bed, and its relationship to the Lockport-Vernon sequence in western New York, is currently under study.

Downing Brook Bed (New)

A thin interval of the Vernon Formation in the type area contains an important faunal association of eurypterids and fish remains (Fisher, 1957). For this unit I propose the name Downing Brook Bed.

Unlike all of the previously described beds, the Downing Brook Bed is a fine-grained dolomitic unit quite unlike the argillaceous units that make up most of the eurypterid-bearing strata of the Vernon Formation. Consequently, with the contrasting lithologies, the fauna consists primarily of pterygotids and fish remains. Also present are *Lingula*, gastropods, and pelecypods.

SYRACUSE FORMATION

Reference Section: Railroad Cut north of Fayatteville, New York.

The Syracuse Formation contains a variety of lithologies. Transitional beds in the lower portion are very argillaceous dolostones with intercalated stromatolite horizons. Middle and upper members exhibit platy and massive dolostones and limestones. Evaporite features are common, including cavities formed by the dissolution of gypsum and halite. See Leutze (1956, 1961, 1964) for a complete discussion of the Syracuse Formation in central-eastern New York.

Sat./Sun. DII

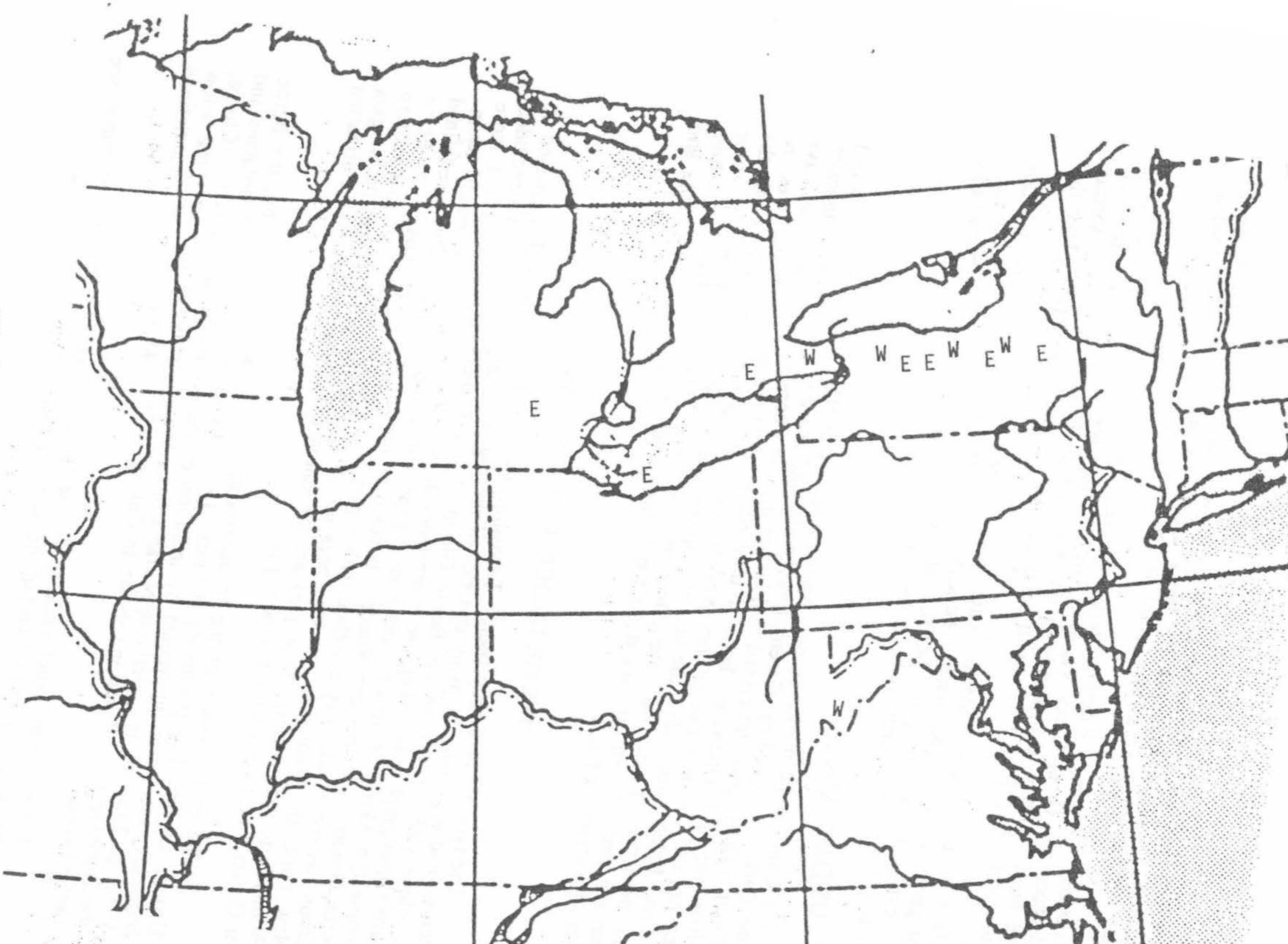


Figure 4. Eurypterid occurrences in Eastern United States and Ontario, Canada.
E *Erieopterus* W *Waeringopterus*

In western New York, the Syracuse Formation was traced from the Clyde River westward through the Oatka Creek Valley, Black Creek east of Batavia, I-990 roadcut (gypsiferous) into Ontario, Canada (Welland Canal Site).

The zonal eurypterid, *Waeringopterus*, was recovered from all of the localities mentioned except the Oatka Creek Valley (Ciurca, in preparation). *Waeringopterus* is one of the most common eurypterids in the Appalachian Basin and its widespread geographic distribution (Figure 4) should be useful in correlating strata throughout the northeast.

Many attempts were made to locate *Waeringopterus* in Pennsylvania, but this important genus has yet to be found. The section near Mt. Union, which has already provided several distinct eurypterid horizons at one locality, is a most likely site worthy of continued effort and is currently being examined.

CAMILLUS FORMATION

The Camillus Formation is most readily observed at NY 19 north of LeRoy and along the Oatka Creek Valley. Greenish mudstones and dolomitic shale constitute much of this unit. Salt hoppers and crystal cavities are present where evaporite minerals have been dissolved by groundwater. Beds of anhydrite/gypsum are known. Waterlimes are present, but little is known about their stratigraphic position and distribution within the formation. Paleontologically, the unit is impoverished. Ostracods and perhaps some poorly preserved pelecypods are uncommon at best. The Fort Hill Waterlime (lowermost Bertie Group) occurs at the top of the Camillus in western New York and carries an *Eurypterus* fauna.

BERTIE GROUP (REDEFINITION)

The type Bertie Group (Bertie Township, Ontario, Canada) is well displayed at outcrops and especially in many quarries of the region. Many authors have excluded the Akron-Cobleskill from the Bertie Group, but it is here suggested that a more refined stratigraphy, revealing the interrelationships of the various units, results by including the Akron-Cobleskill, and even higher undescribed units, within a redefined Bertie Group. Note the term Rondout is not used herein. While portions of the Bertie Group, as redefined, may correlate with a part or all of the Rondout of southeastern New York, the lithologies and sequences we observe in the region from the Niagara Peninsula to near Albany, New York are not equivalent. The term Rondout Formation or Group is best utilized in its type region.

The cyclic nature of the formations and members included within the Bertie Group, as described previously (Ciurca, 1973; 1978), ie. recurring lithofacies and biofacies, are better interpreted by inclusion of the Akron-Cobleskill facies (Ciurca, 1978, p. 233) and the overlying Moran Corner Waterlime (unnamed waterlime below the Honeoye Falls Dolostone shown in Ciurca, 1978, Figures 2-3) within a redefined Bertie Group.

The suggested redefinition shows the Moran Corner Waterlime (new name, see this paper) as the uppermost unit of the Bertie Group.

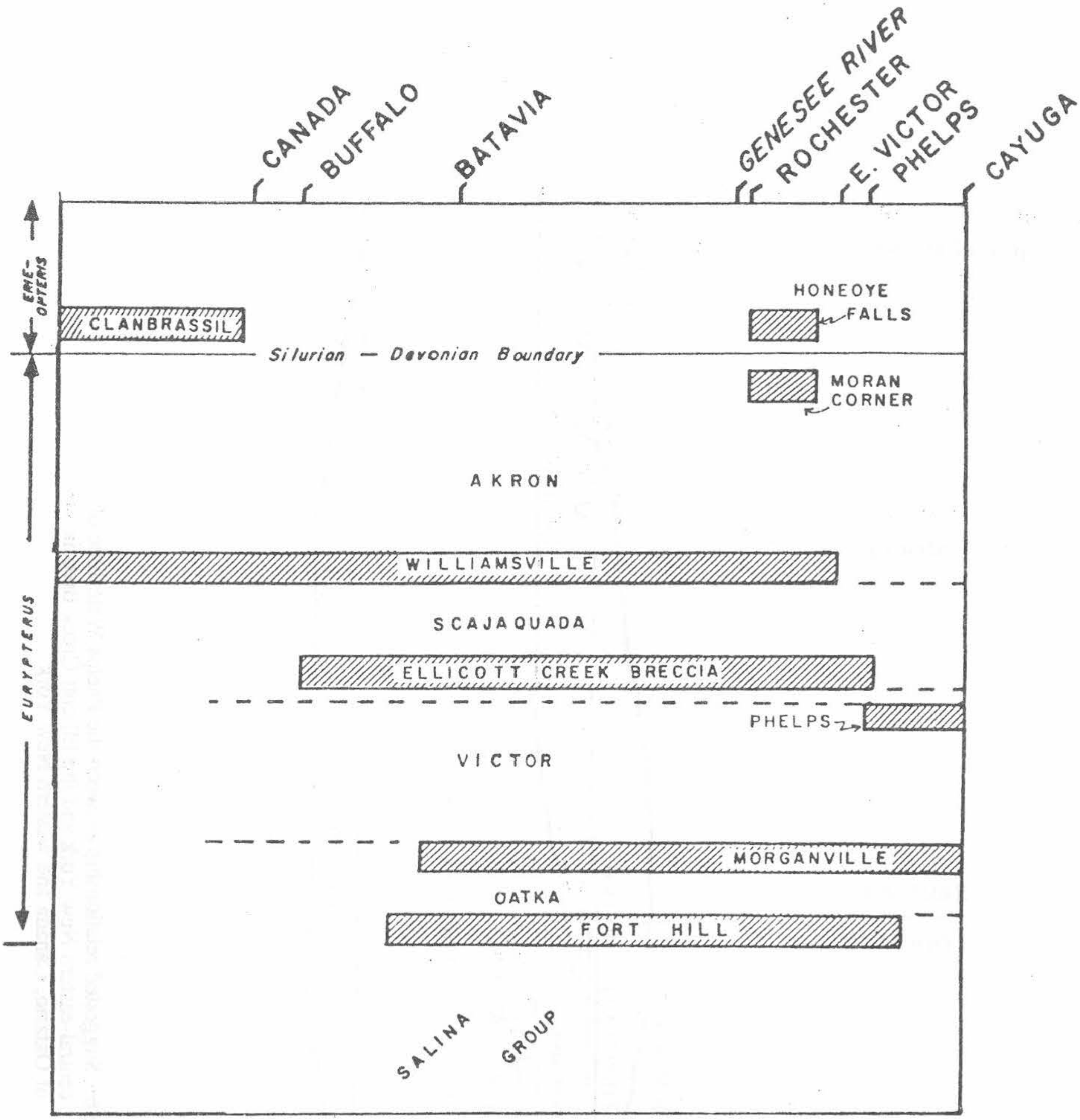


Figure 5. Distribution of the Late Silurian and Early Devonian Eurypterid-bearing Waterlimes (not to scale).

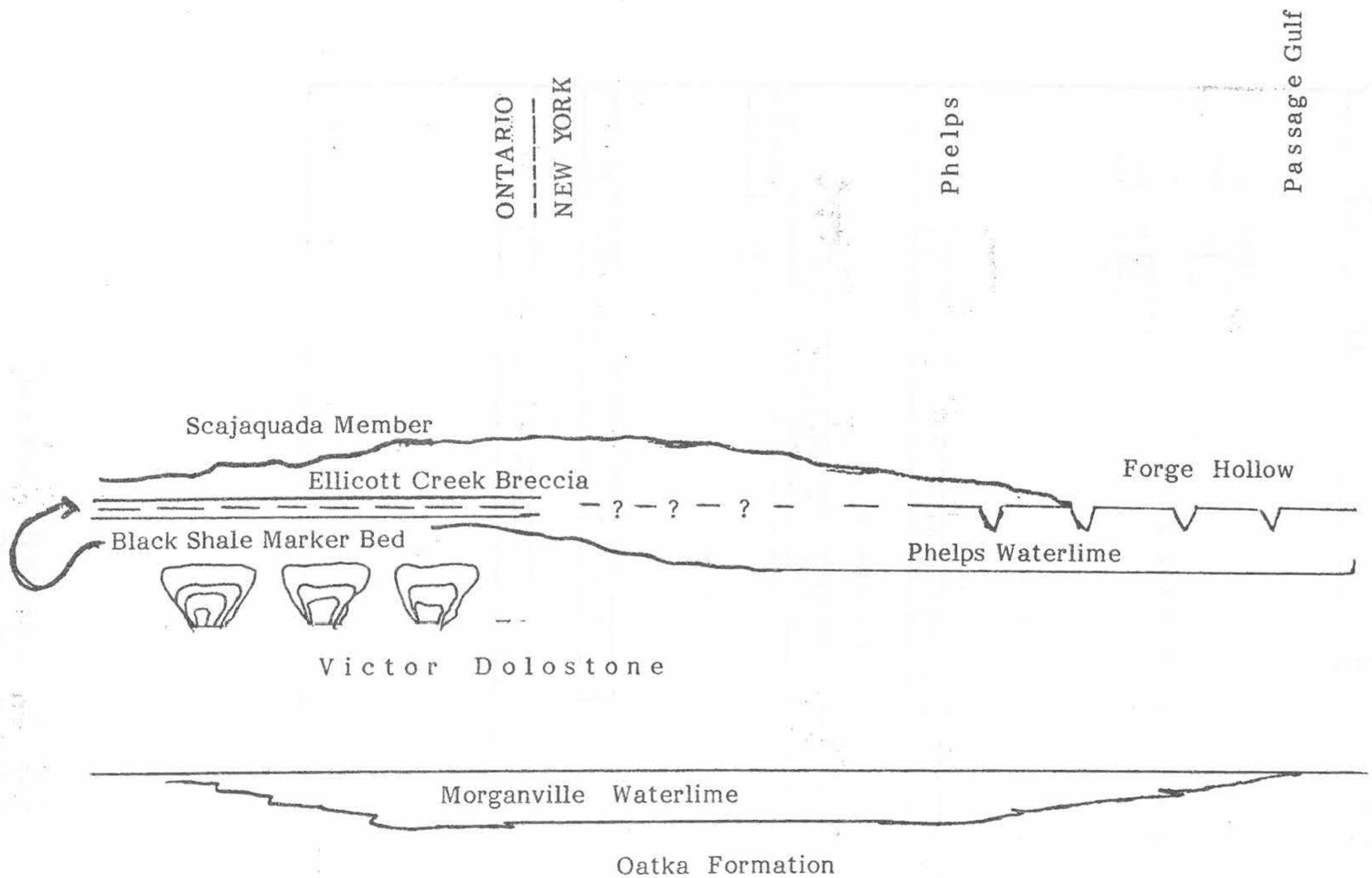


Figure 6. Suggested relationship between the Phelps Waterlime of central-eastern New York and the Ellicott Creek Breccia of Ontario, Canada and western New York.

Ontario, Canada
Bertie Group

Cobleskill Fm.
Williamsville Fm.
Scajaquada Fm.
Fiddlers Green Fm.
- Ellicott Creek Breccia
- Phelps Waterlime?
- black shale unit
- Victor Member
- Morganville Waterlime

Oatka Formation

Western New York
Bertie Group

Moran Corner Waterlime
Cobleskill Fm.
Williamsville Fm.
Scajaquada Fm.
Fiddlers Green Fm.
- Ellicott Creek Breccia
- Phelps Waterlime
- black shale unit
- Victor Member
  - C limestone
  - B dolostone
  - A limestone
- Morganville Waterlime

Oatka Formation
Fort Hill Waterlime

FIDDLERS GREEN FORMATION

Type section: Butternut Creek, north of Jamesville, New York.

In the type area the Fiddlers Green Formation is about 10 meters thick and consists of an upper eurypterid-bearing member that is believed to be continuous from the type Phelps Waterlime of western New York, east to the famous *Eurypterus* bed at Spohn Hill south of Herkimer.

The middle member is mostly dolomitic, but limestone beds are known. This portion of the Fiddlers Green Formation was termed the Victor Dolostone previously and contains a small fauna consisting of brachiopods, ostracods, stromatolites, and eurypterid remains. Throughout much of the region the lowermost Fiddlers Green Formation is a waterlime with fossils only rarely evident. This member is the Morganville Waterlime and is often straticulate and, at least at at some localities, contains salt hoppers and other evidence of hypersalinity.

On the Niagara Peninsula, a brecciated waterlime unit occurs just beneath the Scajaquada Formation. When traced into New York, the beds were found to correspond to a thicker (2.5m.) section at Ellicott Creek, Williamsville that was previously termed Ellicott Creek Breccia (see Figure 5). While the precise relationships across the area are not currently known, I have suggested that the Ellicott Creek Breccia is stratigraphically higher and younger than the Phelps Waterlime that occurs across much of the outcrop belt (east-west). A zone of mudcracks marks the top of the Phelps Waterlime throughout its extent. In Canada, a black shale occurs at the base of the Ellicott Creek Breccia and is used as a marker bed.

Black Shale Marker Bed

Peculiar platy dolomitic black shale, apparently unfossiliferous, occurs as a marker bed separating the Ellicott Creek Breccia from the main mass of the Fiddlers Green Formation below (Ciurca, 1982). I suggest this black shale be termed the Black Shale Marker Bed. This bed is prominent on the Niagara Peninsula and is useful for readily identifying the sequence of waterlimes and dolostones present. I believe the Black Shale Marker Bed is present at Ellicott Creek at Williamsville, New York where a re-entrant occurs at the base of the type Ellicott Creek Breccia and just above a brachiopod-rich waterlime of the Victor or Phelps Member of the Fiddlers Green Formation.

Maximum thickness of the black shale bed is about 15 centimeters. Thickness variation may have resulted from fluctuations in the development of the algal mounds below. Crude stromatolitic structures are also found in the middle of the overlying Ellicott Creek Breccia at several localities. Figure 6 illustrates the current concept of the relationship of western occurrences of the Ellicott Creek Breccia with that of the Phelps Waterlime to the east. For a more complete discussion of the Fiddlers Green Formation see Hopkins (1914, 1969), Rickard (1962, 1969), Ciurca (1973, 1978, 1982, 1986), Hamell (1981, 1985, 1986), qne Hamell and Ciurca (1986).

SCAJAQUADA FORMATION

Argillaceous dolostone constitutes most of the Scajaquada Formation in the Niagara Peninsula of Ontario and western New York. Most of the rock is thin-bedded and chert nodules are evident at several horizons. No fossils have ever been described from this unit except for rare eurypterid remains from a thin waterlime bed near the base in the Batavia area (NYS Thruway roadcut).

Eastward the Scajaquada Formation thickens. At Phelps (NY Rte 88 roadcut) the Onondaga Limestone and a basal sandy unit (Springvale?) overlie a very thin sequence of the Scajaquada Formation. At this locality, red stained salt-hoppers are abundant. Eastward, and presumably southward, the Scajaquada Formation is replaced by the gypsiferous Forge Hollow Formation (an arbitrary cutoff near Auburn, New York).

It is possible that part of the Scajaquada/Forge Hollow interval is correlative with the Ellicott Creek Breccia of western New York and the Niagara Peninsula of Ontario. Favorable localities for studying the Scajaquada Formation include Glen Park at Ellicott Creek, Williamsville; inactive quarry east of Clarence; Neid Road Quarry north of LeRoy; Mud Creek near Victor; and the NY 88 roadcut north of Phelps at the I-90 overpass.

WILLIAMSVILLE WATERLIME

Type Section: Ellicott Creek at Williamsville, N.Y.

The Williamsville Waterlime can readily be divided into four units at a number of localities in Ontario and western New York. The lower waterlime has been the primary supplier of specimens of *Eurypterus lacustris* and *Pterygotus cummingsi* in recent years, particularly from localities in the type area (Ridgemount Quarry) and the famous Bennett Quarry at Buffalo.

Above the lower waterlime unit (A) is a thin sequence (B) bearing brachiopods, the most important being *Eccentricosta jerseyensis* known principally from the Cobleskill Formation of central/eastern New York and from the Keyser Formation of Pennsylvania (see Ciurca, 1982, p. 113).

Williamsville C is almost a repetition of Williamsville A but little is known about the distribution of fossils within the unit. Williamsville D is a transitional unit. It is a waterlime with conchoidal fracture, large ostracods and eurypterid remains. It is slightly mottled, but not as intensely as the Cobleskill Formation above. A stylolitic contact has been observed between Williamsville D and the Cobleskill Formation at some localities. Williamsville D, in Ontario, is believed to be the source of most of the Canadian *Pseudoniscus* remains.

Some portions of the Williamsville Formation, particularly Williamsville A, are varvelike. Fresh rock, as seen mostly in the Ontario quarries, is resistant to collection of fossils. Conchoidal fractures occur in all directions. Upon weathering, however, bedding planes part and the fauna becomes evident. In Ontario, the most important species are *Eurypterus lacustris* and *Pterygotus cummingsi*. The fauna has been described previously (Ciurca, 1982), but I want to emphasize here that *Eurypterus lacustris* is a very distinctive eurypterid and is not known from any other horizon. Its occurrence thus far is limited to the region from Haggersville, Ontario eastward to Victor, New York. I have now observed this eurypterid at Manchester, New York, along Canandaigua Outlet, where it is exceedingly rare (2 specimens), being replaced by *Paracarcinosoma scorpionis* Fauna (Ciurca, 1973; Hamell and Ciurca, 1986).

COBLESKILL FORMATION

Throughout western New York and the Niagara Peninsula, the Williamsville Waterlime is overlain by massive mottled dolostone. This poorly fossiliferous sequence is about 20 feet thick and is usually referred to the Akron/Cobleskill (see Rickard, 1962 for more detail). No eurypterids are definitely known from the Cobleskill Formation of this region, but eastward facies changes occur and *Eurypterus*-bearing limestone and dolostone are encountered. A more detailed description of the Cobleskill Formation was presented earlier (Ciurca, 1982).

MORAN CORNER WATERLIME (New)

Type Section: Small creek south of Moran Corner about 10 miles south of Rochester, New York.

Reference Section: Honeoye Creek, downstream from NY 65, Honeoye Falls, New York.

An unnamed waterlime unit lying stratigraphically above the Cobleskill Formation, was recognized by Ciurca (1982, p. 101, p. 111). It is suggested that this waterlime be termed the Moran Corner Waterlime with important sections as listed above. The Moran Corner Waterlime is lithologically similar to the Phelps Waterlime of the Fiddlers Green Formation. It is about one meter thick and represents a regressive phase of the waterlime (carbonate) cycle with a mudcracked horizon at the top of this unit. Salt hoppers are present as well as an eurypterid fauna.

The Moran Corner Waterlime completes a cycle analogous to the cyclical pattern found in the Fiddlers Green Formation as illustrated below.

Moran Corner Waterlime	Phelps Waterlime
Cobleskill Formation	Victor Member
Williamsville Waterlime	Morganville Waterlime

It should be noted that subsequent waterlimes occur in Ontario, Canada and Western New York are part of the Helderbergian transgressive-cyclic sequences, contain *Erieopterus*, and are believed to be Lower Devonian in age (Gedinnian). A biostratigraphic discontinuity between the Moran Corner Waterlime and the Honeoye Falls Formation, ie. between the *Eurypterus* and *Erieopterus* Zones seems possible (Ciurca, 1967, 1982).

To the west, the Moran Corner Waterlime is unknown. In Canada the Helderbergian Clansbrassil Formation rests directly upon the residual beds of the Cobleskill Formation, apparently disconformably (Figure 5).

CLANBRASSIL FORMATION

Overlying the Cobleskill Formation in the region from Byng to Haggersville, Ontario is a sequence of very fine-grained dolostones, the Clanbrassil Formation (Ciurca, 1982). Type locality is the abandoned quarry at Clanbrassil, but wonderful exposures are available in a number of quarries at Byng, Cayuga, and Haggersville.

The Clanbrassil Formation is about 25 feet thick and is unconformably overlain by the Bois Blanc Limestone. Beneath this unit are the mottled Cobleskill (Akron) fine-grained dolostones (Figure 4).

The eurypterid, *Erieopterus*, is the sole fossil thus far obtained from the Clanbrassil Formation. It is a particularly common form and occurs primarily in the lower portion of the formation. The Clanbrassil Formation is correlative with part or all of the Manlius Group of central New York which is regarded as Gedinnian in age (See Figure 5).

SUMMARY

A variety of eurypterids are found throughout the Late Silurian-Early Devonian sequences of New York and Ontario, Canada. The importance of eurypterids for corhas been albeit neglected, I believe the more we know about their occurrences, the more we will find them useful in relating sequences that are geographically far apart.

Waeringopterus is turning out to be of widespread occurrence and I expect that, sooner or later, it will show up in the Silurian rocks of Pennsylvania. It forms a relatively narrow zone near the base of the Syracuse Formation.

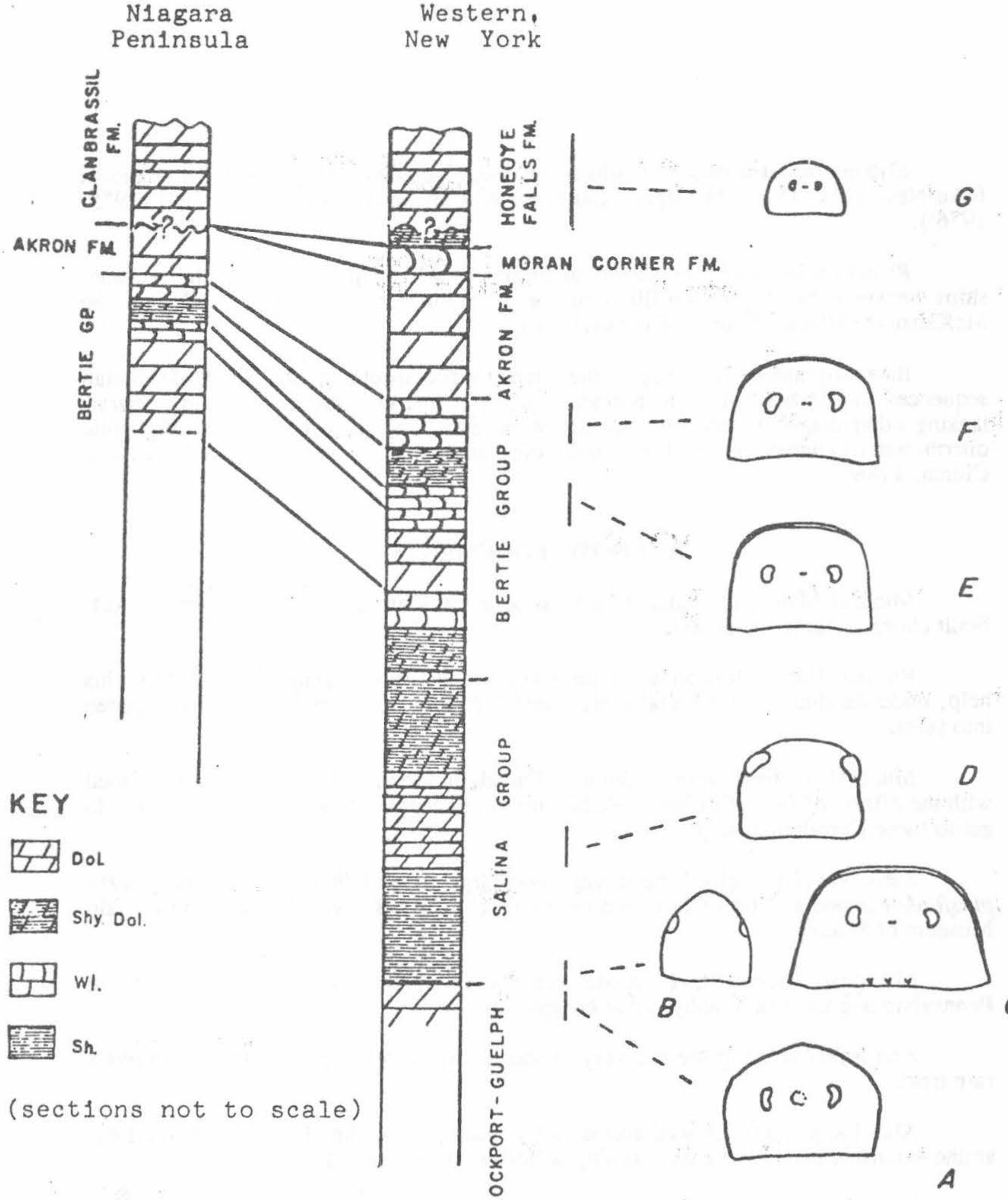


Figure 7 Eurypterid Biostratigraphy: Zonation based on the stratigraphic distribution of characteristic Silurian and Devonian eurypterids, A) Tylopterella, B,C) Hughmilleria socialis, Eurypterus pittsfordensis (respectively), D) Waeringopterus, E) Eurypterus remipes, F) Eurypterus lacustris, and G) Erieopterus.

Erieopterus straddles the Siluro-Devonian boundary and is currently known from New York, Ohio, Michigan, and Ontario, Canada (Ciurca and Gartland, 1975; 1976)).

Rhinocarcinosoma (not shown in Figure 6) will help to establish the relationships between the Sconondoa/Illion of the Lockport Group in New York and the McKenzie/Mifflintown strata of Pennsylvania.

Biostratigraphically, most of the characteristic species of the Silurio-Devonian sequences should be utilized in regional studies. Most eurypterids occur in horizons lacking other diagnostic species. An eurypterid zonation, based upon the stratigraphic distribution of characteristic Silurian and Devonian forms, is given in Figure 7 (see also Ciurca, 1986b).

ACKNOWLEDGEMENTS

Michael Muto gave valuable assistance in excavating the "*Para*" zone, a difficult chore working under water.

Richard Hamell was an accomplice and got this report computized. Without his help, understanding, and especially his friendship, this work would never have gotten into print.

Much of the field work in Ontario, Canada in the mid 1970's was accomplished with the efforts of Gene Gartland. We had many challenging times fighting blizzards to get to those Canadian quarries.

Steve Pavelsky helped me salvage two large stromatolites from the *Eurypterus pittsfordensis* bed at North Chili; one of the specimens has been placed in the Buffalo Museum of Science.

My dear friend Steve Jarose generously helped me do field studies in Pennsylvania-great stratigraphy, great eurypterids.

Ann Raker aided in the recovery of specimens of *Parascarcinosoma scorponis*-a rare treat.

Alan Lang freely allowed access to his quarry on Spohn Hill where I could examine extensive bedding planes exposing windrows of eurypterid material.

I have had the benefit of many interesting geoconversations with Professor Carlton Brett - I really appreciate it Carl.

This work is dedicated to the late Erik N. Kjellesvig-Waering from whom I had the benefit of ten years of fruitful exchange of information and ideas; to my dear friends, Suzi and Rich Hamell; and to those who try to preserve our wild spaces.

REFERENCES

Alling, H. L. 1928. The geology and origin of the Silurian salt of New York State. N.Y.S. Mus. Bull. No. 275.

Ciurca, S. J., Jr. 1967. The Honeoye Falls Dolostone Beds. Preliminary Report, Museum of Petrified Wood, 8 p.

_______________. 1973. Eurypterid horizons and the stratigraphy of Upper Silurian and Lower Devonian rocks of western New York State. N.Y.S.G.A. 45th Ann. Mtg., Monroe Comm. College and SUNY College at Brockport, New York, p. D1-D14.

_______________. 1975. Eurypterids and the position of the Silurian-Devonian boundary in New York State. G.S.A. Bull., v. 7, n. 1, p. 39.

_______________. 1978. Eurypterid horizons and the stratigraphy of Upper Silurian and Lower Devonian rocks of central-eastern New York State. N.Y.S.G.A. 50th Ann. Mtg., Syracuse Univ., Syracuse, N.Y., p. 225-229.

_______________. 1982. Eurypterids, stratigraphy, Late Silurian-Early Devonian of western New York State and Ontario, Canada. N.Y.S.G.A. 54th Ann. Mtg., SUNY Buffalo, Buffalo, N.Y., p. 99-120.

_______________. 1984. New eurypterid zones within the lower Vernon Formation in Western New York State. *In* Rochester Acad. of Sci. Proc. (1986), abstr.

_______________. 1985a. An eurypterid from the Silurian Lockport Group in Western New York. *In* Rochester Acad. of Sci. Proc. (1986), abstr.

_______________. 1985b. A new eurypterid horizon (Harris Hill Bed) in the Vernon Formation, Salina Group, Silurian of Western New York. *In* Rochester Acad. of Sci. Proc. (1986), abstr.

_______________. 1986a. A fantastic new eurypterid occurrence in Eastern New York State: Farmers Mills Bed, Illion-Vernon Transition. *In* Rochester . of Sci. Proc. (1986), abstr.

_______________. 1986b. Eurypterid stratigraphy in a Silurian evaporite sequence. S.E.P.M. 3rd Ann. Midyear Mtg., v. 3, p. 22, Raleigh, N.C.

_______________. 1989. Can of Worms Biostrome, Silurian Lockport Group at Rochester, New York. Rochester Acad. of Sci. Proceedings, abstr.

_______________. 1990. Minerals of the Can of Worms Biostrome. Lapidary Journal, v. 44, no. 4, p. 36-40.

________________ and M. Domagala. 1988. Silurian algal mound/eurypterid association, New York State and Pennsylvania. Rochester Acad. of Sci. Proceedings, abstr.

________________ and E. F. Gartland. 1975. Stratigraphy of the Eurypterid-bearing Bertie Group in the Niagara Peninsula of Southern Ontario, Canada. Abstr., North-Central Mtg., G.S.A. Bull., v. 7, n. 6, p. 737-738.

________________ and ____________. 1976. An Upper Silurian brecciated waterlime unit bearing Eurypterids, Niagara Peninsula of Southern Ontario, Canada. Abstr. North-Central Mtg., G.S.A. Bull., v. 8, n. 4, p. 472.

Clarke, J. M. and R. Ruedemann. 1912. The Eurypterida of New York. N.Y.S. Museum Memior No. 14, 2 vols., 628 p.

Dale, N. C. 1953. Geology and mineral resources of the Oriskany Quadrangle (Rome Quadrangle). N.Y.S. Museum Bulletin 345, 197 p.

Fisher, D. W. 1957. Lithology, paleoecology, and paleontology of the Vernon Shale (Late Silurian) in the type area. N.Y.S. Mus. Bull. 364, 31 p.

___________. 1960. Correlation of the Silurian rocks in New York State. N.Y.S. Mus. Chart and Map Series No. 1.

Hamell, R. D. 1978. A new occurrence of the Silurian Eurypterid: *Eurypterus pittsfordensis*. Rochester Academy of Science Proceedings, abstr., 1980, v. 130, nos. 2-4, p. 149.

___________. 1981. Stratigraphy, petrology and paleoenvironmental interpretation of the Bertie Group (Late Cayugan) in New York State. Unpubl. M.S. thesis, Unv. Rohester, Rochester, New York, 89 p.

___________. 1985. Stratigraphy and paleoenvironmental interpretation of the Bertie Group (Late Cayugan) in New York State. G.S.A. Bull. Abstr., NE Section 10th Ann. Mtg., v. 17, no. 1, p. 40.

___________. 1986. Paleoenvironmental interpretation of the Bertie Group (Late Silurian) in New York State. S.E.P.M., 3rd Midyear Mtg., Raleigh, N.C., v. 3, p. 49.

____________ and S. J. Ciurca, Jr. 1986. Paleoenvironmental analysis of the Fiddlers Green Formation (Late Silurian) in Western New York State. N.Y.S.G.A 58th Ann. Mtg., Cornell Univ., Ithaca, New York, p. 199-218.

Hopkins, T. C. 1914. Geology of the Syracuse Quadrangle. N.Y.S. Mus. Bull. 171, 80 p.

Leutze, W. P. 1956. Faunal stratigraphy of Syracuse Formation, Onondaga and Madison Counties, New York. A.A.P.G. Bull., v. 40, p. 1693-1698.

__________. 1961. Arthropods from the Syracuse Formation, Silurian of New York. Jour. Paleon., v. 33, p. 49-64.

__________. 1964. The Salina Group. N.Y.S.G.A. 36th Ann. Mtg., Syracuse Univ., Syracuse, NY, p. 57-65.

Rickard, L. V. 1962. Late Cayugan (Upper Silurian) and Helderbergian (Lower Devonian) stratigraphy in New York. N.Y.S. Mus. Bull 386, 157 p.

__________. 1969. Stratigraphy of the Upper Salina Group-New York Pennsylvania, Ohio, Ontario. N.Y.S. Mus. Map and Chart Series 12.

__________. 1975. Correlation of the Silurian and Devonian rocks in New York State. N.Y.S. Mus. Map and Chart Series 24, 16 p.

Ruedemann, R. 1919. A recurrent Pittsford (Salina) fauna. N.Y.S. Mus. Bull. 219-220, p. 205-222.

Sarle, C. J. 1903. A new eurypterid fauna from the base of the Salina of Western New York. N.Y.S. Paleontologist's Rep't, p. 1079.

Williams, M. Y. 1915. An eurypterid horizon in the Niagara Formation of Ontario. Geol. Surv. Canada Mus. Bull. 20, p. 21.

Zenger, D. H. 1962. Proposed stratigraphic nomenclature for the Lockport Formation (Middle Silurian) in New York State. A.A.P.G. Bull., v. 46, p. 2249-2253.

__________. 1965. Stratigraphy of the Lockport Formation (Middle Silurian) in New York State. N.Y.S. Mus. Bull. 404, 210 p.

ROAD LOG

Road log is divided into two parts. Part 2 only if time permits.

PART 1

Point-Point Mileage	Cummalative Mileage	Description
0.0	0.0	Inspection Station entering Canada.
0.1	0.1	EXIT Fort Erie. Follow sign to Rte 3.
0.5	0.6	TURN RIGHT onto Rte. 3 West.
0.2	0.8	Ideal Donuts on left.
0.3	1.1	McDonalds on right, proceed west.
2.6	3.7	Pink Elephant on right.
1.1	4.8	Stonemill Road.
0.2	5.0	TURN RIGHT onto Ridgemount Road.
0.7	5.7	Ridgemount Quarries Ltd. Aggregate Div. of Walker Industries. Quarry on left.
0.8	6.5	TURN LEFT onto Bridge Street.
0.2	6.7	Quarry overpass. Park on left.

STOP 1 Ridgemount (Campbell) Quarry

Camillus Shale or Oatka Shale forms the quarry floor. Above this is a thin representation of the Morganville Waterlime followed by massive beds of the Victor Dolostone. Several feet of Ellicott Creek Breccia are accessible on the northeast side of the quarry and from this many eurypterids have been obtained (Ciurca and Gartland, 1976). The higher strata, including the Williamsville Waterlime, are all visible in the quarry walls. The Devonian Bois Blanc Formation is particularly fossiliferous in this quarry.

0.2	6.9	Return to Ridgemount Road. TURN RIGHT, follow Ridgemount Road south.
1.5	8.4	Route 3 intersection. TURN RIGHT (west) toward Port Colburne.
2.0	10.4	Ridgeway Battlefield.
0.1	10.5	Intersection Ridge Road (signal light), continue west.
3.9	14.4	Entering Port Colbourne Township.
2.9	17.3	Gasline.
3.2	20.5	TURN RIGHT, Rte. 140 North.
0.6	21.1	Chippawa Road. Port Colbourne Quarries Ltd. entrance on right. Fine section of the Bertie Group and overlying strata in this quarry complex. CONTINUE north toward Welland.
4.0	25.1	Welland City Line. Population 46,000.
0.1	25.2	TURN RIGHT. Intersection Rte. 58A.
0.1	25.3	BEAR RIGHT
0.1	25.4	TURN RIGHT, Rte. 58A West.
0.5	25.9	Intersection Rusholme Road. CONTINUE west.

STOP 2 **Welland Canal Underpass Roadcut**

SYRACUSE FORMATION (SALINA GROUP). A portion of the Syracuse Fm. is well exposed exhibiting gypsum beds, fine-grained dolostone replete with cavities formed by dissolution of evaporites. Crude algal mounds are present and the facies developed here is identical to that displayed in the type area in central New York.

Waeringopterus, characteristic of the lower or transitional beds of the Syracuse Fm., occurs here in the argillaceous dolostones.

0.5	26.4	CONTINUE into tunnel.
0.3	26.7	Leaving tunnel.
1.6	28.3	TURN LEFT, intersection Rte. 58. Port Colbourne 8 km., CONTINUE SOUTH.
1.3	29.6	Entering City of Port Colbourne. Population 16,000.
2.0	31.6	Outcrop.
1.2	32.8	Jct. Rte. 3, McDonald's, TURN RIGHT (west).
1.2	34.0	Entering WainFleet Township.
0.8	34.8	Law Quarry on right. Fine sections exposing Bertie Group, Bois Blanc Limestone.
2.6	37.4	Ostryhon Corners.
3.0	40.4	Wainfleet
1.6	42.0	FOLLOW Rte. 3 west.
3.1	45.1	Winger
3.2	48.3	Town of Dunnville. Population 11,500.
7.1	55.4	Downtown Dunnville. CONTINUE west Rte. 3.
0.8	56.2	STOP SIGN. TURN LEFT. DO NOT GO RIGHT onto Main Street west. Grand River in sight.
1.4	57.6	SIGNAL LIGHT, Queen Street. TURN RIGHT.
0.1	57.7	Grand River.
0.2	57.9	Byng Island.
0.2	58.1	Grand Island BBQ. TURN LEFT, Rte. 11.
0.3	58.4	Dunnville Rock Products Ltd. Quarry and Plant. TH & B Bldg., 6620 Broad Street East. Trip Part 1 ends. Go back or begin Trip Part 2.

STOP 3 **Quarry at Byng**

The Bertie Group is prominent in the lower portions of the quarry. Particularly important is the sequence above the Cobllekill Formation at this site. The Clanbrassil Formation makes its appearance beneath the unconformity at the base of the Bois Blanc Formation. In this quarry the Clanbrassil Fm. consists primarily of waterlimes, i.e. very fine-grained dolostones having conchoidal fractures. The only fossil obtained from the Clanbrassil Fm. at this locality is *Erieopterus*.

TRIP PART 2 (if time permits)

0.0		Leave quarry.
0.2	0.2	TURN LEFT. Head back to RTE. 3
0.1	0.3	JCT., TURN RIGHT to Grand River.
0.4	0.7	Grand River.
0.1	0.8	JCT. Main Street. Dunnville. TURN LEFT. Proceed to Rte.3 (Grand River on left).
0.6	1.4	STOP SIGN, JCT. Rte. 3, proceed west on Rte 3.
5.0	6.4	Railroad Crossing.
0.2	6.6	Railroad Crossing.
1.7	8.3	Canborough, follow Rte. 3 west.
1.9	10.2	Town of Haldiman, Population 18,500.
1.8	12.0	Canfield
1.3	13.3	JCT. 56
0.3	13.6	Railroad Crossing.
3.8	17.4	Cayuga
0.6	18.0	JCT. Rte. 54.
0.2	18.2	Grand River.
1.2	19.4	JCT. Rte. 8.
1.2	20.6	Decewsville.
1.3	21.9	Cayuga Quarry (Crushed Stone).

STOP 4 Cayuga Quarry

This large quarry complex contains an important section including the Williamsville Waterlime and the Clanbrassil Formation. *Erieopterus* seems to be more common in the Clanbrassil Fm. and in the quarries at Haggersville.

0.6	22.5	Dry Lake Road. TURN RIGHT.
0.1	22.6	Railroad Crossing.
2.4	25.0	STOP SIGN, TURN RIGHT.
0.1	25.1	Clanbrassil.
0.7	25.8	Abandoned quarries on both sides of road. Stone Crest Farm.

STOP 5 Abandoned Quarry

Type section Clanbrassil Formation. The Clanbrassil Formation occurs in the walls of an abandoned quarry at the Stone Crest Farm.

END OF TRIP

TRIP II - SUNDAY

ROAD LOG

SCENIC ROUTE

Starting Point: Sign (N.Y. Power Authority, Administration Bldg., Switchyard, Warehouse) on NY 265 heading south from I-190 near Lewiston, New York.

STOP 1

Section high in Lockport Group ("Oak Orchard" of Zenger). Large stromatolite mounds.

START LOG

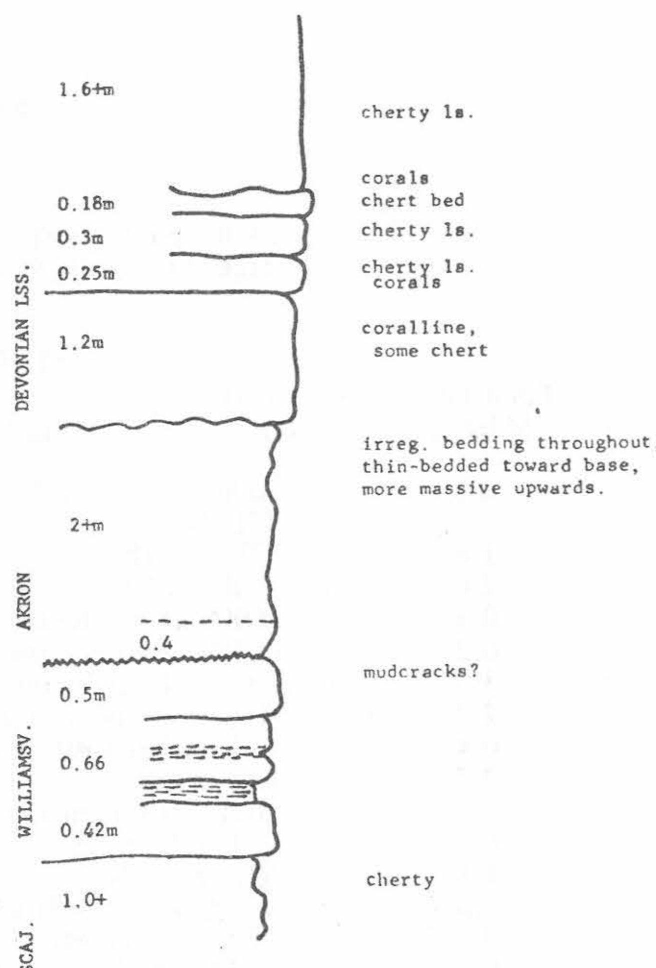
Point-Point Mileage	Cummalative Mileage	Description
0.0	0.0	Follow NY 265 (Military Rd.) south to NY 324 (Sheridan Drive).
0.7	0.7	JCT. NY 31.
1.8	2.5	JCT. NY 182.
2.0	4.5	TURN RIGHT, follow NY 265.
0.4	4.9	TURN LEFT, follow NY 265.
0.2	5.1	TURN LEFT, follow NY 265.
3.0	8.1	City of North Tonowanda. Niagara River on right.
2.0	10.1	JCT. 429, continue south on NY 265.
0.4	10.5	JCT. NY 384 BEAR LEFT. Follow NY 384.
0.7	11.2	Erie County.
		JCT 324 (Sheridan Dr.) TURN LEFT, follow NY 324 east.
6.5	17.7	JCT. NY 263.
1.8	19.5	JCT. NY 277.
1.0	20.5	Mill Street. TURN RIGHT.
0.9	21.4	Glen Ave. TURN RIGHT.
0.1	21.5	GLEN PARK-manmade wonder complete with ducks.

STOP 2 <u>ELLICOTT CREEK BRECCIA</u>

Type Sections: Ellicott Creek Breccia, Williamsville Waterlime.

Figure below shows stratigraphic section of part of Bertie Group.

0.0	21.5	Head back to Mill Street.
0.1	21.6	Mill Street. TURN RIGHT to JCT. NY 5 (Main Street). TURN LEFT (east). FOLLOW NY 5.
2.5	24.1	JCT. NY 78.
2.5	26.6	JCT. NY 324.
2.6	29.2	Clarence, New York.
1.1	30.3	Antique World on right.
1.9	32.2	View from escarpment on left-WHEW!
0.5	32.7	Clarence Materials Corp. Redi Mix Div. Main St. Plant.



Section at Williamsville, New York, Ellico Creek, west side, tributary showing Scajaqu through Akron Formational relationships.

STOP 3 FIDDLERS GREEN FM. ETC

Inactive quarry exposing a portion of Fiddlers Green Fm. up through Cobleskill Fm. Large quarry in Onondaga Limestone just to the east.

1.6	34.3	JCT. to NY 93, continue east.
3.0	37.3	Pembroke, New York. JCT. NY 77 TURN LEFT. Proceed to Indian Falls.

STOP 4 INDIAN FALLS-FALCON CREST

4.4	41.7	Falls consists of massive Victor Dolostone-lush green algal mats still thrive on bedding planes, in pools and channels. No eurypterids have been seen.

0.1	41.8	TURN RIGHT, head back to NY 5 via NY 77.
2.3	44.1	JCT. NY 5 TURN LEFT, heading to Batavia and points beyond.
11.4	55.5	Downtown Batavia-nice place.
1.1	56.6	Leaving Batavia.
2.2	58.8	Neat boulder on left, continue east on NY 5.
1.5	60.3	Genesee LeRoy Stone Quarry on RIght.
0.6	60.9	Stafford, N.Y.
3.8	64.7	LeRoy, N.Y.
0.7	65.4	JCT. NY 19 TURN LEFT (north).
2.0	67.4	Railroad overpass. BE PREPARED TO PULL OFF ROAD.
0.7	68.1	JCT Oatka Trail - TURN RIGHT, PULL OVER.

STOP 5

Stratigraphic section showing Camillus Shale (Salina Group) and overlying Fort Hill Waterlime.

0.0	68.1	FOLLOW Oatka Trail (east).
0.3	68.4	TURN RIGHT - Oatka Trail (WINDING ROAD).
0.8	69.2	JCT Circular Hill Rd. TURN RIGHT, proceed up the escarpment.
1.7	70.9	JCT. Gulf Road. TURN LEFT.

STOP 6

0.8	71.7	Dolomite Products Co., Inc. LeRoy Plant. Old locomotive ensemble, quarries in Onondaga.

0.0	71.7	Continue east on Gulf Road.
0.5	72.2	Neid Road. TURN LEFT.

STOP 7

0.2	72.4	Erratics of Victor Dolostone - note *Whitfieldella* brachiopods littering bedding planes. Continue north on Neid Rd.
0.1	72.5	JCT. Town of LeRoy Refuse Area. TURN LEFT.

STOP 8

0.1	72.6	Neid Road Quarry (Ciurca, 1973). Bertie Group and overlying Onondaga Limestone exposed here.
0.1	72.7	JCT. Neid Road. TURN RIGHT.
0.3	73.0	STOP - JCT. Gulf Road. TURN LEFT (east).
1.0	74.0	County of Monroe - Flint Hill Road.
1.2	75.2	Genesee Country Museum, Nature Center on Right.
1.3	76.5	JCT. NY 36. TURN LEFT.
0.2	76.7	Oatka Creek.
0.1	76.8	JCT. 383. TURN RIGHT. Follow NY 383 north - ENJOY BEAUTIFUL SCENERY.
3.7	80.5	Garbutt, N.Y.
0.2	80.7	JCT. Union St. TURN RIGHT.
0.1	80.8	Railroad Crossing. Ronzo's Grocery on right.

STOP 9

TURN RIGHT into Oatka Park. Exposures of Syracuse Fm. (Salina Group) along Oatka Creek. Ruins of buildings-old gypsum mines.

0.1	80.9	FOLLOW Union Street back to NY 383.
0.1	81.0	STOP-TURN RIGHT onto NY 383 north.
0.8	81.8	Railroad overpass.
0.6	82.4	Scottsville, N.Y.
1.5	83.9	JCT. NY 253. Follow NY 383 north (Rochester-6.0 mi.).
8.6	92.5	JCT. I-390.

END OF TRIP

UPPER DEVONIAN TURBIDITES IN WESTERN NEW YORK: CHARACTERISTICS AND IMPLICATIONS FOR SUBMARINE FAN DEPOSITION MODELS

ROBERT JACOBI
Department of Geology; SUNY at Buffalo; 4240 Ridge Lea Rd; Buffalo, NY 14226

MICHAEL GUTMANN
URS Consultants; 570 Delaware; Buffalo, NY 14202

AL PIECHOCKI
Department of Geology; SUNY at Buffalo; 4240 Ridge Lea Rd;
Buffalo, NY 14226

SUZANNE O'CONNELL
Department of Earth and Environmental Sciences; Wesleyan University; Middletown, CT 06457

JILL SINGER
Department of Earth Sciences; State University College at Buffalo; 1300 Elmwood Ave.;
Buffalo, NY 14222

CHARLES MITCHELL
Department of Geology; SUNY at Buffalo; 4240 Ridge Lea Rd;
Buffalo, NY 14226

STEVE FRANK
Department of Geology; SUNY at Buffalo; 4240 Ridge Lea Rd;
Buffalo, NY 14226

DAVID SCHEUING
Lamont-Doherty Geological Observatory and Department of Geological Sciences;
Columbia University; Palisades, NY 10964

STEVE HASIOTIS
Department of Geology; SUNY at Buffalo; 4240 Ridge Lea Rd;
Buffalo, NY 14226

PREFACE

The object of this field trip is to acquaint you with some of the characteristics of Upper Devonian turbidites in western New York* (WNY). We will examine three coarsening-upward clastic depositional cycles within the West Falls, Perrysburg and Java formations. The many gorges and ravines in WNY provide the control necessary to trace individual sandstone beds across the region, allowing us to characterize their interrelationships and their lateral variabilty in bedforms, texture, and thickness. From these and other data we believe the WNY turbidites in this part of the sequence represent lobe fringe and sand lobe deposition on a submarine fan. Continued study of these turbidites will help resolve problems encountered in both modern and ancient submarine fan research. Construction of a depositional model based on this research will entail 1) deposition and flow characteristics for individual non-channelized turbidity currents, 2) relationships among various fan elements, and 3) fan vs. clastic ramp development. A more detailed discussion of the research rationale and results is presented in Jacobi et al. (in press).

STRATIGRAPHY AND SEDIMENTOLOGY OF THE UPPER DEVONIAN

Overview

The Upper Devonian section in WNY and central New York (CNY; Figs. 1,2) records an infilling of the Catskill Sea from sources to the east. The Upper Devonian in this region consists of a number of major sedimentation cycles; each coarsening-upward cycle is marked by a basal black shale that grades upward into gray and greenish-gray shales, which in turn are interbedded with sandstone/siltstone** beds near the top of a particular cycle (e.g., Pepper and de Witt, 1951; Pepper et al., 1956; Sutton, 1963; Buehler and Tesmer, 1963; Kirchgasser and House, 1981; van Tyne, 1982, 1983; Sevon and Woodrow, 1985). It is these sandstone beds that we will observe on this field trip.

The origin proposed for these sandstones varies markedly among different units, different regions of the same unit, and even among different researchers. A large number of paleoecological studies have been conducted on the Upper Devonian "sandy" sections east of, and older than, the units we will examine on this field trip (e.g., McGhee and Sutton, 1985, and references therein). These paleoecological studies suggest that the Frasnian Chemung facies (Caster, 1934) of CNY represent a collage of shallow marine environments, including delta front, delta platform, and channel/estuary environments.

** WNY is defined here as the area west of the Genesee Gorge (Fig.2).*
*** "sandstone/siltstone" will be referred to as simply "sandstone".*

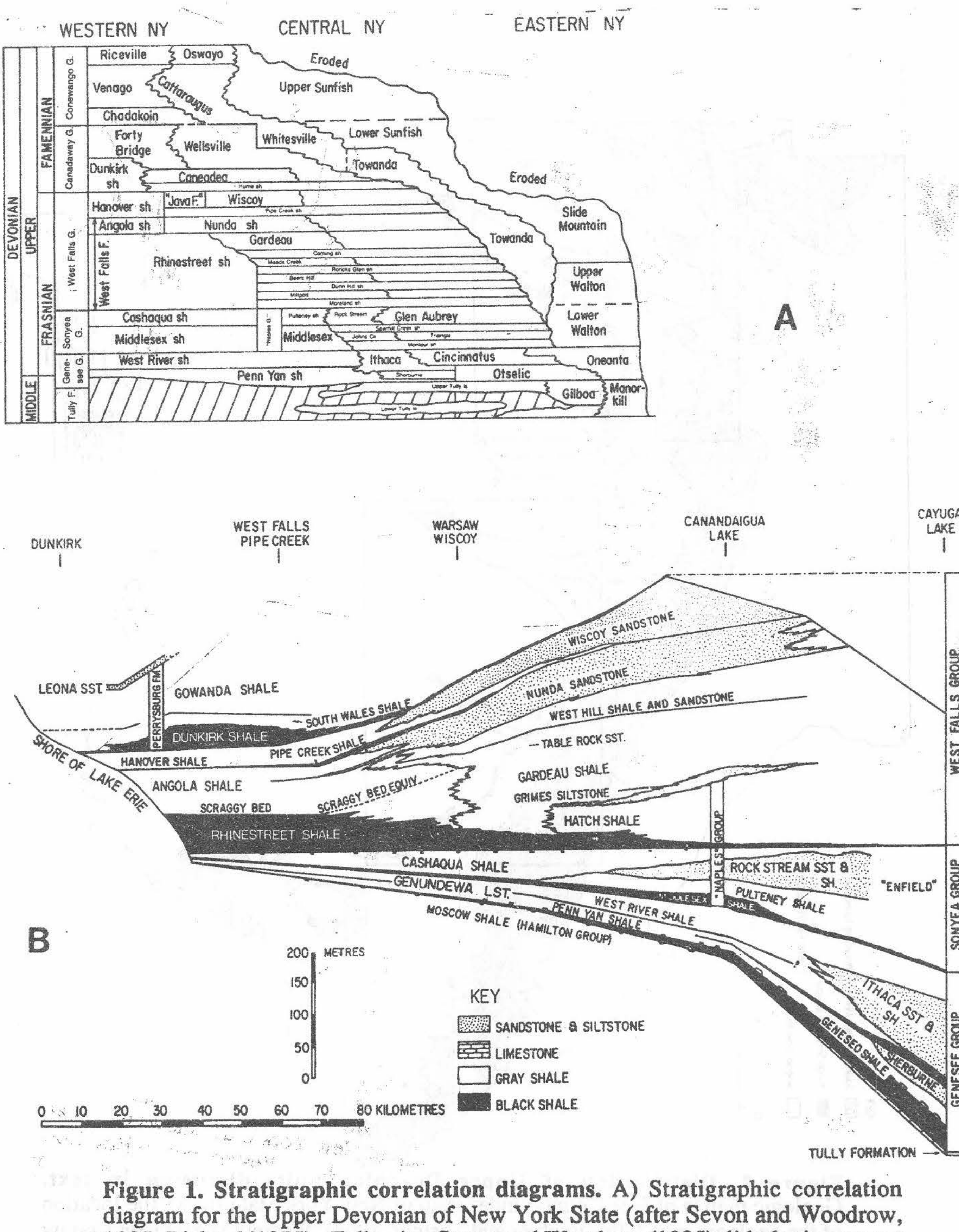


Figure 1. Stratigraphic correlation diagrams. A) Stratigraphic correlation diagram for the Upper Devonian of New York State (after Sevon and Woodrow, 1985; Rickard, 1975). Following Sevon and Woodrow (1985), lithologies are not capitalized. In this chart, rocks of the Perrysburg Formation are included in the Dunkirk shale. B) Facies diagram of the Upper Devonian in western and central New York (from Kirchgasser and House, 1981; Johnson, et al., 1985).

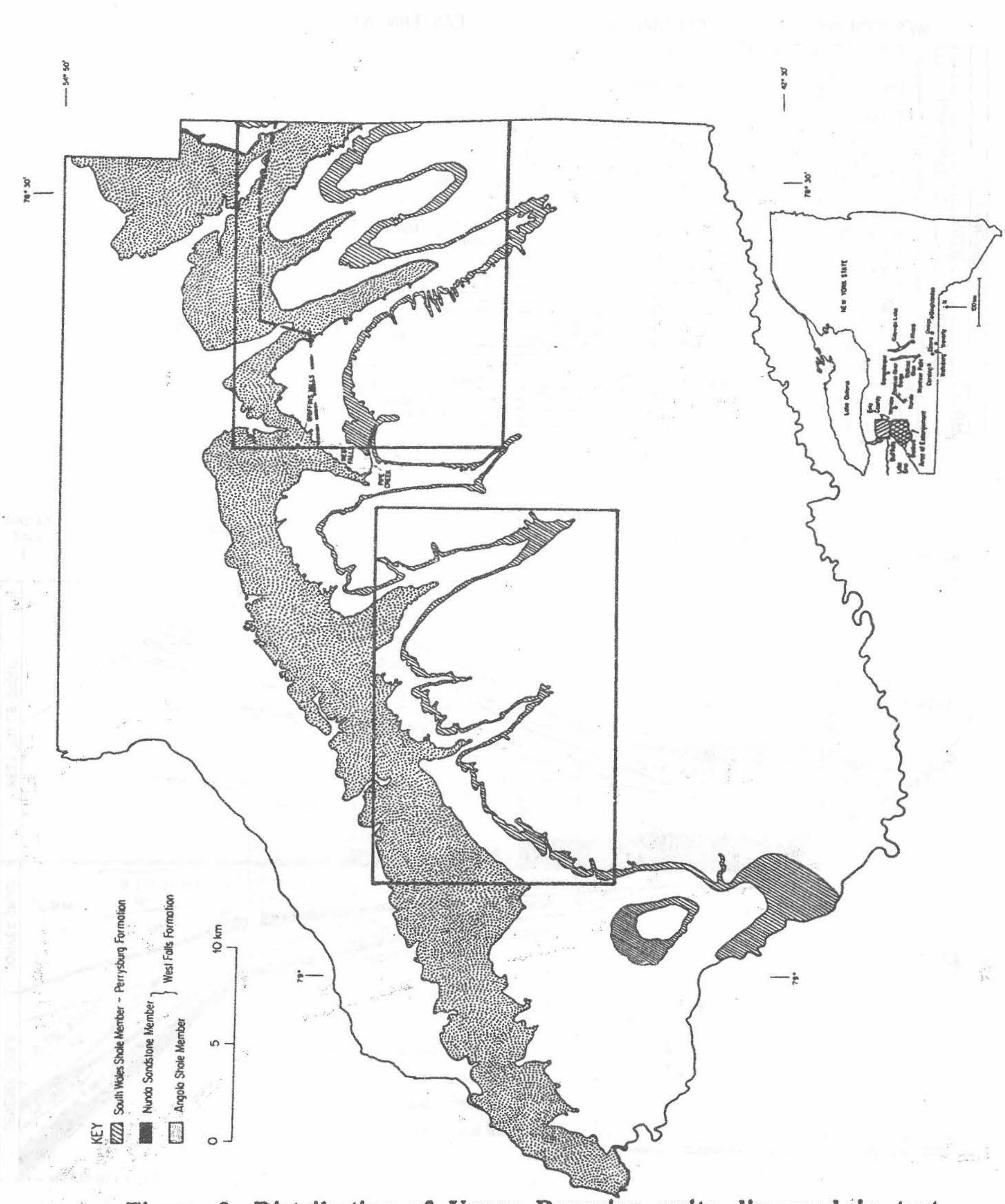


Figure 2. Distribution of Upper Devonian units discussed in text. Geologic map is of Erie County; index map of New York State shows the location of the geologic map (after Buehler and Tesmer, 1963). Western box in Erie County shows location of South Wales Shale Member maps (displayed in Fig. 4), and eastern box shows location of Nunda Sandstone Member maps (displayed in Fig. 6). Dashed line locates stratigraphic cross section displayed in Figure 5.

Sedimentological studies typically have concentrated on a few discrete vertical sections in the lower sandy units of the Frasnian in CNY (Walker and Sutton, 1967; Woodrow and Isley, 1983; Craft and Bridge, 1987). These detailed studies suggest that sandstones in the western part of CNY represent turbidites deposited on the basin slope and floor (broadly, the Portage facies of Caster, 1934), whereas some of the eastern sands represent storm deposits on a platform. In WNY there have been no sedimentological studies of the sandy units in the Upper Devonian Portage facies that postdate the recognition of turbidites and storm deposits.

Devonian turbidites in the Catskill Sea generally have been thought to be more similar to nonchannelized, ramp deposits rather than to modern submarine fan deposits (Woodrow and Isley, 1983; Woodrow, 1985; Lundegard et al.,1985; Van Tassell, 1987). The clastic ramp model involved sheet flows from line sources such as storms, rather than from point sources at channel mouths or slide complexes. However, data gathered by our research group suggest that a submarine fan model <u>is</u> appropriate for the Devonian turbidites in WNY. The arguments promoting a clastic ramp model centered on two lines of negative evidence: 1) absence of submarine channels; and 2) the lack of radial flow patterns in nonchannelized turbidites. However, our initial research in WNY reveals a systematic areal variation in paleoflow that describes a partial radial flow pattern of 40° over a distance of 18 km., and van Tyne (1982) identified macroscale submarine channels in the Upper Devonian sequence, based on isopach variations.

"SANDSTONE" UNITS TO BE OBSERVED ON THE FIELD TRIP

We will examine the Portage facies sandstone packets in three coarsening-upward cycles that occur above the lower West Falls Formation in WNY (Figs. 1,2). Our main focus will be on the South Wales Shale Member of the Perrysburg Formation and the Nunda Sandstone Member of the West Falls Formation. We also will observe the Wiscoy Sandstone Member of the Java Formation.

*South Wales Shale Member of the Perrysburg Formation (Pepper and de Witt, 1951)**

The South Wales Shale Member consists of fine-grained sandstones and siltstones interbedded with greenish-gray shale (Figs. 1, 3; Pepper and de

**Stratigraphic designations for each unit we propose to study have not been consistently employed (Clarke, 1903; Clarke and Luther, 1908; Hartnagel, 1912; Pepper and de Witt, 1951; Pepper et al., 1956; Buehler and Tesmer, 1963; Rickard, 1975; Kirchgasser and House, 1981; Sevon and Woodrow, 1985; compare Fig. 1A vs. Fig. 1B). We will employ the conventions of Pepper and de Witt, (1951), Pepper et al. (1956), and Kirchgasser and House (1981) (Fig. 1B).*

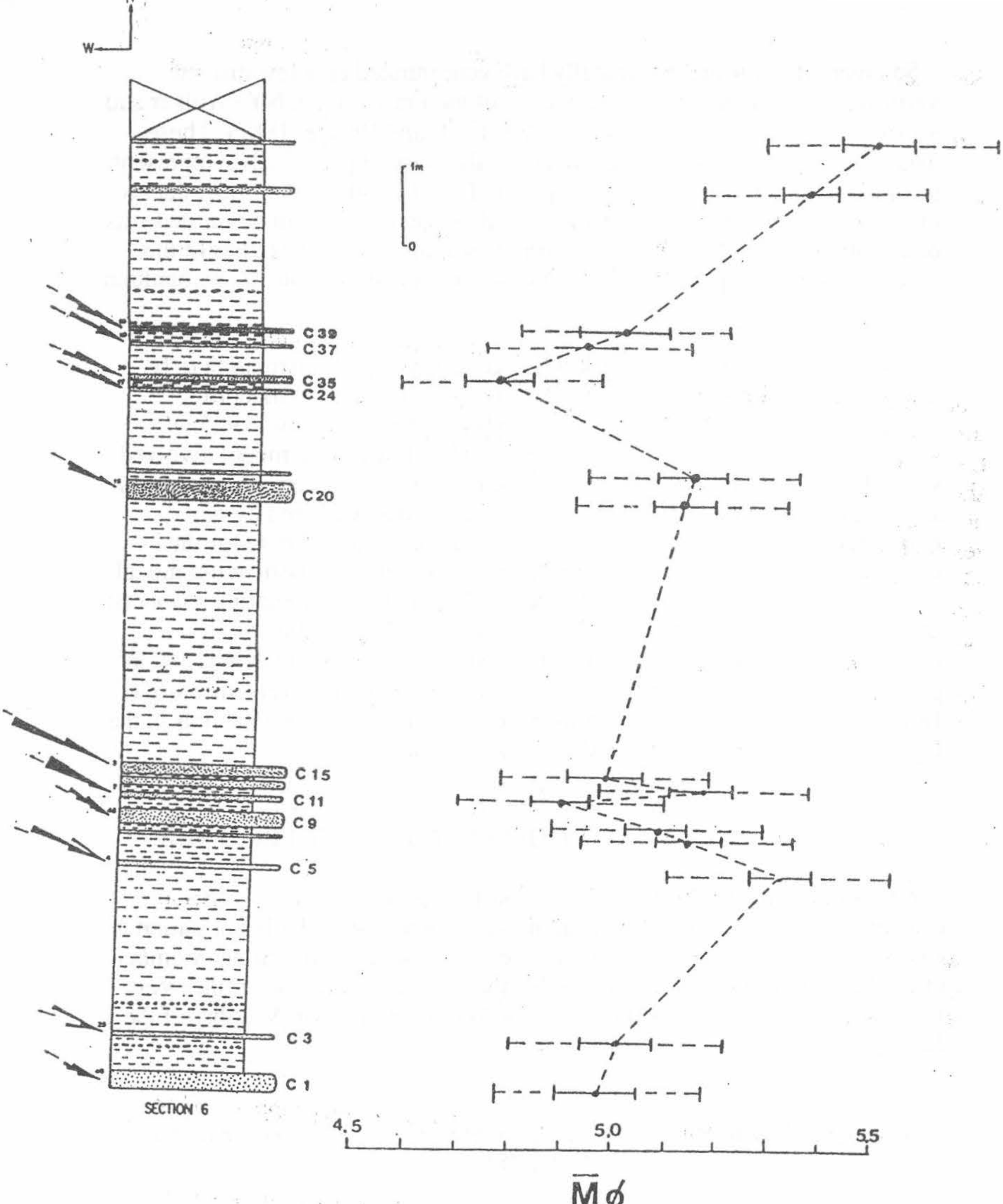


Figure 3. Columnar section of the South Wales Shale Member at STOP 1 and upsection grain size variations. Locality of measured section shown in Figures 7 and 8. In the columnar section calcareous concretion horizons are shown as layers of small circles. Both rose diagrams and averages for current indicators in sandstone units are displayed left of the columnar section. Upsection variations in grain size are displayed right of the columnar section. Solid error bar indicates sampling error and dashed error bar indicates measuring error. Figure after Gutmann and Jacobi (1988) and Gutmann (1989).

Witt, 1951; Buehler and Tesmer, 1963). The South Wales Shale thins from a maximum of 24m at Lake Erie to a minimum of 6m at the Genesee Gorge (Pepper and de Witt, 1951). Because the only study of the South Wales that included sedimentology predates the common knowledge of turbidites, the turbiditic nature of the South Wales sandstone units was not recognized; rather, Pepper and de Witt (1951) believed the sands represented stream deposits and delta deposits.

Nunda Sandstone Member of the West Falls Formation (Clarke , 1897)

The Nunda Sandstone generally displays either a massive or an undulatory, flaggy bedded character (Pepper et al., 1956; Buehler and Tesmer, 1963). A tongue of Nunda Sandstone contained within the Angola Shale extends west into Erie County. The Nunda is ~75 m thick near the Genesee Gorge and thins westward to an abrupt pinchout in central Erie County (Pepper et al., 1956; Buehler and Tesmer, 1963; Piechocki et al., 1990).

Wiscoy Sandstone Member of the Java Formation (Group) (Hartnagel, 1912)

The Wiscoy Sandstone Member consists of interbedded sandstone and shale. To the west the Wiscoy correlates with the Hanover Shale which consists of ~30m of gray shale interbedded with black shales at the base and thin sandstones near the middle. The thick sections of sandstones typical of the Wiscoy in the east (e.g., Elmira) thin to the west (e.g. the Genesee Valley) and correlate with the thin sections of Sandstone in the middle of the Angola. From west (the Genesee Valley) to east (Elmira), the Wiscoy sandstone is thought to represent a series of different environments: open shelf, prodelta and delta platform (McGhee and Sutton, 1981). To the best of our knowledge there are no definitive paleoenvironmental studies of the Java Formation west of the Genesee Valley.

RESULTS OF OUR PRELIMINARY FIELDWORK

We have begun fieldwork in WNY on all the units described above; greater effort has been devoted to fieldwork and laboratory analyses on the South Wales Shale Member and the Nunda Sandstone Member (e.g., Gutmann and Jacobi, 1988; Piechocki et al., 1990; Jacobi et al., in press). The following section summarizes our findings for primarily the South Wales Shale in western and central Erie County (Fig. 2).

South Wales Shale Member

In the western exposures of the South Wales Shale Member (SWSM, Fig. 2), the SWSM can be divided into two units-a lower division with a relatively thick basal sandstone (bed C1 in section 6 at stop 1, Fig. 3), and an upper division that also has a thick basal sandstone (bed C20 in section 6, Fig. 3). The lower division varies from 5 to 7 m thick, and the upper division ranges from 4.5 to 6.5 m thick. Both divisions generally consist of two or more packets of sandstones (Fig. 3). Within the lower division, the average thickness of the sandstone beds varies from about 5 cm in the northeast to more than 9 cm in the southwest and southcentral areas. The average thickness of sandstone beds in the upper division varies from 1 cm to more than 10 cm. The percent of sandstone in the lower division varies from 0% to more than 10%. The percent-of-sandstone contours form a distinctly lobate pattern, which is somewhat consistent with the observed transport directions, especially in the western sections. The percent of sandstone in the upper division varies from 7% to 17%.

The lower contacts of the thicker sands in the South Wales Shale Member typically display abundant sole marks, including groove casts, groove with chevron casts, bounce and prod marks, and flute casts. Grooves and striations orthogonal to the ripple crestlines are both prominent and abundant, and were carved most likely by water-logged plant remains, although shell debris cannot be totally dismissed. Based on the sole marks and bedforms, paleocurrent determinations in all sandstone units reveal an overall west-northwesterly transport direction (Fig. 4). In detail, however, the paleoflow indicators on most individual beds display a systematic swing of about 40^{o} across western Erie County, a distance of some 18km (Fig. 4).

Many sandstone units display prominant straight-crested climbing ripples; these asymmetrical ripples are transverse to current flow deduced from sole marks, as well as parting lineations. Most of the sandstones exhibit Bouma sequences Tc, Tcd/e and a few display Tbc, Ta, and Tab. Additionally, the basal sandstone shows a systematic progression of Bouma sequences from east to west across Erie County, from Ta/Tab and Tbc to Tc. Based on the thickness and texture of the sandstones, Walker's (1967) ABC index, and the sand/shale ratio, most of the sandstones in the South Wales Shale Member are "distal" turbidites. Significantly, the sandstone units lack shallow marine or nearshore bedforms such as hummocky cross-stratification and herringbone pattern.

A working hypothesis for the origin and depositional environment of sandstones in the South Wales Shale Member can be constructed from the results of our preliminary study. First, as stated above, the bedforms are consistent with a "distal" turbidite origin. Second, individual turbidites appear to be nonchannelized because the sandstones display minimal sharp variations in character (e.g., thickness) across Erie County, and we observe neither significant erosion at the basal contacts nor typical channel facies.

The critical remaining question concerns the depositional setting of the non-channelized "distal" turbidites. The thin-bedded and continuous nonchannelized nature of the sands, the sand/shale ratio, and the bedforms

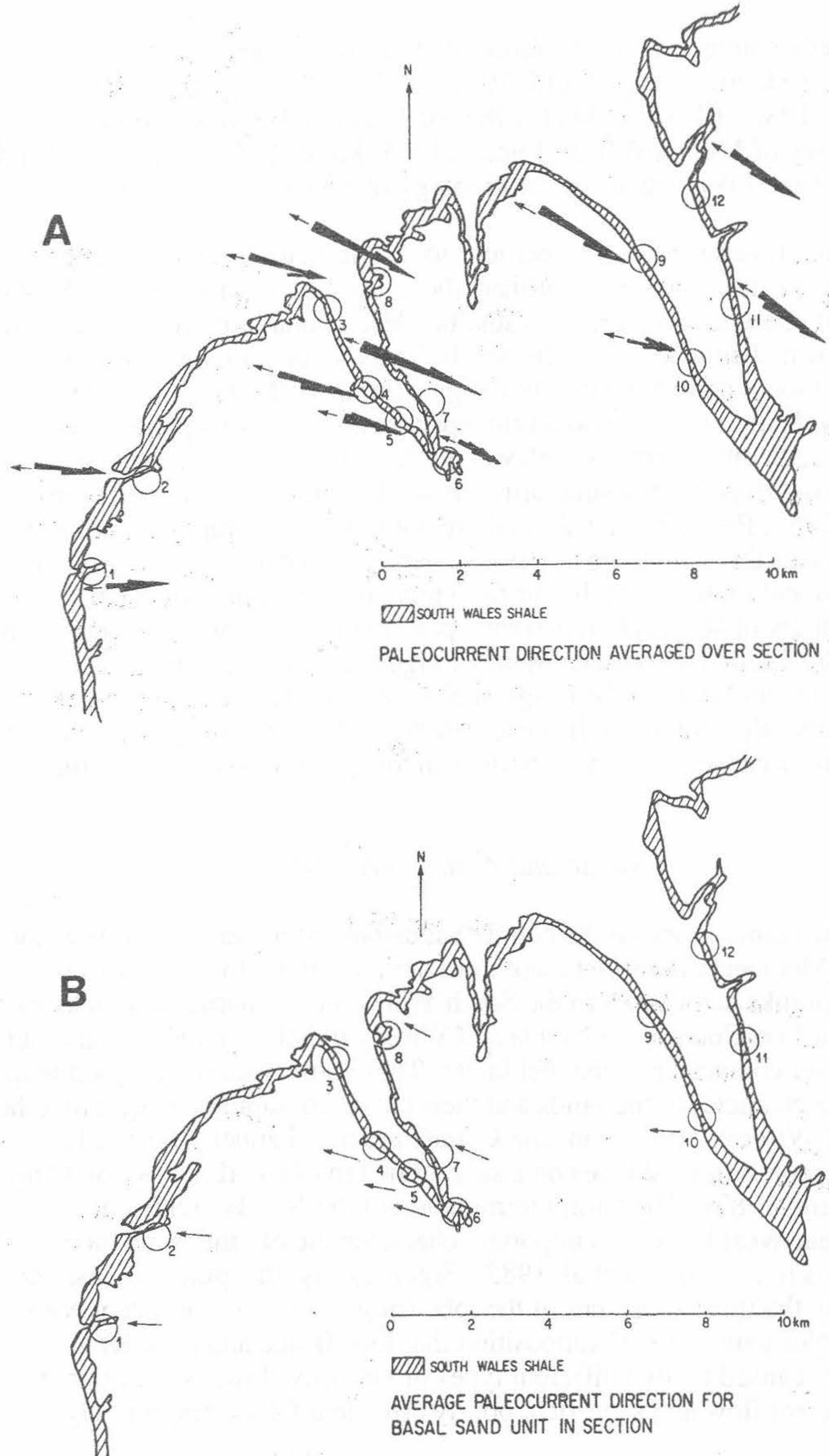


Figure 4. Paleocurrent directions maps. A) Paleocurrent directions for all sandstones at each measured section of the South Wales Shale Member; arrow denotes average direction. B) Average paleocurrent direction for the basal sand unit in the South Wales Shale Member. Figure from Gutmann and Jacobi (1988) and Gutmann (1989).

taken together suggest that the thin-bedded sandstones are either: 1) interchannel deposits (e.g., Mutti, 1977), 2) "distal" ramp deposits (e.g., Pickering, 1982; Chan and Dott, 1983) or 3) sand lobe-fringe deposits (terminology of Mutti and Ricci Lucchi, 1975; Mutti, 1977; "nonchannelized lobes" of Type I deposits in the terminology of Mutti and Normark, 1987).

We can utilize several considerations to discriminate among the three possible depositional settings outlined above for the sandstone beds. We have discounted interchannel facies as a suitable depositional setting because there are no known channel facies at the South Wales Shale horizon in the field area. Discrimination among the latter two origins (ramp vs. fan) may be achieved through evaluation of the regional flow patterns. Radial flow patterns are consistent with either crevasse splay deposits or lobe-fringe deposits that develop downslope from a submarine channel mouth (e.g. Hampton and Colburn, 1967; Pickering, 1982), and are not typical of ramp (or modified ramp) deposits that exhibit no systematic variation in paleocurrent direction (e.g., Chan and Dott, 1983). In our field area, the swing in paleoflow direction of about 40° might represent a portion of the radial flow pattern. The facies of the South Wales Shale Member argues against a crevasse splay interpretation, and thus a lobe fringe deposit seems consistent with our observations. We cannot yet dismiss, however, the possibility that the radial flow pattern is caused by local variations in topography on a clastic ramp.

Nunda and Wiscoy members

Reconnaissance work on the Nunda Sandstone Member and the Wiscoy snadstone Member reveals that these units are not similar to either "distal" turbidites (unlike sandstones in the South Wales) or submarine channel sands. Throughout Erie County, the Nunda and Wiscoy members exhibit neither an erosive lower contact s nor channel facies. These observations, coupled with the massive character of the sands and their lateral continuity, suggest that the Nunda and Wiscoy members in Erie County are not channel sands; rather, they could represent sand lobes on a submarine fan (Type II deposit of Mutti and Normark, 1987). The abrupt termination of the Nunda Sandstone Member near West Falls is an important characteristic of other sand lobe terminations (e.g., Cazzola et al, 1985). Significantly, this pinchout does not "grade" into the thin sandstones of the lobe fringe deposits, consistent with Mutti and Normark's (1987) supposition that lobe fringe and sand lobe deposits are caused by two different types of turbidity flows: sand-poor, highly efficient flow and sand-rich, poorly effeicient flows, respectively.

In western New York the Nunda Sandstone Member generally consists of two packets of sandstone beds -- one at the base of the Nunda Sandstone Member and one near (or at) the top of the Nunda Sandstone Member (Fig. 5). The total number of sandstone beds in the Nunda Sandstone Member varies from 1 to 8, and the total thickness of the Nunda Sandstone Member varies from 0 to 38 m (Fig. 6). The mean grain size of the uppermost sand bed

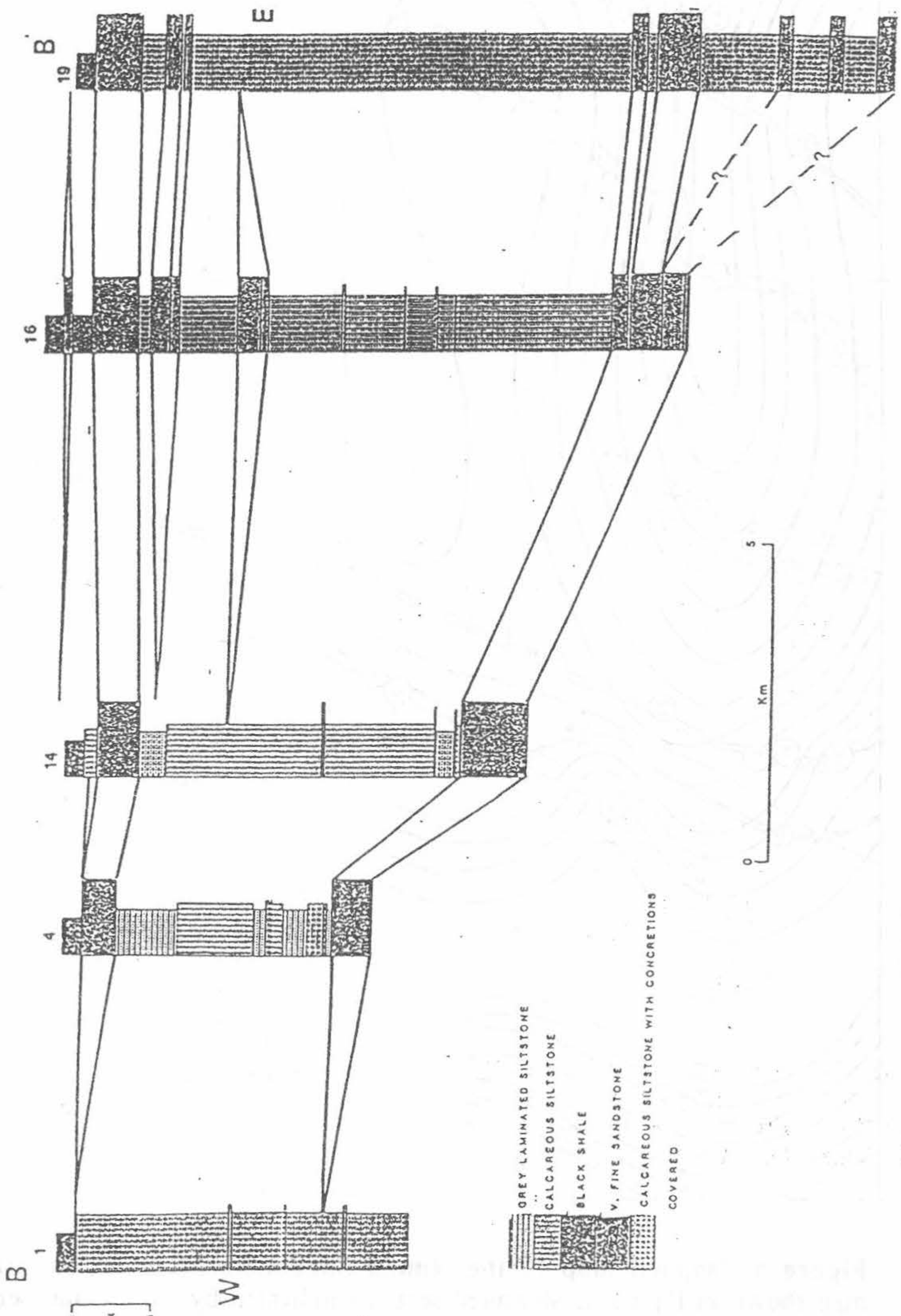


Figure 5. East-West stratigraphic cross section of Nunda Sandstone Member. Location of cross section shown in Figure 2. Figure from Piechocki et al. (1990) and Piechocki (1990).

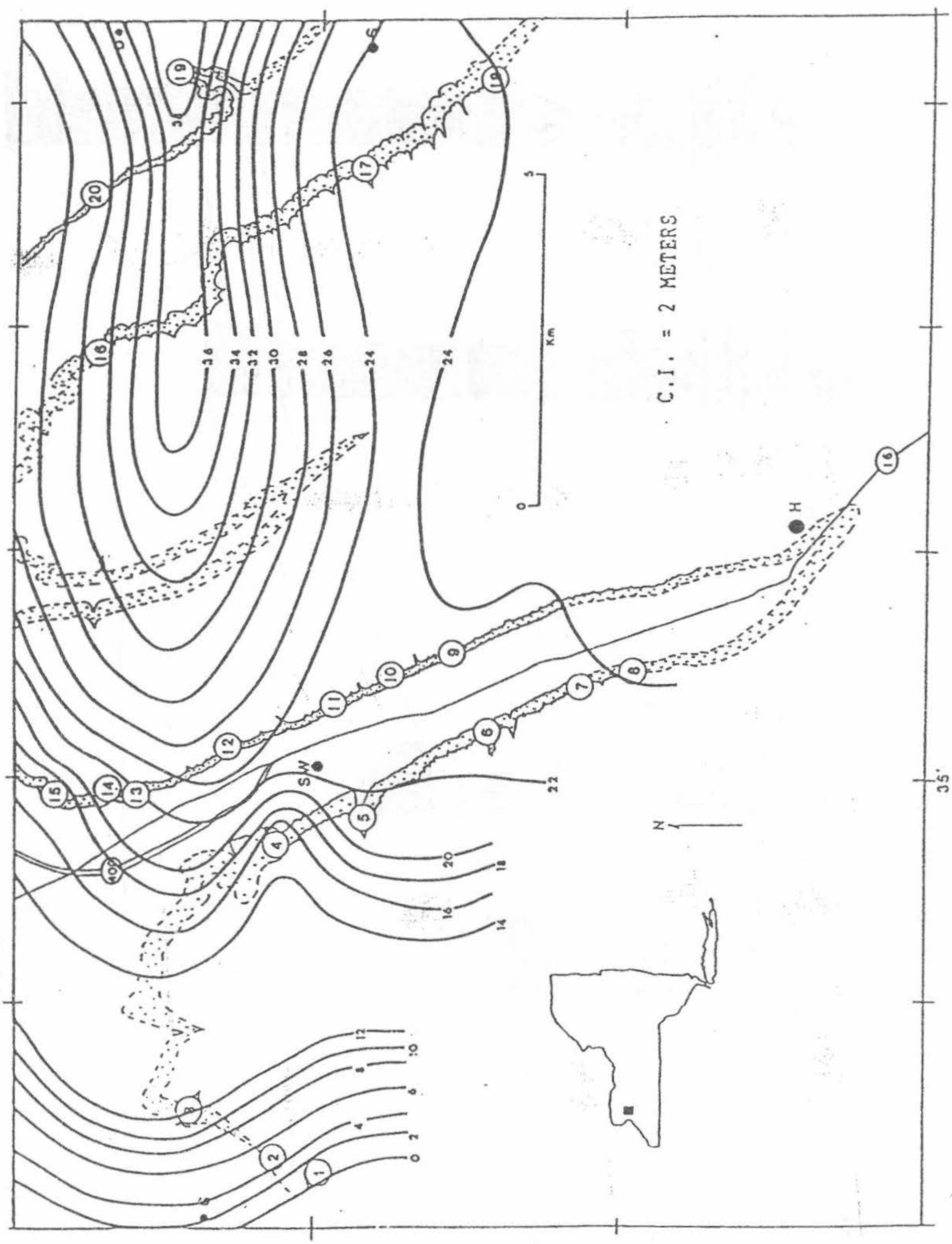


Figure 6. Isopach map of the Nunda Sandstone Member. Location of map shown in Figure 2. Measured sections indicated by circled numbers. G = Griffin's Mills, H = Holland, S = Strykersville, SW = South Wales. Figure from Piechocki et al. (1990) and Piechocki (1990).

varies from 7.5 Ø in the central area to 8.2 Ø at the termination. The overall east-west trend of the grain size is coincident with the zone of thickest Nunda Sandstone Member. The grain size contours display several apparent lobes near the Nunda Sandstone Member termination, and the grain size dramatically fines in the area where the uppermost sand abruptly terminates. There are very few recognizable flow directional indicators in the Nunda Sandstone Member; the few grooves that we have observed in the western sections are transverse or highly oblique to the local isopach contours for the upper and lower sandstone packets. Vertical escape burrows are common in the sandstones of the Nunda Sandstone Member (e.g., Hasiotis and Piechocki, 1990), and indicate very rapid deposition.

ACKNOWLEDGMENTS

This research was partly supported by NSF grant EAR 8904713 awarded to RDJ, SOC and CM.

REFERENCES CITED

Buehler, E. J., and Tesmer, I. H., 1963, Geology of Erie County: Buffalo Society of Natural Sciences Bulletin, v. 21, n. 3,118 p.

Caster, K. E., 1934, The stratigraphy and paleontology of northwestern Pennsylvania: Bull. Am. Paleo., v. 21,185 p.

Cazzola, C., Mutti, E., and Vigna, B., 1985, Cengio turbidite system, Italy, *in* Bouma, A. H., Normark, W. R., and Barnes, N. E., eds., Submarine Fans and Related Turbidite Systems: Springer-Verlag, p. 179-184.

Chan, M. A., and Dott, R. H., Jr., 1983, Shelf and deep-sea sedimentation in Eocene forearc basin, Western Oregon-fan or non-fan?: Am. Assoc. Pet. Geol. Bull. v. 67, p. 2100-2116.

Clarke, J. M., 1897, The stratigraphic and faunal relations of the Oneonta sandstones and shales, the Ithaca and portage groups in central New York: New York State Mus., Annual Report 49, n. 2, p. 27-81.

Clarke, J. M., 1903, Classification of New York series of geologic formations: New York State Museum Handbook (1st ed.), 19 p.

Clarke, J. M., and Luther, D. D., 1908, Geologic map and descriptions of the Portage and Nunda quadrangles: New York State Mus. Bull. 118, 88 p.

Craft, J. H., and Bridge, J. S., 1987, Shallow-marine sedimentary processes in the Late Devonian Catskill Sea, New York State: Geol. Soc. Am. Bull. v. 98, p. 338-355.

Gutmann, M. P., 1989, Upper Devonian turbidites in the South Wales Shale Member of the Perrysburg Formation: Unpub. MA thesis, SUNY at Buffalo, Buffalo, N.Y., 185 p.

Gutmann, M. P., and Jacobi, R. D. 1988. The Devonian South Wales Member: A suprafan lobe? Geol. Soc. Am., Abstr. with Programs, v. 20, n. 1, p. 24.

Hampton, M. A., and Colburn, I. P., 1967, Directional features of an experimental turbidite fan: A report of progress: Jour. Sed. Pet., v. 37, p. 509-513.

Hartnagel, C. A., 1912, Classification of the geologic formations of the state of New York: New York State Mus. Handbook 19, 2d ed.

Hasiotis, S. T., and Piechocki, A. L., 1990, Preliminary report on the fauna, flora, and ichnofauna content of the Nunda sandstone: Geol. Soc. Am., Abstr. with Programs, v. 22, n. 2, p. 22-23.

Jacobi, R. D., Gutmann, M., Piechocki, A., Singer, J., O'Connell, S., Mitchell, C., in press, Upper Devonian turbidites in western New York: preliminary observations and implications, *in* Landing, E., ed., Studies in Paleontology and Stratigraphy in Honor of Donald W. Fisher: New York State Geol. Surv.

Johnson, J. G., Klapper, G., and Sandberg, C. A., 1985, Devonian eustatic fluctuations in Euramerica: Geol. Soc. Am., v. 96, p. 567-587.

Kirchgasser, W. T., and House, M. R. ,1981, Upper Devonian goniatite biostratigraphy, *in* Oliver, W. A. Jr. and Klapper, G. eds., Devonian Biostratigraphy of New York, Part I, Text: International Union of Geological Sciences, Subcommission on Devonian Stratigraphy, Washington, D. C., p. 57-66.

Lundegard, P. D., Samuels, N. D., and Pryor, W. A, 1985, Upper Devonian turbidite sequence, central and southern Appalachian basin: Contrasts with submarine fan deposits, *in* Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geol. Soc. Am., Sp. Paper 201, p. 107-122,

McGhee, G. R., Jr., and Sutton, R. G., 1981, Late Devonian marine ecology and zoogeography of the central Appalachians and New York: Lethaia, v. 14, p. 27-43.

McGhee, G. R., Jr., and Sutton, R. G, 1985, Late Devonian marine ecosystems of the lower West Falls Group in New York, *in* Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geol. Soc. Am., Sp. Paper 201, p.199-210.

Mutti, E. 1977. Distinctive thin-bedded turbidite facies and related depositional environments in the Eocene Hecho Group (South-central Pyrenees, Spain): Sediment., v. 24, p.107-131.

Mutti, E., and Normark, W. R., 1987, Comparing examples of modern and ancient turbidite systems: problems and concepts, *in* Leggett, J. K., ed.,: Marine Clastic Sedimentology: Kluwer Academics Press, p. 1-38.

Mutti, E., and Ricci Lucchi, F., 1975, Turbidite facies and facies associations, *in* Mutti, E. et al., eds., Examples of Turbidite Facies and Associations from Selected Formations of the Northern Apennines, Field Trip Guidebook A-11: 9th International Association of Sedimentologists Congress, Nice, p. 21-36.

Pepper, J. F., and de Witt, W., 1951, Stratigraphy of the Late Devonian Perrysburg Formation in western and west-central New York: U. S. Geol. Sur., Oil and Gas Investigations, Prelim. Chart OC45.

Pepper, J. F., de Witt, W., and Colton, G. W., 1956, Stratigraphy of the West Falls Formation of Late Devonian age in western and west-central New York: U. S. Geol. Sur. , Oil and Gas Investigations Chart OC55.

Pickering, K. T., 1982, The shape of deep-water siliciclastic systems: A discussion. Geo-Marine Let., v. 2, p. 41-46.

Piechocki, A. L., 1990, The Nunda Sandstone Member of the West Falls Formation: a possible submarine fan sand lobe. Unpub. MA thesis, SUNY at Buffalo, Buffalo, N.Y., 136 p.

Piechocki, A. L., Jacobi, R. D., and Hasiotis, S. T., 1990, The Nunda Sandstone: A Devonian non-channelized sand lobe of a submarine fan: Geol. Soc. Am., Abstr. with Programs, v. 22, n. 2, p. 62.

Rickard, L. V., 1975, Correlation of the Silurian and Devonian rocks in New York State: New York State Mus. and Science Service, Map and Chart Series no. 24.

Sevon, W. D., and Woodrow, D. L. 1985. Middle and Upper Devonian stratigraphy within the Appalachian basin, *in* Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geol. Soc. Am., Sp. Paper 201, p. 1-8.

Sutton, R. G. 1963. Correlation of Upper Devonian strata in south-central New York, *in* Shepps, V. C., ed., Symposium on Middle and Upper Devonian Stratigraphy of Pennsylvania and Adjacent States: Pennsylvania Geol. Sur. Bull. G39, p. 87-101.

Van Tassel, J. ,1987, Upper Devonian Catskill Delta margin cyclic sedimentation: Brallier, Scherr, and Foreknobs Formations of Virginia and West Virginia: Geol. Soc. Am. Bull. v. 99, p. 414-426.

van Tyne, A. M., 1982, Subsurface expression and gas production of Devonian Black Shales in western New York, *In* Buehler, E. J., and Calkin, P. E., eds., Guidebook for Field Trips in Western New York, Northern Pennsylvania and Adjacent, Southern Ontario: New York State Geol. Assoc., Hofstra University, Hempstead, N.Y., p. 371-385.

van Tyne, A. M., 1983, Natural gas potential of the Devonian black shales of New York: Northeastern Geol., v. 5, p. 209-216.

Walker, R. G. 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. Journal of Sedimentary Petrology 37:25-43.

Walker, R. G., and Sutton, R. G., 1967, Quantitative analysis of turbidites in the Upper Devonian Sonyea Group, New York: Jour. Sed. Pet., v. 37, p. 1012-1022.

Woodrow, D. L., 1985, Paleogeography, paleoclimate, and sedimentary processes of the Late Devonian Catskill Delta, *in* Woodrow, D. L., and Sevon, W. D., eds., The Catskill Delta: Geol. Soc. Am., Sp. Paper 201, p. 51-64.

Woodrow, D. L., and Isley, A. M., 1983, Facies, topography, and sedimentary processes in the Catskill Sea (Devonian), New York and Pennsylvania: Geol. Soc. Am., v. 94, p. 459-470.

Road Log

(see Fig. 7)

Mileage		
Interval	*Cumulative*	
0.00	0.00	Start mileage at NY Thruway Exit 57 Toll Barrier (Eden, NY).
0.35	0.35	T-intersection of Thruway exit ramp and Eden-Evans Center Road. Turn left onto Eden-Evans Center Rd. (this road is called Church Street in the town of Eden).
3.25	3.60	Intersection of Church St. and Route 62. Continue straight on Church St.
1.35	4.95	Intersection of Church St. and Jennings Road. Continue straight on Church St.
1.30	6.25	T-intersection of Church St. and Route 75. Turn right onto Rte. 75
1.00	7.25	Bridge crossing the South Branch of 18 Mile Creek. Excellent exposures of Java Formation shales on the left.
0.30	7.55	Intersection of Rte. 75 with New Oregon Road. Turn left onto New Oregon Rd.
0.30	7.85	Bridge crossing the South Branch of 18 Mile Creek. The contact between Java Formation shales and the overlying Dunkirk Shale Member of the Canadaway Formation is exposed near the base of the section to the left. For permission to inspect outcrop, ask owners at house along the creek (driveway to the west of bridge)
0.48	8.33	Intersection of New Oregon Rd. and Clarksburg Road. Proceed straight on New Oregon Rd.
0.07	8.40	Waterfall in creek to the right of road. Falls is in the Dunkirk Shale. In periods of summer overgrowth, the falls can be viewed from the bridge on Clarksburg Rd., 0.07 mi. back.
0.75	9.15	**Stop 1** Creek bed near a sharp turn to left on New Oregon Rd. For permission to view outcrop, ask owners at the second house past barn along the turn and also ask at the house on the left past the private bridge on the driveway to the outcrop (Fig. 8)

Figure 7. Generalized route of field trip. Note that U.S.G.S. base map is outdated, and does not show the modern route of U.S. 219.

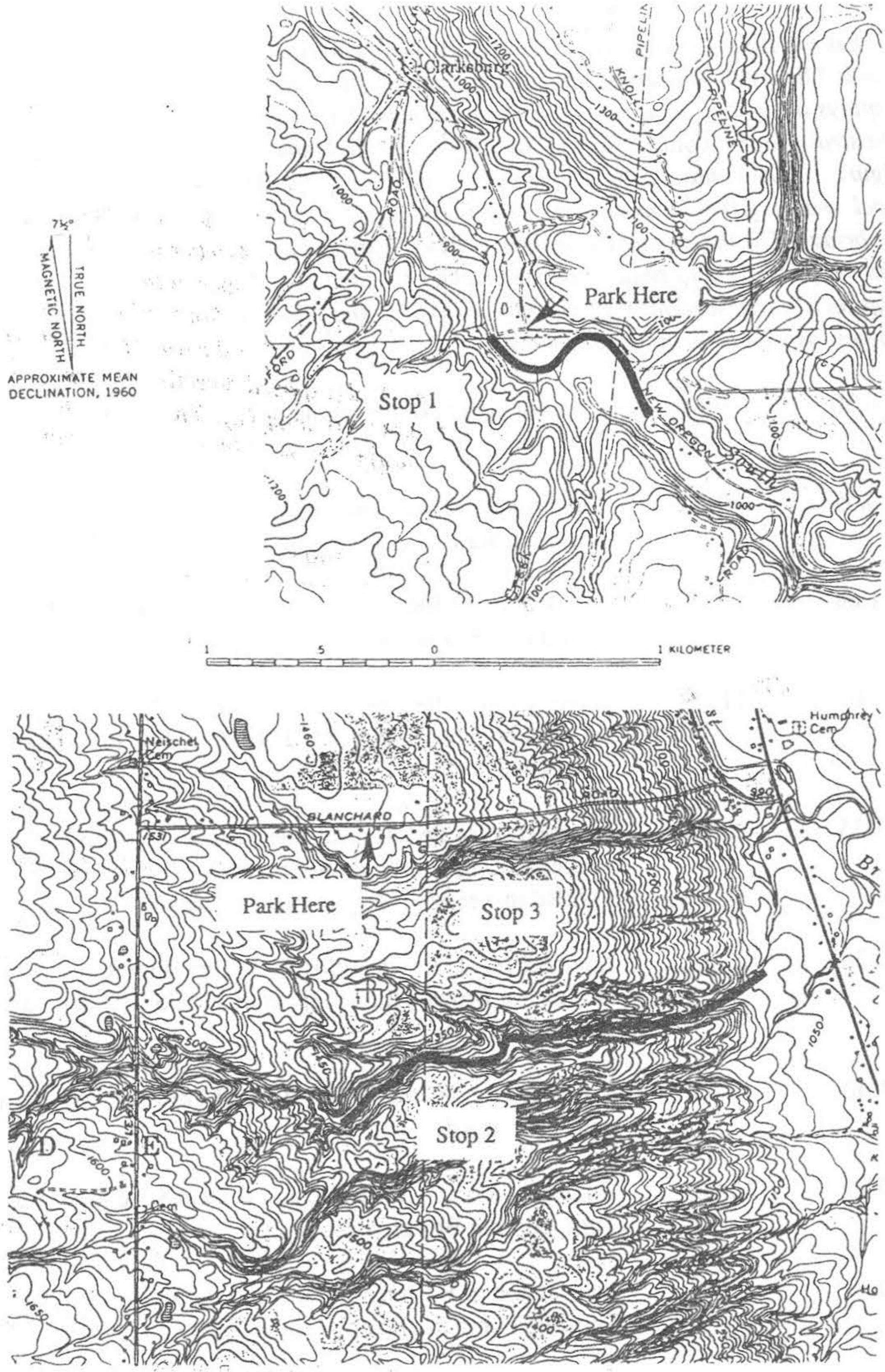


Figure 8. Topographic maps displaying locations of stops 1, 2 and 3.

Walk down the driveway toward the private bridge over the South Branch Eighteen Mile Creek. The creek bed directly below the bridge, as well as the creek bed upstream, (south) displays a "typical" exposure of the South Wales Shale Member of the Perrysburg Formation. The stratigraphic column for the South Wales Shale Member at this locality is portrayed in Figure 3. The contact between the basal sandstone of the South Wales Shale Member and black shales of the Dunkirk Shale Member is exposed at the waterfall north (downstream) of the bridge. The upper surface of the basal bed (exposed directly beneath the bridge) displays prominent straight to sinuous-crested ripples (Bouma Tc). The sandstone bed that forms the cap rock of the waterfalls upstream (south) from the bridge also shows prominent ripples, as well as orthogonal grooves.

0.30	9.45	View of upper portion of the South Wales Shale Member of the Perrysburg Formation on the right.
0.05	9.50	Intersection of New Oregon Rd. and Belcher Road. Turn left onto Belcher Rd.
1.68	11.18	Intersection of Belcher Rd. and paved Boston Road. Continue straight on Belcher Rd.
0.02	11.20	Intersection of Belcher Rd. and Feddick Road. Continue straight on what is now called Feddick Rd. (Feddick Rd. makes a 90° turn at this intersection).
1.55	12.75	Intersection of Feddick, Zimmerman, and Brown Hill Roads. Proceed straight through intersection on Brown Hill Rd. (Brown Hill Rd makes a 90° turn at this intersection).
0.30	13.05	Intersection of Brown Hill Rd. and Emerling Road. Continue straight on Brown Hill Rd. (pass under US 219 immediately after intersection).
1.40	14.45	T-intersection of Brown Hill Rd. and Trevett Road. Turn left onto Trevett Rd.
0.55	15.00	T-intersection of Trevett Rd. and Boston State Road (the old 219). Turn left onto Boston State Rd.
0.30	15.30	Intersection of Boston State Rd. and the Boston Colden Road. Turn right onto the Boston Colden Rd.
1.09	16.39	Y-intersection of the Boston Colden Rd. and Cole Road. Bear to right on the Boston Colden Rd.

1.96	18.35	Intersection of the Boston Colden Road with Route 240 in the center of the town of Colden. Proceed straight across this intersection to the bridge on Heath Rd. (the Boston Colden Rd. becomes Heath Rd. east of this intersection).
0.05	18.40	View of waterfall on right from bridge. This falls is also in the Dunkirk Shale Member of the Perrysburg Formation.
1.50	19.90	T-intersection of Heath Rd. and Hayes Hollow Road (Irish Road). Turn left onto Hayes Hollow Rd.
0.55	20.45	Intersection of Hayes Hollow Rd. and Partridge Road. Turn right onto Partridge Rd.
1.65	22.10	Intersection of Partridge Rd. and Center Road. Continue straight on Partridge Rd.
1.45	23.55	Intersection of Partridge Rd. and Lewis Road. Turn left onto Lewis Rd.
1.60	25.15	Intersection of Lewis Rd. and Blanchard Road. Turn right onto Blanchard Rd.
0.72	25.87	**Stop 2** For permission to view outcrop, ask owners at: 10983 Blanchard and also at two homes at Route 16 (see road log below and Fig. 8).

This stop consists of much of the ravine (indicated on Figure 8) south of Blanchard Road. We will walk south from the parking site to the ravine via farmers' roads, then logging roads, and finally deer trails. We will walk down (east) the ravine and come out on Rte 16. This stop displays a complete section from Nunda Sandstone Member at the base near Route 16 to South Wales Shale Member near the top of the ravine. We will hand out a measured section of this locality on the field trip. In the South Wales Shale Member section, notice that the basal sandstone is no longer prominently rippled, unlike the same sand at stop 1 to the west; rather, it is massive to planar bedded (Bouma Ta/b). The transition from rippled to massive/planar occurs between this section and that exposed at optional stop 3. In the Nunda Sandstone Member note the escape burrows (vertical worm burrows) and undulatory bedding

0.13	26.00	**Stop 3** (optional) To view outcrop, ask owners' permission at 11025 1/2 Blanchard Rd. The owners need proof of self insurance or a signed statement releasing them of all responsibility (Fig. 8).

Walk south from the A-frame house (west of the owner's house) to the waterfall in the ravine that is closest to the Blanchard Road. This stop displays a fairly complete section of the South Wales Shale Member. Although we are about 22 km east-northeast of stop 1, the general appearance of the sandstone beds in the South Wales Shale Member remains the same as that at stop 1, including the prominently rippled basal sandstone.

1.00	27.00	Intersection of Blanchard Rd. and Rte 16. Turn right onto Rte 16.
0.40	27.40	Residence of landowner on right for permission to view Stop 2.
0.10	27.50	Residence of landowner #3 on right for permission to view Stop 2 Reverse direction at this point, returning to Blanchard Rd.
0.40	27.90	Intersection of Rte. 16 and Blanchard Rd. Turn left onto Blanchard Rd.
1.80	29.70	Intersection of Blanchard Rd. and Lewis Rd. Turn right onto Lewis Rd.
0.80	30.50	Intersection of Lewis Rd. and Darien Road. Turn left onto Darien Rd.
1.40	31.90	Intersection of Darien Rd. and Center Road. Turn right onto Center Rd.
3.00	34.90	Intersection of Center Rd. and Blakely Corners Road. Turn left onto Blakely Corners Rd.
1.23	36.13	Intersection of Blakely Corners Rd. and Mill Road. Turn left onto Mill Rd.
0.07	36.20	**Stop 4** (optional) For permission to view outcrop, ask owners at house adjacent to creek section (1437 Mill Rd). The owners need proof of self insurance or a signed statement releasing them of all responsibility (Fig. 9).

This stop displays the upper sandstone of the Nunda Sandstone Member along the north wall of the ravine immediately west of the road. Note that its appearance (e.g., thickness) has not changed significantly from the Nunda exposed at stop 2, some 10 km to the southeast. However, this sandstone bed abruptly pinches out about 3 km south of here (in a ravine 2 km south of here the sandstone is still a meter thick, but in an adjacent ravine at Pipe Creek 4 km south of here, the Nunda is nonexistent).

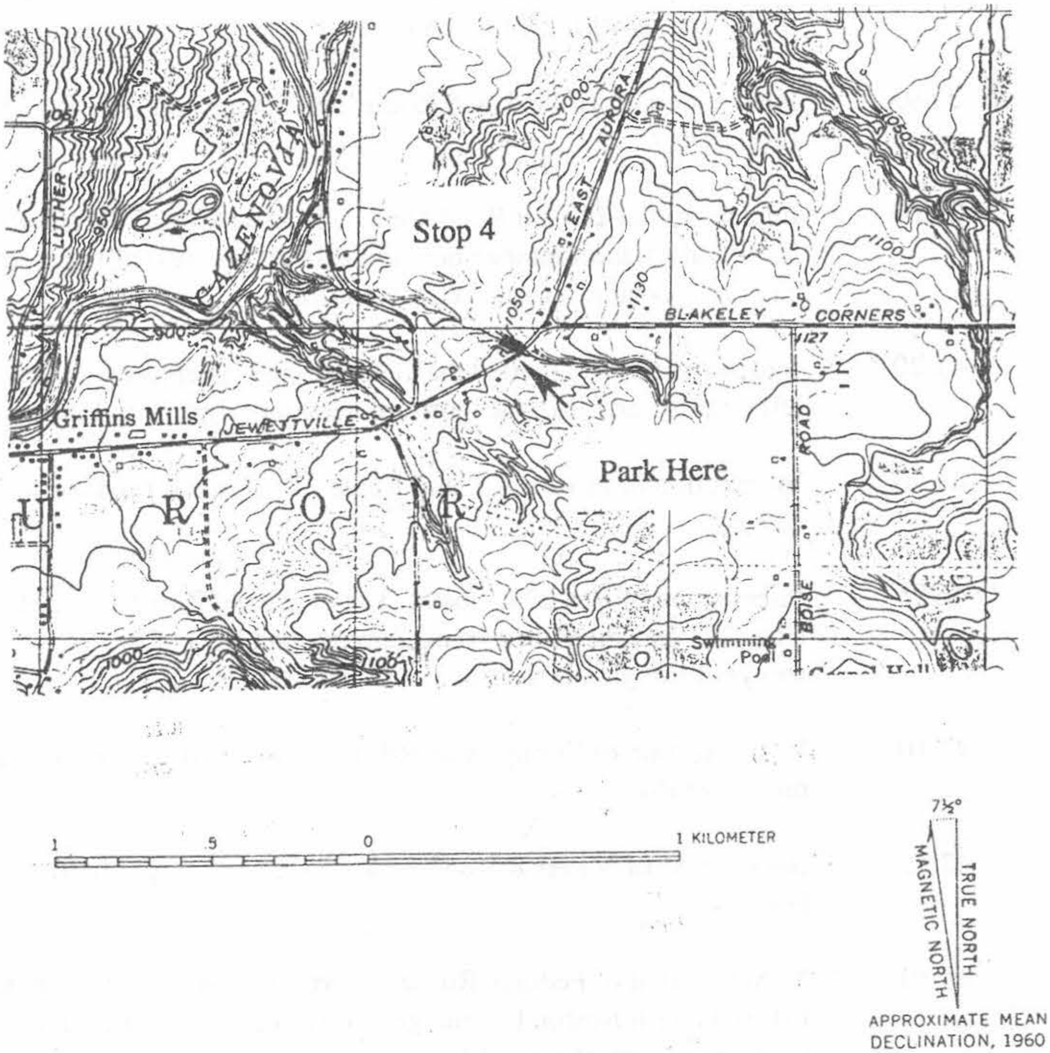


Figure 9. Topographic map displaying location of stop 4.

1.05	37.25	Bridge crossing the West Branch of Cazenovia Creek. Exposures of the Angola Shale Member of the West Falls Formation are seen in the vertical cliff face to the right.
0.70	37.95	Intersection of Mill Rd. and Route 240. Turn right onto Rte. 240 (240 north).
0.65	38.60	Intersection of Ellicott Road and Davis Road (Rte. 240 follows Davis Rd south of intersection and Ellicott Rd west of intersection Turn left, continuing on Rte. 240 "north".
2.60	41.20	Intersection of Rte. 240, Scherff Road and Powers Road. Turn left onto Scherff Rd., then turn immediately right onto Powers Rd
1.00	42.20	Intersection of Powers Rd. and Route 277. Turn left onto Rte. 277
4.15	46.35	Intersection of Rte. 277, Boston State Rd., and Zimmerman Rd (Rte. 277 ends here, becoming Zimmerman Rd.). Proceed straight through intersection on Zimmerman Rd.
0.75	47.10	Y-intersection of Zimmerman Rd. and Mayer Road. Bear to right on Mayer Rd.
0.85	47.95	Intersection of Mayer Rd. and Feddick Rd. Turn right onto Feddick Rd.
0.85	48.80	Y-intersection of Feddick Rd. and North Boston Road (Feddick Rd. and North Boston Rd. merge at this intersection, Feddick Rd. becoming North Boston Rd.).
0.20	49.00	Intersection of North Boston Rd. and Taylor Road. Continue straight on North Boston Rd.
1.35	50.35	Intersection of North Boston Rd. and East Eden Road. Continue straight on North Boston Rd.
1.55	51.90	Intersection of North Boston Rd. and Route 75. Continue straight on North Boston Rd.
0.85	52.75	Intersection of North Boston Rd. and Eden Valley Road. Turn right onto Eden Valley Rd.
0.05	52.80	Intersection of Eden Valley Rd. and Route 62. Turn left onto Rte. 62.

0.20	53.00	Cross bridge over Eden Valley.
2.17	55.17	Intersection of Rte. 62 and Church St. (Eden-Evans Center Rd.). Turn right onto Church St.
3.25	58.42	Intersection of Church St. (Eden-Evans Center Rd) and onramp for NY Thruway Exit 57 (Eden, NY). Turn right for Thruway.
0.35	58.77	NY Thruway Exit 57 Toll Barrier.

BEACH PROCESSES AND COASTAL MORPHOLOGY ALONG A RECURVED SAND SPIT: PRESQUE ISLE, PENNSYLVANIA

KENT TAYLOR and RAYMOND BUYCE
(Mercyhurst College)

INTRODUCTION

Presque Isle is an 11 km long recurved sand spit on the south shore of Lake Erie partially enclosing Pennsylvania's only recreational and commerical shipping port, Presque Isle Bay and the City of Erie (Figure 1). A new $15 million system of detached breakwaters is presently under construction along the lake side of the Presque Isle beaches (Figure 2). The U.S. Army Corps of Engineers and the State of Pennsylvania are sharing the cost of this Shoreline Erosion Control Project. A great diversity of coastal engineering structures have been built along Presque Isle dating from the early 1800's to the present (Figure 3) and beach nourishment has been ongoing since 1955 (see Thomas and others, 1987, p. 33-38). The estimated cost of erosion control including beach nourishment from 1829 is $32 million, of which $20 million was spent since 1975. It is hoped that with the completion of the breakwaters the Presque Isle maintenance and nourishment costs will be reduced by $1 million per year for the next fifty years (personal communication Dale Hamlin, Pennsylvania Bureau of Water Projects, 1990).

The U.S. Corps of Engineers and the State of Pennsylvania have shared the responsibility of protecting Presque Isle from significant erosion. From the U.S. Corps of Engineers perspective the primary goal has been to ensure that the peninsula will continue to protect the Port of Erie. Only recently have the beaches on the lake side of the peninsula played a significant economic role for the City of Erie due to their immense recreational potential. Some conflicts have arisen recently between those who see the primary goal of erosion control projects as maintenance of recreational beaches and others who continue to pursue the long standing goal of harbor protection. Some progress has been made toward addressing both issues. One example is the recent spreading of a fine-grained sand layer over the beach replenishment materials selected for maximum cost effectiveness for erosion control (the so-called dirty sand and gravel taken yearly from upland gravel pits).

Effective erosion control relies on an understanding of the dynamics of the sand spit and preliminary studies have been made, mostly supported by the U.S. Corps of Engineers. The most recent studies include:
(1) The U.S. Army Corps of Engineers Final Phase 1 General Design Memorandum (1980).
(2) An extensive study of a 1:50 scale physical model of the neck of Presque Isle completed at the Waterways Experiment Station at Vicksburg, Mississippi.

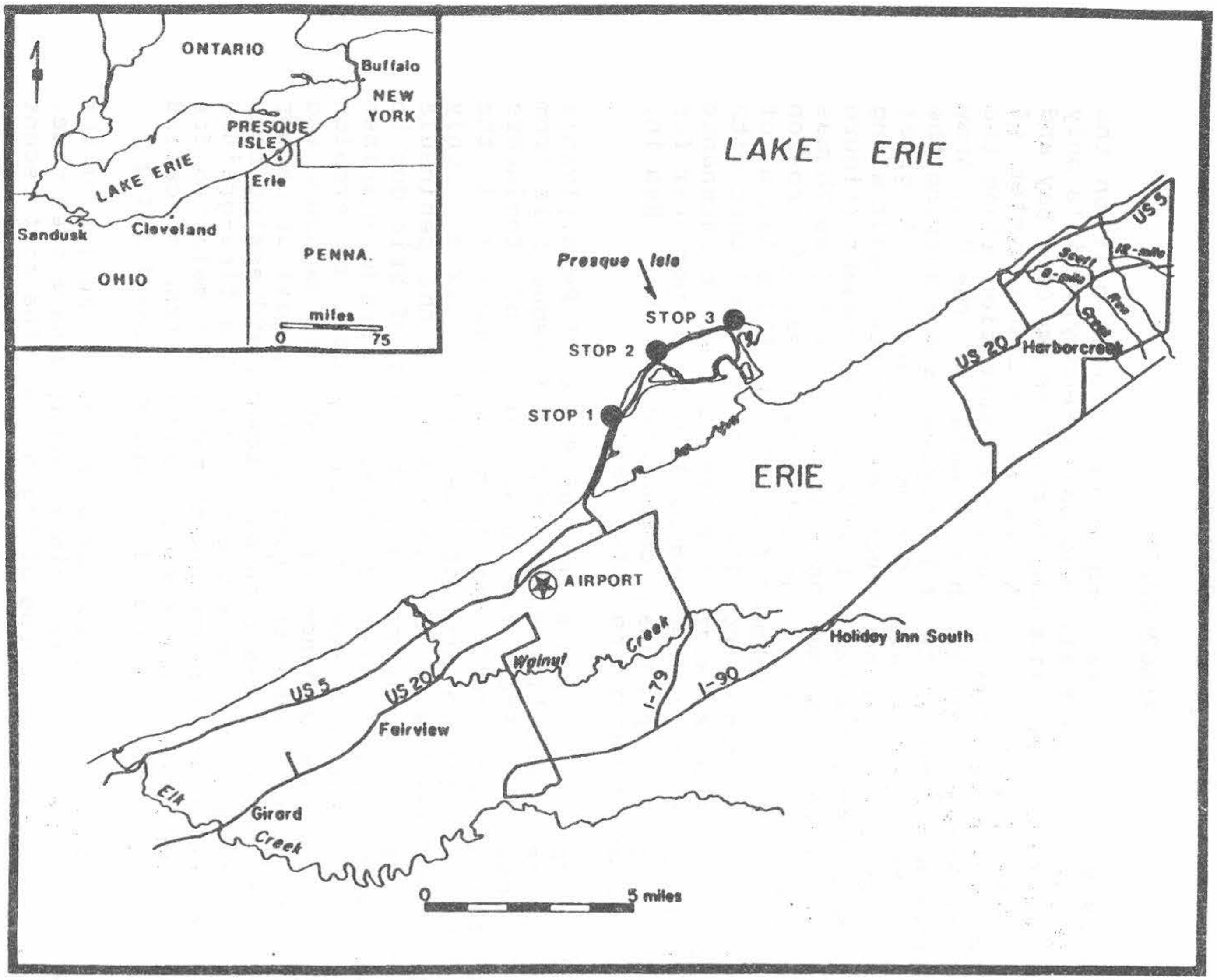


Figure 1. Location map of Presque Isle, PA, showing its position relative to the tri-state area, the route map for the Presque Isle field trip, and the location of the stops at the peninsula.

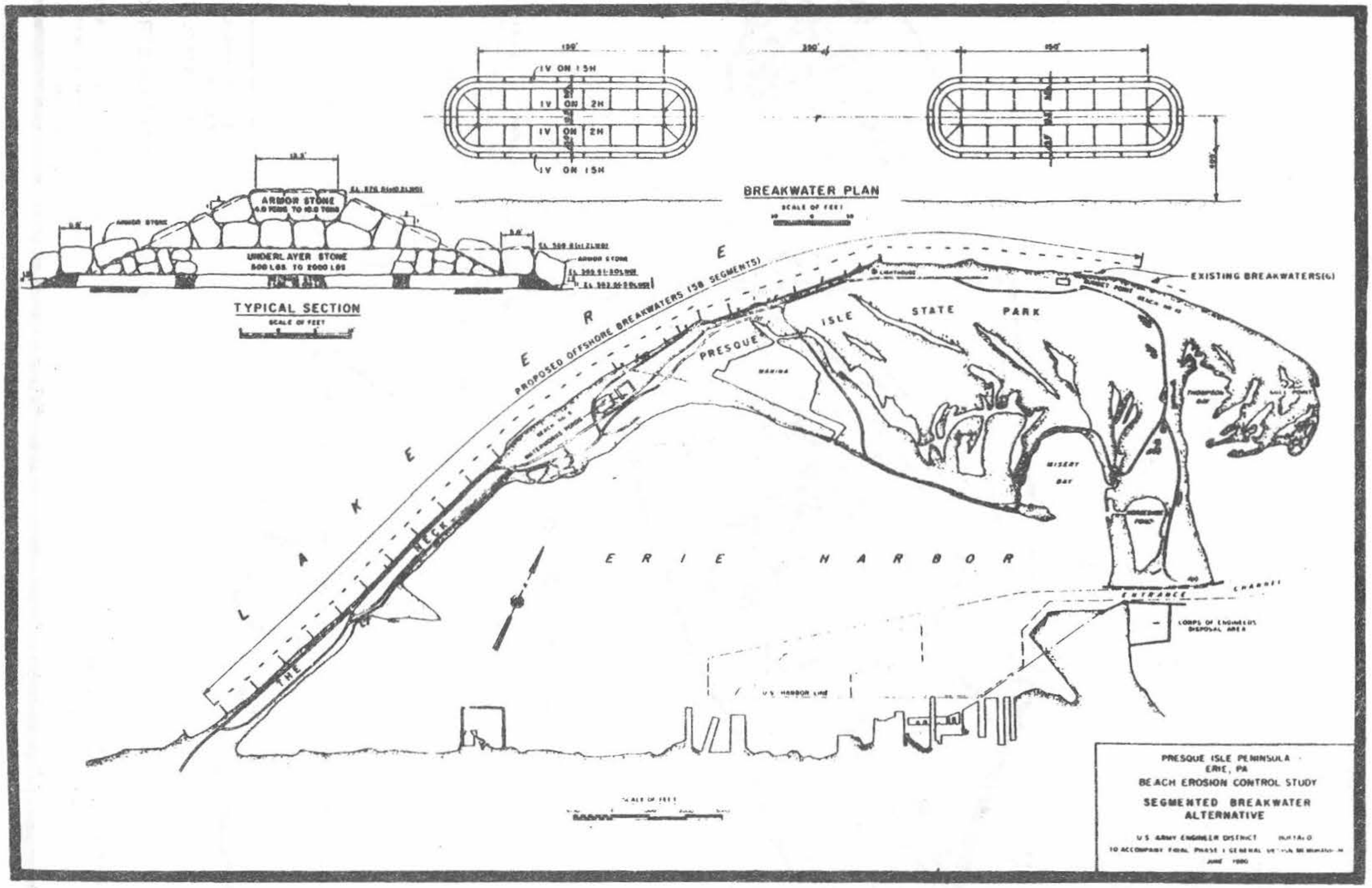


Figure 2. U. S. Army Corps Engineers plan for the 58 offshore breakwater segments designed to protect the peninsula's beaches.

Sat. F3

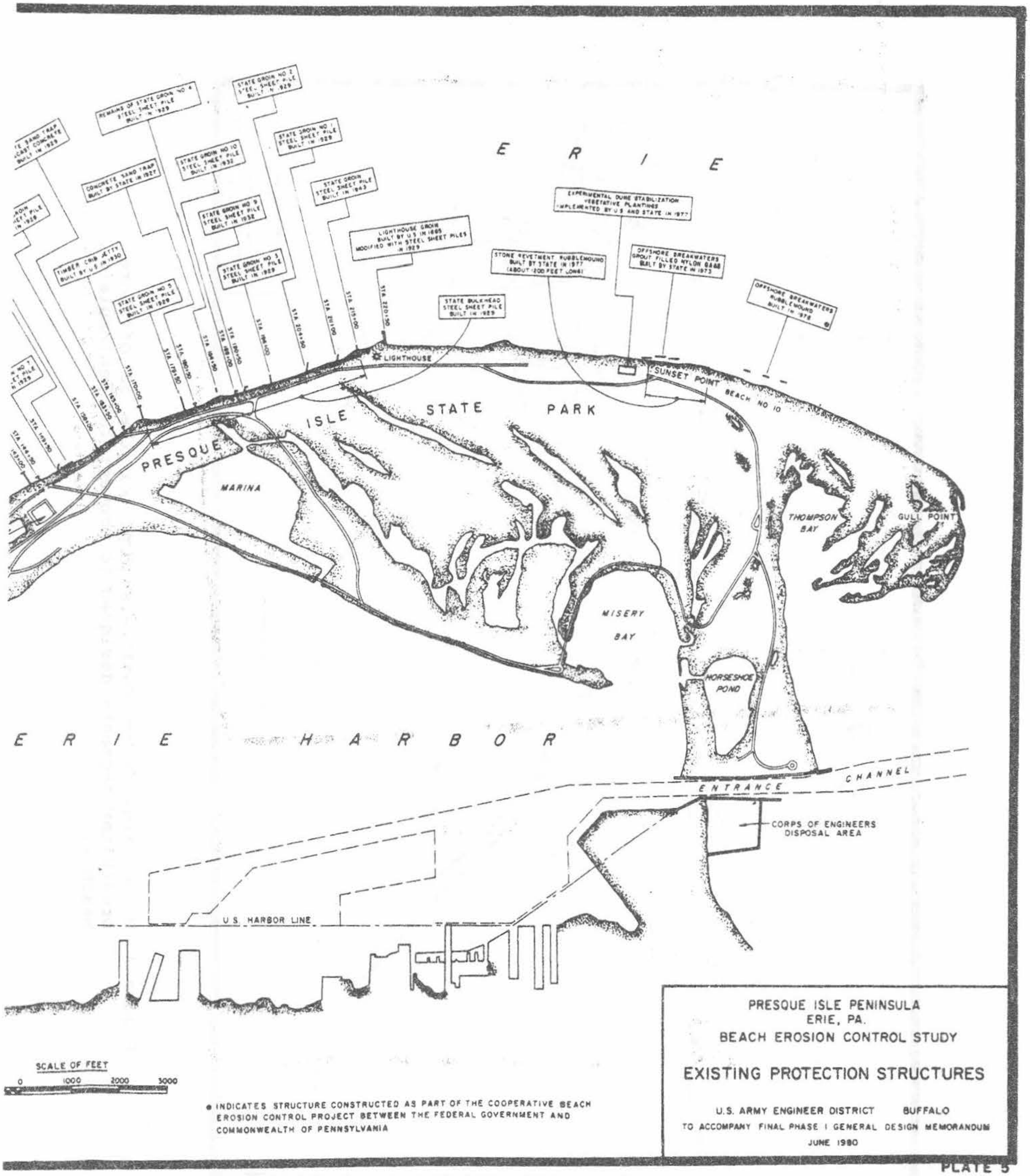
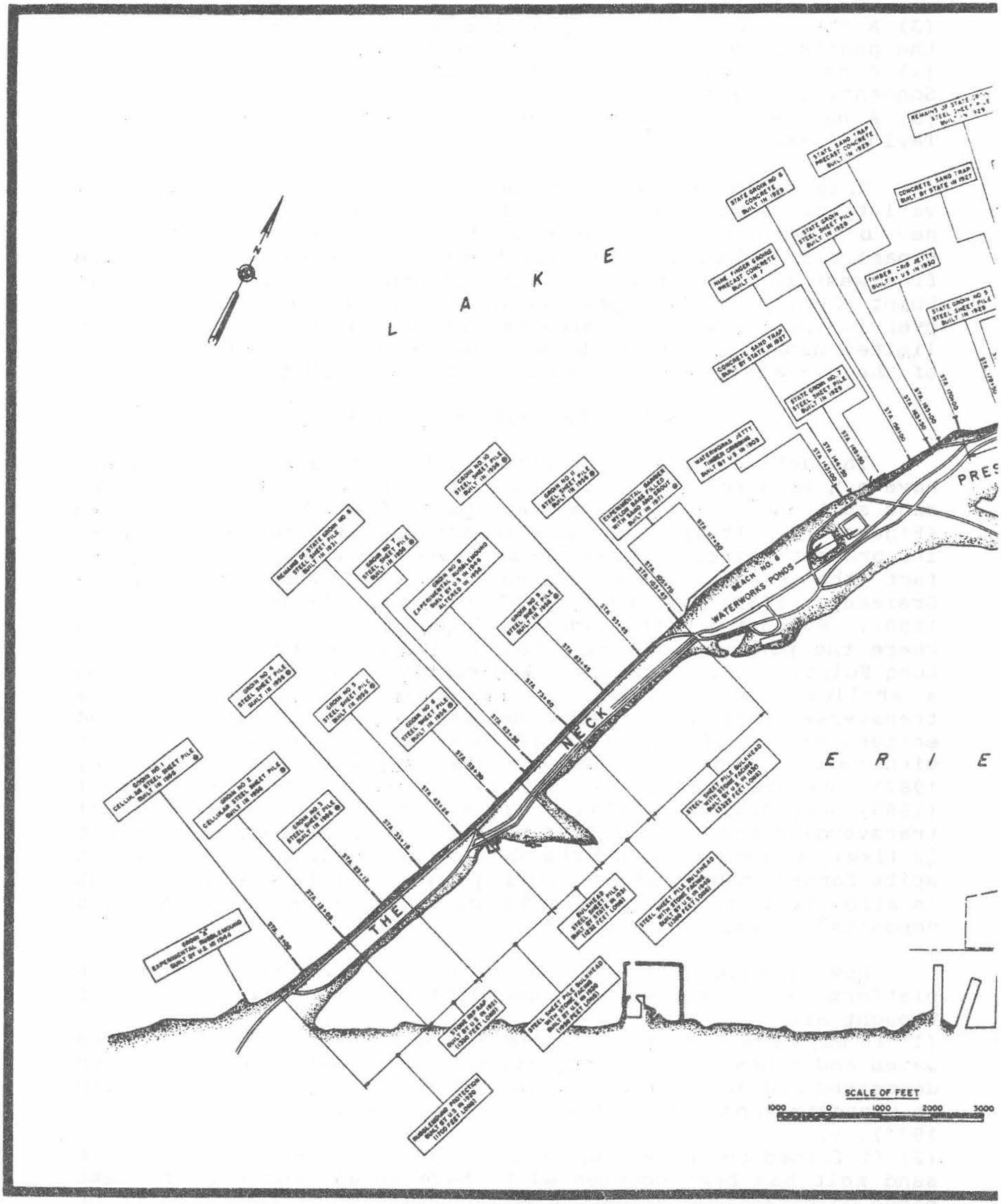


Figure 3. Existing erosion control structures along the Presque Isle shore.

Sat. F4



L A K E
E R I E
THE NECK
PRES
BEACH NO. 6
WATERWORKS PONDS
SCALE OF FEET
1000 0 1000 2000 3000

(3) A three year time study of the beach and bar systems along the peninsula by Nummedal (1979, 1980, 1981).
(4) A native sand tracer study of the inner bar at Beach 6 by Sonnenfeld (1981).
(5) A native sand tracer study of the outer bar at Beach 6 by Taylor (1981, unpublished).

This field trip has been devised to allow us to observe a variety of coastal engineering structures including some of the new breakwaters. One stop each is planned for the three dynamically and morphologically distinct segments of Presque Isle from west to east: the Neck, the Lighthouse Beaches, and Gull Point (Figure 6). The optional overflight will permit a dramatic overview of the entire system helping us to fit the necessarily limited number of ground-based observations into the framework of the overall coastal dynamics of the spit system.

ORIGIN OF PRESQUE ISLE PENINSULA

The details of the formation of the Presque Isle peninsula have not been fully established. It is known that the recurved sand spit now sits on the eastern part of a subaqueous platform (Figure 4). It is reasonable to assume that Presque Isle owes its present location along the southern shore of the lake to the fact the platform was present there (Lewis, 1966, 1969; Dreimanis, 1969; Messinger, 1977; U.S. Army Corps of Engineers, 1980). Support for this idea is found directly across the lake where the peninsula is mirrored by a similar but larger spit, Long Point, Ontario, Canada. Apparently Long Point also sits on a shallow platform that is the other end of a subaqueous transverse ridge that crosses the lake separating the central and eastern basins of Lake Erie (Figure 5). The ridge is covered with wave-sorted sands and gravels (Williams and Meisberger, 1982) and has been identified as a glacial moraine by Lewis (1969) and Dreimanis (1969). Thus the glacial moraine ridge transversing the lake at this particular place along its length is likely to have provided the shallow water platforms upon which spits formed on both sides. During lower lake levels the moraine is also likely to have provided sediments that were reworked and deposited as part of the spits.

How Presque Isle came to be on the eastern part of the platform is the subject of some disagreement. Two schools of thought are:
(1) It may have formed on the eastern portion - Eastward directed waves and longshore currents may have reworked, transported and deposited sediments on the leeward side of the platform producing an elongate sand body that evolved into Presque Isle (Messinger, 1977). or
(2) It formed on the western portion and migrated eastward - The sand spit has been documented to have moved from west to east based on botanical data (e.g. age of trees on different parts of the peninsula) and on charts and surveys dating back to 1730 (Jennings, 1930). Using the rate of migration established from

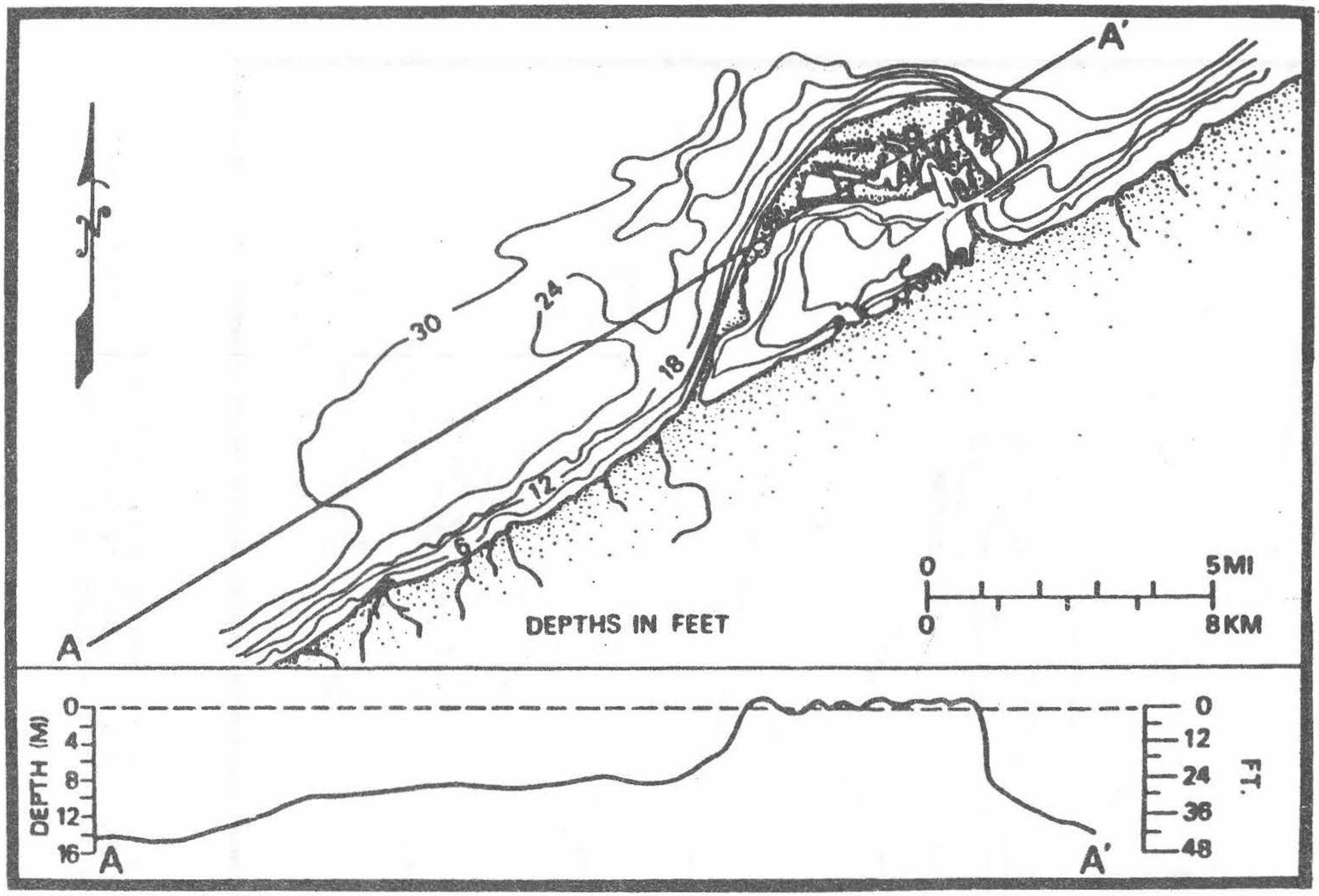


Figure 4. Bathymetry of the lake area surrounding Presque Isle, showing the platform on which the spit is built (from Nummedal, 1983).

Sat. F7

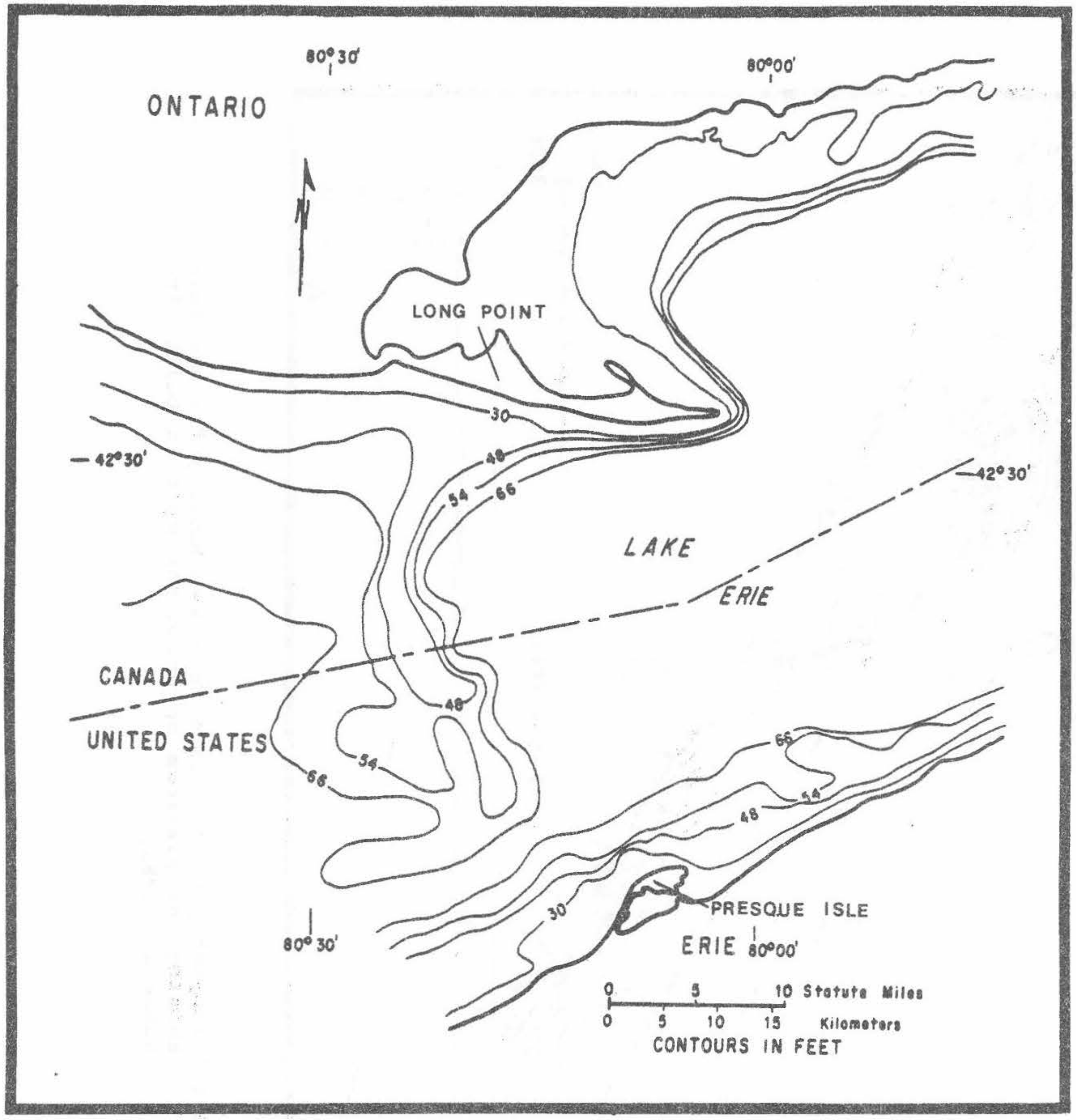


Figure 5. Map of the central Lake Erie basin showing the traverse ridge that connects with Long Point and projects toward Presque Isle (from Williams and Meisburger, 1982).

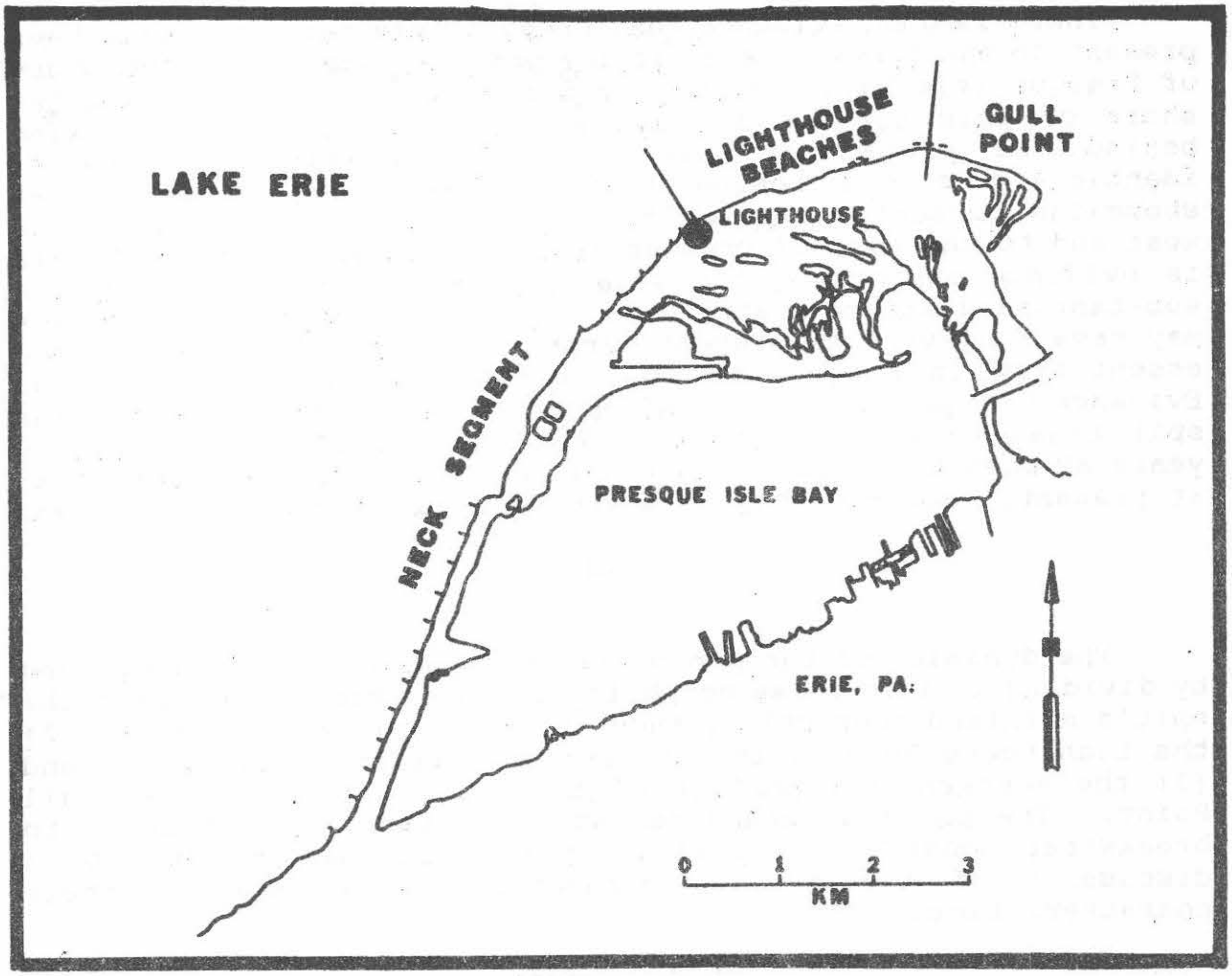


Figure 6. Location of the three major morphologic zones of Presque Isle. The divisions are based on changes in the physiography of the spit and the nearshore zone.

1730 to present to project the location of Presque Isle back 500 years and more ago is not necessarily valid.

There is good evidence that Presque Isle has not always been present on the lakeward side of the present location of the shore of Presque Isle Bay. The wave-cut bluffs that extend along the shore of this portion of lake are A) not displaced lakeward behind the protective shield of the present spit and B) identically eroded behind the spit as along stretches of shoreline presently exposed to lake waves and currents to the west and to the east of Presque Isle. It is not clear that this is evidence supporting the idea that the spit has migrated a substantial distance eastward to its present location; the coast may have achieved its present morphology before the spit formed essentially in place with only a minor amount of migration. Evidence has not yet been obtained establishing the time when the spit began to form; certainly it cannot be older than 12,000 years BP when the level of Lake Erie was some 120 feet lower than at present. How recently it may have begun forming is unknown.

PHYSIOGRAPHY

The dynamics of the peninsula can be more easily understood by dividing it into three morphologic zones (Figure 6). From the spit's mainland connection, they are: (1) the Neck segment, (2) the Lighthouse Beaches in the central portion of the spit, and (3) the eastern-most portion of the spit, referred to as Gull Point. The physical characteristics of each segment prior to breakwater construction will first be described followed by a discussion of the dynamic parameters which generate their characteristics.

Neck Segment

From its connection with the mainland, extending 3.2 km to the northeast, the neck of Presque Isle is narrow and low in relief. Its average width is 245 m and the maximum elevation is 2.5 m above low water datum (LWD for Lake Erie is 173.3 m (568.6 ft) (IGLD, 1955). This segment has a history of severe erosion and has been breached several times by intense storms. Pre-breakwater shore protection structures along this segment include eleven sheet pile groins located at 300 m intervals, several sheet pile bulkheads reinforced with armor stone, and a barrier of grout- and sand-filled nylon bags. An extensive beach nourishment program has also been an important part of the erosion control program since 1955.

Prior to construction of breakwaters along the neck segment the morphology of the nearshore zone included an ephemeral inner bar and a single permanent outer bar, separated by a deep (4.0-4.5 m) through (Figure 3 from Nummedal, 1979). A third poorly developed outer bar was sporadically present along portions of this segment. The inner bar within the groin field was linear or crescentic in plan view. The bar crested in 0.5-1.5 m of water, and was located 10-50 m offshore. The outer bar was located 150-200 m offshore, with the crest of the bar in 3.5 m of water. The outer bar was also crescentic in plan view, with lows or "saddles" separating the 400-1000 m segments. In profile, this bar was typically asymmetric, with a steep landward flank and a gentle lakeward flank. The trough separating the inner and outer bar was typically 4.0-4.5 m deep, giving the outer bar substantial relief.

Lighthouses Beaches

East of the neck, the Lighthouse Beaches segment of the peninsula widens abruptly to more than 1.5 km, and the maximum elevation increases to 6.0 m above low water datum. The orientation of the shoreline changes abruptly at the lighthouse, as the peninsula's shoreline begins to turn landward. The dominant features of the inland portion of the spit include sand plains, east-west trending dune ridges, and ponds or swales occupying the areas between the ridges. Repetition of these features throughout the interior of this segment illustrates the accretionary nature of the spit throughout its evolution. Unlike the neck, pre-breakwater shore protection structures are less regular in spacing and type having been built to stabilize specific pockets of erosion. Groins, stone filled crib jetties, sheet pile bulkheads, rip-rap revetments, detached breakwater segments, and beach nourishment have all been employed to help stabilize the beaches.

Prior to breakwater construction the morphology of the beaches and the nearshore zone changed significantly to the east of the lighthouse. Both the average depth and the slope of the nearshore zone decreased. The outer bar system in this broad platform area consisted of multiple longshore bars, and a system of transverse bars (Figure 7 modified from Nummedal, 1979). The inner bars were usually straight, with their crests located 50 meters offshore in 1.0-1.5 m of water. Megacusps were common features along the beaches of this segment. The megacusps were rhythmic features, with a wavelength of 230 m and an amplitude of 11 m (Nummedal, 1979).

Gull Point

The eastern-most portion of Presque Isle, referred to as Gull Point, is the only depositional area along the peninsula. A series of beach ridges, separated by captured ponds reflects the rapid growth of this segment. This growth occurs by the process of west to east migration of megacusps and welding of offshore bars to the beach along Gull Point (Nummedal, 1978). Although the slopes are much steeper the bar morphology of the nearshore zone along Gull Point is similar to that described for the pre-breakwater neck segment. Bar forms include a single recurved inner bar, and a single high relief outer bar. (Figure 7 from Nummedal, 1979).

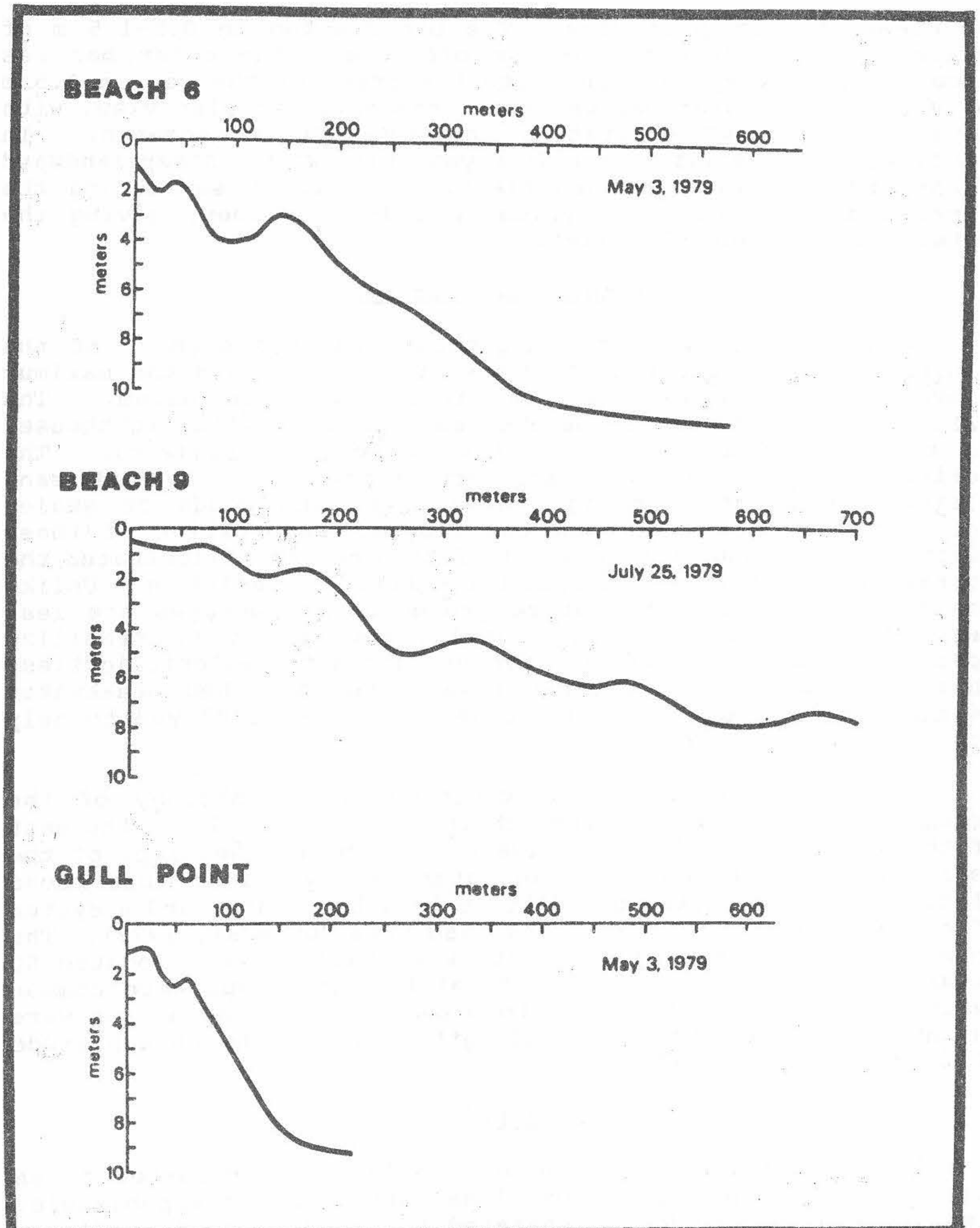


Figure 7. Bathymetric profiles representative of the nearshore bathymetry for each field stop (Beach 6 - neck segment (Stop 1): Beach 9 - lighthouse beaches (Stop 2): Gull Point - Stop 3) (Nummedal, 1979).

WAVE CLIMATE

The wave climate for Lake Erie is best described as a storm wave environment (Nummedal et al., 1976). Long periods of calm are punctuated by short-lived, high energy storm events. Calm conditions, or wave heights less than 0.15 m occur 75% of the time along the Presque Isle shoreline (Nummedal et al., 1984).

The wave climate also has a seasonal aspect. High intensity storms occur in the fall (October and November) before the lake freezes and during the spring (March, April and May) after the lake ice begins to break-up (Nummedal et al., 1974). The summer months are dominated by moderate to low wave energies.

The dominant winds responsible for wave generation on Lake Erie are most commonly a function of extratropical cyclone activity (Nummedal et al., 1976). Low pressure systems generated to the west commonly track eastward along the north shore of Lake Erie (Figure 8A). As the lows pass to the north of the lake, the counter-clockwise circulation causes the wind to shift from the south, through the southwest, to the northwest (Figure 8C). The strongest winds associated with the passage of the lows are from the southwest to west which corresponds to the maximum fetch direction for Lake Erie.

Because the dominant wind direction corresponds to the long axis of the lake, the wave energy flux is also at a maximum from the west-southwest (Nummedal et al., 1974). An annual wave-energy budget based on hindcast wave conditions was calculated by Saville (1953). Figure 9 summarizes the relationship between the fetch length and Saville's calculations for wave energy flux along specified bearings for Erie, PA. The diagram emphasizes that:
(1) the maximum wave power approaches Presque Isle's shoreline from the west and (2) the eastward directed wave energy flux would result in the generation of eastward directed coastal currents.

SEDIMENT TRANSPORT RATES

Nummedal. (1983) has demonstrated that changes in shoreline orientation with respect to the dominant wave approach direction are responsible for increases and decreases in the rate of net sediment transport along Presque Isle. Completion of the detached breakwater project will result in changes that are unpredictable. The rest of this discussion of sediment transport refers to the dynamics and morphology prior to breakwater construction. Figure 10 gives the rates of longshore sediment transport and the rates of net erosion and net accretion for five shoreline segments along the lakeward perimeter of Presque Isle (Nummedal, 1983). The Neck segment of this guide includes Nummedal's I, II, and III segments; the Lighthouse Beaches

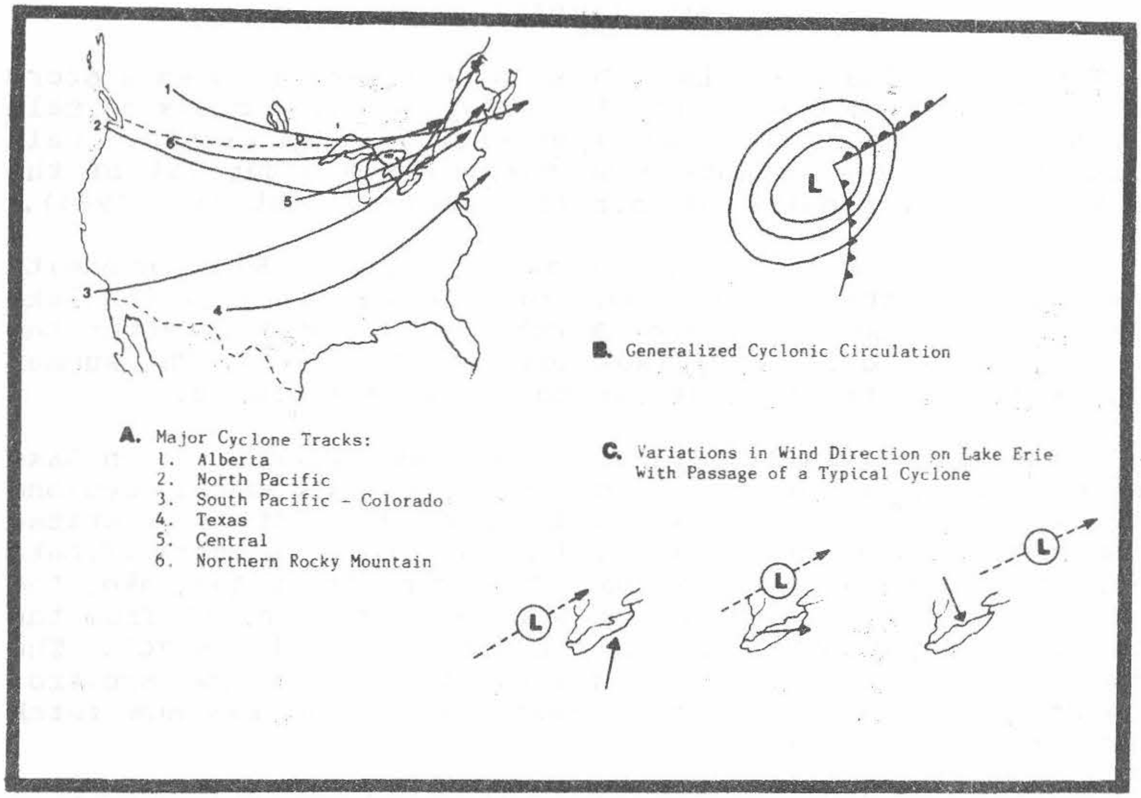


Figure 8. The characteristics and migration of major cyclones effecting the Great Lakes Region.

(A) Major North American Cyclone tracks. Five of six major cyclone tracks converge, and track to the north of Lake Erie. Lows passing to the north of Lake Erie are generally responsible for the generation of storm events. Data from Goode's World Atlas (1970) and Petterson (1969).

(B) Generalized cyclonic pattern. Geostrophic winds are controlled by the equilibrium beetween the pressure gradient and the Coriolis force.

(C) Variation in wind direction on Lake Erie with the passing of a cyclone. Winds change in direction from the south, through west, to northwest as the low tracks to the northeast. (Nummedal et al., 1974)

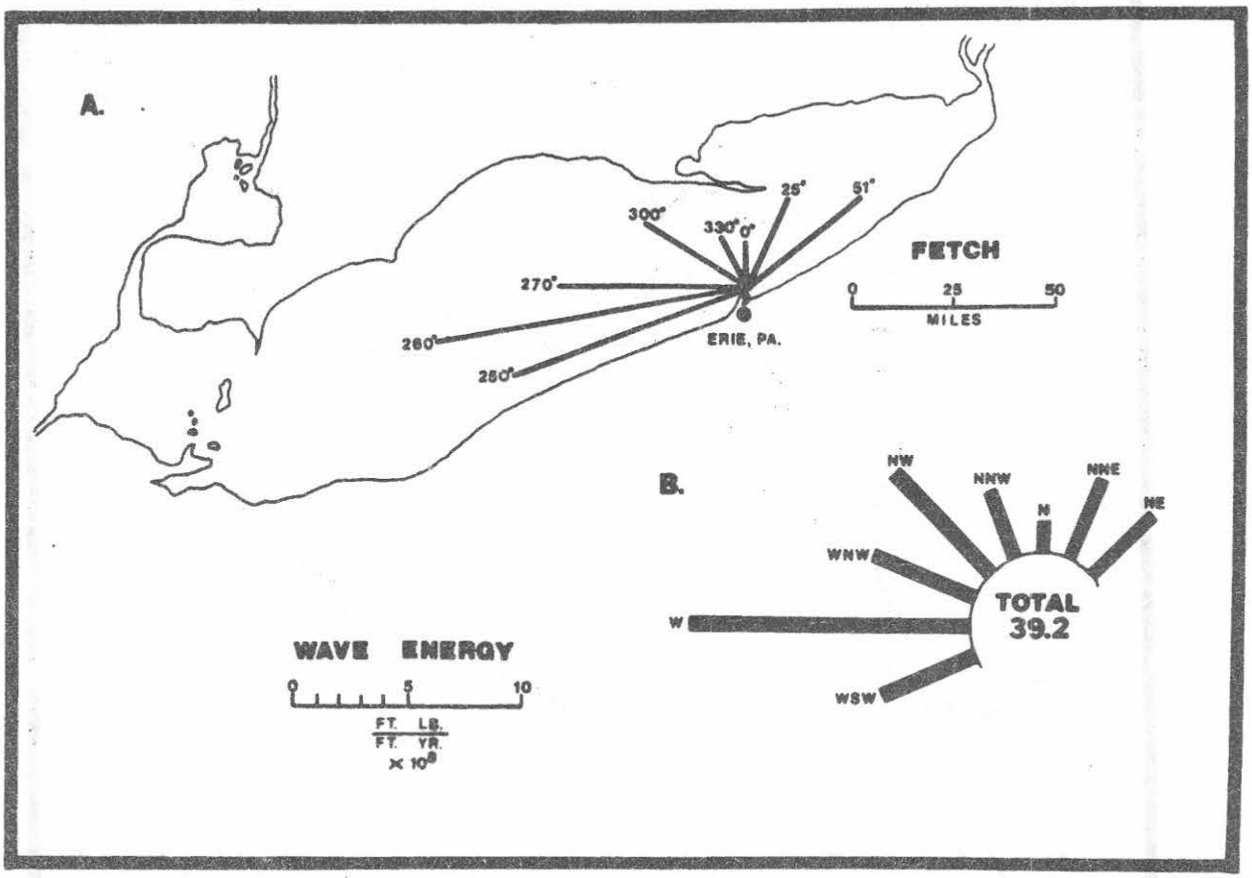


Figure 9. Relationship between fetch and wave energy flux for Erie, PA.

(A) Fetch diagram for Lake Erie with respect to Erie, PA. Bar lengths are proportional to the fetch along the specified azimuths.

(B) Summary of annual wave energy flux for Erie, PA. The principle direction of wave energy flux is from the west. Note the close correlation between the direction of maximum fetch and wave power.

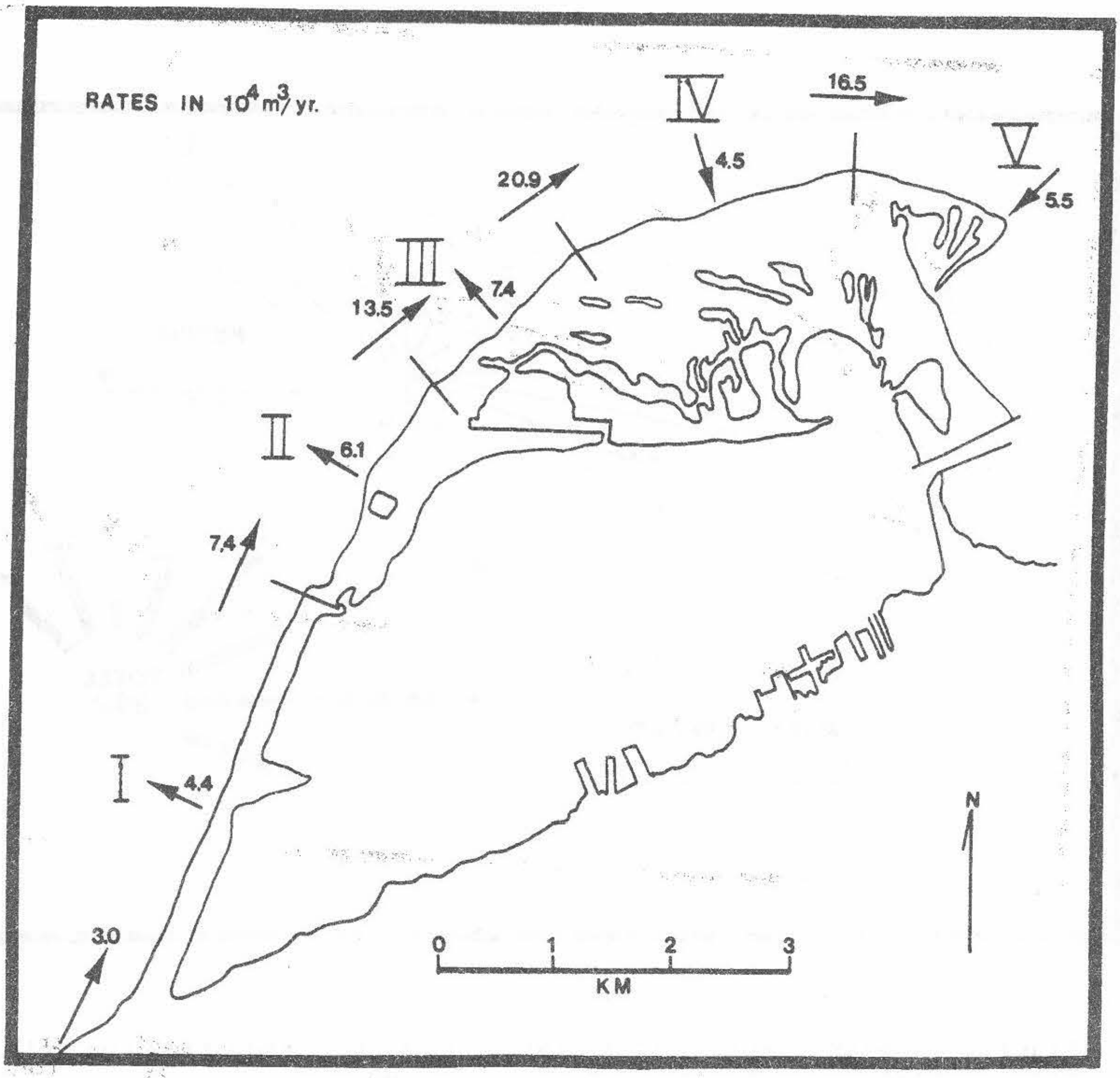


Figure 10. Calculated rates of sediment transport, erosion, and accretion along Presque Isle (Nummedal, 1983). The rates of sediment transport are derived from the longshore distribution of wave power calculated from hindcast wave data, historical accretion rates at Gull Poiont, and Erie Harbor dredging records. The arrows parallel to the shoreline indicate the longshore sediment transport rate for each segment. The arrows perpendicular to the shore line indicate the net loss (arrows directed offshore) or the net gain (arrows directed onshore) of sediment for a given segment. The neck of the Peninsula includes segments I, II and III; the lighthouse beaches correspond to segment IV and Gull Point to segment V.

correspond to his segment IV and Gull Point to segment V. The distribution of wave power along the shore controls the rate of sediment transport and therefore the rate of erosion or accretion along the spit. The differences in the rate of longshore sediment transport are reflected by systematic changes in the characteristics of the beaches and bar systems along the peninsula's shoreline.

The net longshore wave power, and therefore the rate of longshore sediment transport, increases systematically along the Neck of Presque Isle from shoreline segment I to shoreline segment III (Figure 10). The shoreface is eroded as each segment suffers a net loss of sediment. These conditions are reflected in the nearshore zone by: (1) a steep shoreface slope, (2) a single outer bar and trough, and (3) systematic losses of sediment from the beaches (Nummedal, 1983). Dominant storm waves approach segments I through III from the west-southwest and strike normal to the shore producing a strong cell circulation pattern, particularly within the groin field. Sediments are transported offshore by a series of groin-associated rip currents, and alongshore by coastal currents directed to the east. Along these segments 80% of the mean longshore sediment transport occurs within the outer bar-trough system (Nummedal, 1979).

East of the lighthouse (segment III-IV boundary), the orientation of the peninsula's shoreline changes with respect to the dominant wave approach direction. This results in a decrease in the longshore wave power and a decrease in the rate of longshore sediment transport (Figure 10). Consequently, segment IV (Lighthouse Beaches segment) has a positive sediment budget and is accretional through time. Formation of a large nearshore platform within this segment reflects the large influx of sediments from the peninsula's neck to the west (Nummedal, 1983). Associated with the platform is: (1) shallow water depths, (2) a decrease in the shoreface slope, and (3) multiple longshore bars and a transverse bar component (Nummedal, 1983).

Segment V (Gull Point segment) is dominated by net spit progradation (Figure 10). Storm waves from the west-southwest lose their effectiveness as the shoreline curves to the southeast. As a result the longshore energy flux decreases and the segment undergoes a net gain of sediment through time. Sediments carried in the littoral system are deposited along the beach as bar systems migrate landward. The shoreface slope is steep, and a single outer bar may be present close to shore (Nummedal, 1983).

PROGNOSIS

Presque Isle Peninsula is many things to many people but all agree that it is a valuable resource worth protecting. How the 1989-91 Segmented Breakwater Project will function to accomplish that goal is still an open question and the subject of some heated debate. The balancing of priorities among its values as a scientific phenomenon, an aesthetic treasure, a recreational resource, and a protective shield for a Great Lakes port is a formidable task. Those who feel that the sand spit should be allowed to evolve naturally responding freely to the dynamics of wind and wave should remember that the natural geologic feature has been the site of extensive coastal engineering dating back to the 1820's. It is fortunate that we do have the results of some scientific studies referred to herein to serve as a basis for the understanding of the dynamics of the system prior to breakwater construction. Part of the construction cost has been allocated to monitoring studies because no one really knows in detail how the peninsula will respond to this latest engineering project. These and other scientific studies will take place during and after the breakwater installation and we invite everyone to observe carefully today and to keep track of Presque Isle in the years that follow to see what happens.

ROAD LOG

The road log begins at Exit 5: I-79 North-Erie of Interstate 90 in Pennsylvania.

Cumm. Miles	Miles from last point	Description
0.0	0.0	Turn right off I-90 at Exit 5: Interstate 79 North-Erie. Proceed to the end of I-79 (12th Street West).
5.5	5.5	End of I-79. Exit onto 12th Street West - PA Rt. 5 West.
7.1	1.6	Intersection Rt. 832 N, Peninsula Drive leads to Presque Isle State Park. Continue on PA Rt. 5W to Erie Airways at Airport for overflight.
8.3	1.2	Intersection Rt. 299N, Powell Avenue. Continue on PA Rt. 5W.
9.2	0.9	Entrance Erie International Airport on left. Do not turn! Continue on PA Rt. 5W.
9.7	0.5	Turn left (S) on Asbury Road. Turn left again immediately onto Kudlak Drive.
9.75	0.05	Turn left on Kudlak Drive. First building on left is Erie Airways.

OVERFLIGHT DEPARTURE POINT

9.75	0.0	Turn right (S) from Kudlak Drive onto Asbury Road.
9.80	0.05	Turn right (E) onto PA Rt. 5 East.
11.2	1.4	Cross Rt. 299 Powell Avenue. Continue on Rt. 5E.
12.4	1.2	Turn left (N). Intersection Rt. 832 N, Peninsula Drive. Proceed North to Presque Isle Park.
12.7	0.3	Intersection Alt. Rt. 5 West Lake Road. Continue on Rt. 832 N.
12.8	0.1	Intersection West Sixth Street. Continue on Rt. 832 N.

13.4	0.6	New condominiums on left. Pile of large sandstone blocks is remains of 1948 vintage seawall, which was removed during condo construction.
13.5	0.1	Entrance to Presque Isle State Park. Bike path leaves road on right.
13.6	0.1	View of Presque Isle Bay (Erie Harbor) to right.
14.5	0.9	Extensive swampy land.
15.7	1.2	Park headquarters building on right.
16.0	0.3	Niagara boat launch area on right.
16.2	0.2	Turn right into parking area for Cookhouse Pavilion. LUNCH STOP
16.3	0.1	Leave parking lot. Cross median strip and TURN LEFT onto Peninsula Drive.
16.4	0.1	Waterworks on right.
16.8	0.4	Turn right into entrance to Beach No. 6 parking lot.
16.9	0.1	Stop sign. Experimental electric generating windmill on right. Turn left. Go past bathhouse into west parking lot. Proceed to far end of parking lot.
17.0	0.1	STOP 1. NECK SEGMENT
		Leave Stop 1. Retrace route through parking lot. Turn right at windmill and bathhouse.
17.3	0.3	Stop sign at main park road. Cross median and TURN LEFT. Park office is straight ahead.
17.8	0.5	Cookhouse pavilion entrance on right.
18.0	0.2	Swampy area on right, large number of dead trees due to high lake levels during the last several years.

18.1	0.1	Road to Marina and West Pier on right runs along an old dune ridge. It was the old park perimeter road before the 1956 dredging of nourishment material opened the Marina area.
18.4	0.3	Dunes and beach visible to left.
18.7	0.3	GET IN LEFT LANE, TURN LEFT at stop sign.
18.8	0.1	Stop sign. Turn right onto Pine Tree road. Dunes and beaches visible on left, old dune ridges on right.
19.0	0.2	Entrance to park maintenance area on right.
19.3	0.3	Lighthouse (built in 1871) on left. Sidewalk Trail trailhead on right. This concrete paved trail runs along the base of a dune ridge from the lighthouse to the bay shore near the East Boat Livery. It is shown on the 1900 15-minute topographic map, and was originally used as an access route for the lighthouse keepers, who crossed from Erie by boat. It was paved in 1913. The park road originally ran in front of the lighthouse, but was damaged by erosion in 1946, and relocated to its present position in 1948.
19.7	0.4	Stop sign. Road curves right, go straight ahead into parking lot for Beach No. 9.
19.8	0.1	Park in parking lot.

STOP 2. LIGHTHOUSE BEACHES

19.9	0.1	Leave parking lot, turn left at stop sign.
20.6	0.7	Sunset Point area, beach and lake visible to left.
20.8	0.2	Parking lot entrance on left.

20.9	0.1	West end of Budny Beach parking lot, bathhouse on left.
21.0	0.1	Turn left into east section of Budny Beach (Beach No. 10) parking lot. Proceed to far east end of lot.
21.1	0.1	Park in parking lot.

STOP 3. GULL POINT

Leave Stop 3. Retrace route out of parking lot.

21.3	0.2	Leave parking lot. Turn left at stop sign onto Thompson Drive.
21.7	0.4	Thompson Circle loop road on right.
21.8	0.1	Road to Coast Guard Station on left.
21.9	0.1	Entrance to Beach No. 11 on left.
22.2	0.3	Misery Bay on left. Admiral Perry's fleet returned here after the battle of Lake Erie to spend the winter. The Perry Monument is visible on the point across the bay. Niagara Pond is on the right.
22.3	0.1	Lawrence Boat Launch on left.
22.7	0.3	South end of Sidewalk Trail on right.
22.8	0.1	Entrance to East Boat Livery (Park franchised boat rental area).
22.9	0.1	Grave Yard Pond on right.
22.1	0.2	Cross Misery Bay Bridge.
23.2	0.1	Parking area for Perry Monument on left.
23.7	0.5	Bike trail crosses road. Stone rip-rap along bay shore on left. View to left across Erie Harbor to city of Erie.
24.1	0.4	Entrance to East Pier area parking on left, Continue straight.

24.4	0.3	Dredged harbor area on left is location of the park Marina. This area supplied much of the initial beach nourishment material in the 1950's. Range lights on left and right are navigational aids for marking the marina entrance.
24.7	0.3	Bridge over inlet to Long Pond.
24.9	0.2	Bike trail crosses road. Road divides, BEAR RIGHT.
25.0	0.1	Stop sign, turn left onto Peninsula Drive. This is an area of frequent winter washovers and dune migration across road.
25.3	0.3	Junction with old shore road on right.
25.6	0.3	Pettinato Beach entrance on right.
25.9	0.3	Waterworks lagoons on right, picnic area on left.
26.4	0.5	Beach No. 6 entrance on right.
27.4	1.0	Modern restrooms on right (Barracks Beach access).
27.8	0.4	Nature Center on right.
28.4	0.6	Beach No. 1 entrance on right.
28.5	0.1	Leave park.
29.4	0.9	Peninsula Drive and 6th Street traffic light. GO STRAIGHT.
29.5	0.1	Peninsula Drive and 8th Street traffic light. GO STRAIGHT.
29.8	0.3	Peninsula Drive and 12th Street (PA Route 5) traffic light. GET IN LEFT LANE and TURN LEFT.
30.4	0.6	Villa Maria College on left. GET INTO RIGHT LANE.
30.6	0.2	Traffic light at shopping center entrance. GO STRAIGHT.

30.65	0.05	Traffic light at Pittsburgh Avenue. GO STRAIGHT.
30.9	0.25	BEAR RIGHT onto entrance ramp for I-79 South.
36.2	5.3	Junction I-79 and I-90 West. STAY on I-79.
36.6	0.4	Junction I-79 South and I-90 East. TURN RIGHT onto I-90 East.

PRESQUE ISLE OVERFLIGHT: ERIE AIRWAYS, ERIE INTERNATIONAL AIRPORT

The overflight of Presque Isle will set the stage for the ground based field stops and will help you visualize Presque Isle as a dynamic coastal system. The half-hour overflight will be just enough time to make one complete circuit around the peninsula and several circles around the Gull Point area. Before the overflight, familiarize yourself with Figure 11, and review the list of suggested observations provided below.

The Neck Segment (Includes location of Stop 1)

(1) Notice the system of sheet pile groins along the neck and their relationship to beach morphology and orientation.
(2) Notice the orientation of incident waves with respect to the shoreline and look for patterns of wave refraction (wave bending).
(3) Look for evidence of nearshore currents as evidenced by turbidity plumes along the beaches and adjacent to the groins.
(4) Look for nearshore bars if the water is clear or look for linear patterns of surf that may be the result of waves breaking over the crests of nearshore bars.
(5) Notice the characteristics of the backshore area including the discontinuous vegetated dune ridge along the neck.
(6) Notice the newly constructed breakwaters offshore along the neck.

The Lighthouse Beaches (Includes location of Stop 2)

(1) Notice how the orientation of the shoreline changes to the east of the lighthouse.
(2) Look for changes in nearshore processes (wave angle with respect to the shoreline, nearshore currents, etc.).
(3) Look for changes in the morphology of the beaches and notice the presence of megacusps and other rhythmic features.
(4) Notice the beach ridges, ponds and climax forest in the interior of the spit.
(5) Observe the effects of different types of engineering structures on the morphology of the beaches along this segment (e.g. buildup of sediment on the updrift side of structures and erosional bays immediately downdrift).

Gull Point (Includes location of Stop 3

(1) Notice how the shoreline orientation continues to rotate to the southeast.
(2) Look for the orientation of incident waves and note wave refraction patterns.
(3) Look for megacusps and evidence of nearshore bars welding to the beaches.
(4) Notice the newly formed spits at the end of Gull Point.
(5) Try to pick out the location and orientation of relic shorelines and spit complexes.

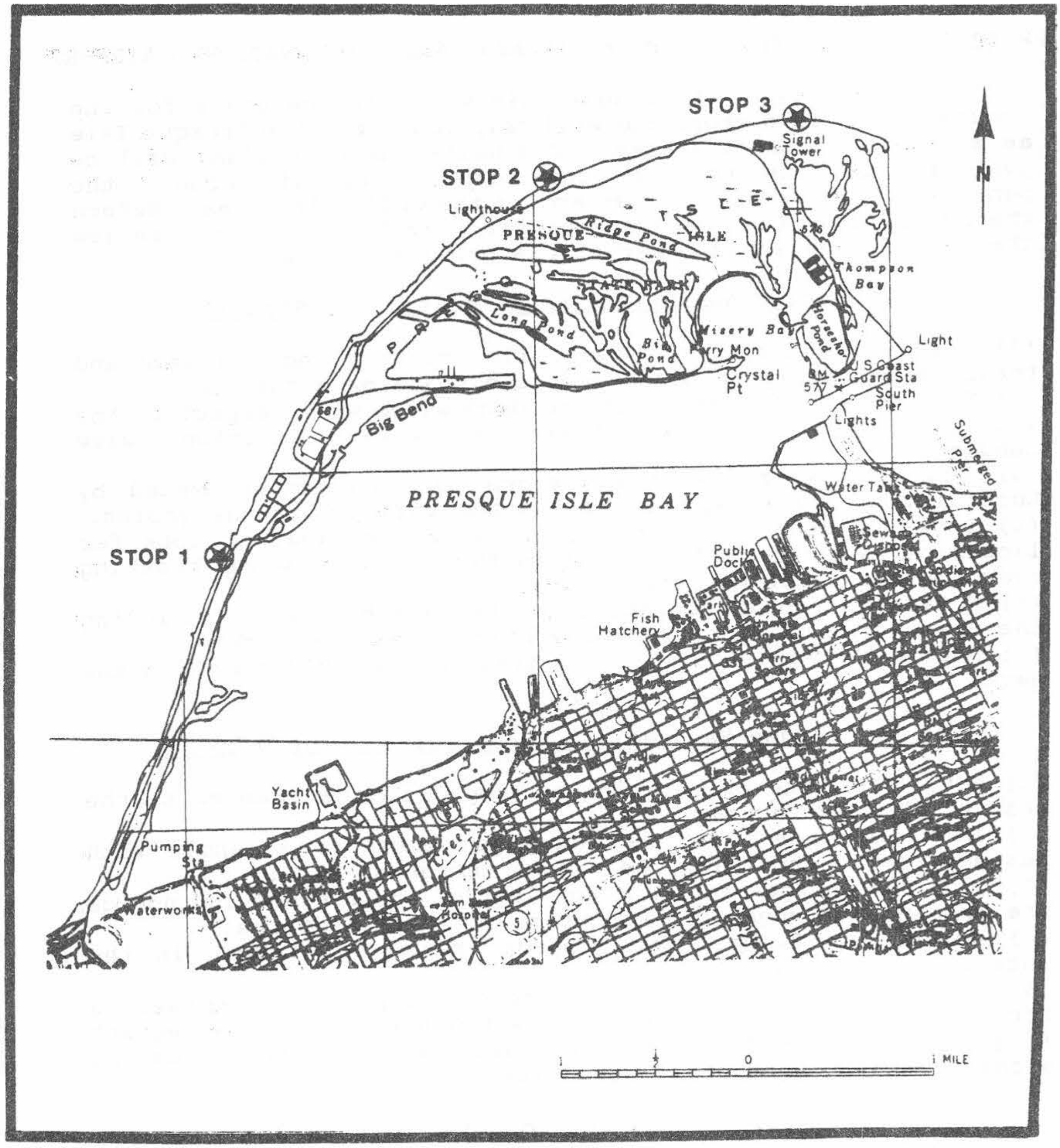


Figure 11. Stop locations for the field trip. Stop 1 is located just west of Beach 6, Stop 2 is located at Beach 9 and Stop 3 is located east of Beach 10 on Gull Point.

(6) Observe the orientation and spacing of the dune ridges and look for changes in their stage of development, particularly the succession of vegetation types associated with the ridges

STOP DESCRIPTIONS

STOP 1. BEACH 6: NECK SEGMENT

This stop is located to the west of Beach 6 between Groins 10 and 11 on the Neck segment of Presque Isle (Figure 11). Beach 6 is at the downdrift end of the groin field constructed to stabilize the Neck as part of the cooperative shore protection program beginning in 1954. This area is one of the erosional hot-spots along the peninsula's neck. Each year approximately 75,000 cubic yards of sand are dumped at this site as part of Presque Isle's annual beach nourishment program. This sand usually disappears by December after the intense storms in the fall.

Storm waves commonly arrive nearly perpendicular to this section of the Presque Isle shore. A large portion of the wave energy is transformed into storm water-level set-up against the beach. This results in a stratified current pattern with net lakeward underflow along the bottom. During severe storms with strong winds from the southwest, high waves superimposed on elevated lake levels commonly wash over the roadways and sometimes entirely across the spit into the bay.

Within the groin field the nearshore slope is steep. Bar forms include an outer bar and an inner bar separated by a bar trough. The inner bar is linear or crescentic in form. The outer bar is crescentic with lows or saddles separating the 400-1000 m segments. The beaches within the groin field are narrow and generally erosional. Downdrift offsets of the beaches adjacent to the groins demonstrates the net west to east transport of sediment called littoral drift. Other engineering structures such as rip-rap revetments significantly influence the character of the beaches.

Results of a comprehensive nearshore current study and native-sand tracer study completed prior to breakwater construction on the inner bar system (Sonnenfeld, 1981) and the outer bar-trough system (Taylor, 1981) indicate the following nearshore dynamics for storm events (wave heights greater than 60 cm):

(1) Sediment from the beach and the inner bar is transported lakeward towards the outer bar trough by rip currents located on the updrift sides of the groins, rip currents associated with cell circulation systems located between the groins and the net lakeward flow near the bed.

(2) Sediment from the lakeward flank of the outer bar and the bar crest is transported toward the trough inside the outer

bar (outer-bar trough) in response to currents generated by incident waves shoaling and breaking on the outer bar.

(3) Sediment within the outer bar trough is transported by an eastward directed shore parallel current within the trough producing an effective mechanism for sediment bypassing the beaches downdrift of the source areas.

(4) Scouring within the outer bar trough is common during these conditions and results in an increase in the relief of the outer bar.

For wave heights less than 60 cm, the net sediment transport is onshore and alongshore. Sediments move from the outer bar crest into the bar trough resulting in a decrease in bar relief. The inner bar moves onshore and may weld to the beach forming a cuspate berm.

At this time it is unclear what effect(s) the new breakwaters will have on the morphodynamics of the peninsulas neck. Some questions that need to be addressed include:

(1) What effect will the breakwaters have on sediment exchange between the beaches and the nearshore zone?

(2) What kinds of currents will be generated by the interaction of incident waves and the breakwaters?

(3) How much nourishment, if any, will be required to maintain the beaches?

(4) What will be the fate of the outer bar if offshore sediment transport is decreased?

THINGS TO LOOK FOR

* Notice wave height and approach direction with respect to the shoreline orientation.

* Notice longshore current direction and speed and look for evidence of rip currents (e.g. wave suppression in certain areas due to current outflow and suspended sediment plumes)

* Observe the texture of sediments including the beach replenishment materials.

* Observe the effects of engineering structures on beach morphology and wave reflection and refraction patterns.

* Notice the washovers in the backshore areas.

* Notice the newly constructed detached breakwaters. Can you see any difference in the beach behind the structures?

* Zebra mussels

STOP 2. BEACH 9: LIGHTHOUSE BEACHES

This stop is located to the east of the lighthouse at Beach 9. The dramatically different morphology and dynamics along this section of the peninsula are a consequence of the change in shoreline orientation with respect to the dominant southwest wave approach direction and the influx of sediment eroded from the peninsula's neck.

The Lighthouse Beaches (Figure 11) are much wider with established multiple vegetated dune ridges on the backshore. Relic coastal dune ridges, separated by swales and ponds dominate the climax forest behind the beaches. The most striking feature of the beaches is the system of large-scale shoreline megacusps spaced at 400-800 m. These rhythmically spaced features are substantial packets of sand which typically migrate eastward maintaining their identifiable form over periods ranging from months to years. Sediment which is transported along the beach itself occurs within these features. It may be that as little as 20 percent of sediment transported along the Lighthouse Beaches occurs as beach transport; the bulk of sediment transport takes place offshore. As megacusps move eastward so do the rip-channel embayments between them. Beach erosion is tied to these rip-channel embayments and consequently it is these pockets of erosion that have been targets of beach nourishment. It has been established that little <u>net</u> erosion or deposition of the lighthouse shoreline occurs through time (Nummedal, 1979).

The nearshore zone morphology of the Lighthouse Beaches is the most complex of the three segments. It includes up to four longshore bars parallel to the beach which are superimposed on transverse bar components and an arcuate bar system (Figure 7). The significantly gentler slope of this multibarred nearshore zone is a direct consequence of the high rate of sediment supply to this area. The large sediment load for the nearshore zone is largely derived from sources updrift (west of the lighthouse along the erosional neck) rather than the local beaches.

Sediment exchange between the beach and offshore bars is limited to the arcuate bar system and megacusps. Sediment moves eastward along the arcuate bar system with some sediment re-entering the beach where the bars attach to the horns of the megacusps. The rest of the sediment continues along the bar system. Sediment is transferred back to the bar system from the beach by rip currents flowing within bar-rip channels emerging from the embayments between megacusps. The entire megacusp system is strongly skewed to the east because of eastward directed incident waves.

Many questions about the spatial and temporal nature of the shore parallel and transverse bar components remain to be answered. For example, the rates, patterns and mechanisms for sediment dispersal in the deeper portion of the nearshore zone are not understood. It is difficult to assess how the detached breakwaters will effect the dynamics of this area.

THINGS TO LOOK FOR

* Notice the wave height, wave approach direction and shoreline orientation.

* Notice the longshore current direction and speed.

* Observe the texture of sediments at different locations on the beach and backshore.

* The characteristics and spacing of megacusps and inner bars that may be recognized by observing waves breaking over their crests.

* Notice the distribution of pockets of erosion and deposition along the beach and their relationship to beach morphology.

* Notice the vegetated dune ridges and incipient dunes formed by the interaction of vegetation and aeolian activity in the backshore.

* Notice beach ridges and climax forest across the road.

STOP 3. BEACH 10 AND EAST: GULL POINT

This stop on the Gull Point segment of Presque Isle (Figure 11) features the beaches east of the prototype breakwaters built at beach 10 in 1978. Depending on the time available we will walk eastward onto Gull Point to see the extensive sand plain and series of accretionary dune ridges.

By this stage in your observations, it must be clear to you that Presque Isle is an eastward migrating spit system that feeds upon itself as it migrates. Within this system, sediment is eroded from the neck, transported across the broad shallow multibarred nearshore zone of the Lighthouse Beaches and is finally deposited adding to Gull Point, the only net depositional feature along the peninsula. The rapid growth of Gull Point is evidenced by the formation of a series of beach ridges superimposed on recurved spits which wrap around the end of the spit enclosing ponds.

The rapid accretion of Gull Point began in the 1930's possibly in response to the sudden increase in available sand supply from the breached neck in 1917-23, and is continuing today in response to the annual nourishment program. According to the U.S. Army Corps of Engineers (1979) the present volumetric migration rate of 289,100 cubic yards per year reflects the replenishment input which has averaged 259,000 cubic yards per year since 1955. The peninsula's continued eastward migration operates at an annual sediment budget deficit. The present deficit results partly from the spit having migrated to the eastern edge of its underlying platform in response to a long-term rise in lake level (post glacial rise of 1 ft. per 300 yr.). Each future increment of eastern migration of Gull Point requires much larger volumes of sand because the platform itself must prograde into deep water before the spit can be superimposed. Also, the present sediment volume being supplied to the Neck from streams and bluff recession updrift has been estimated at only 30,000 cubic yards per year (Nummedal, 1984).

The Gull Point beaches receive the lowest annual wave power along the entire outer shore of the peninsula. The orientation of the shoreline is such that only storms with incident waves from the northeast can generate sufficient energy for offshore transport and shoreface broadening. Such storms are infrequent and moderate in strength. This wave climate is such that sediment is kept close to shore maintaining a steep prograding profile.

Because the nearshore slope is steep and the surf zone is narrow only one, or at most two, longshore bars are present close to shore. Because the strongly eastward directed incident waves, rip channels and transverse bar components are strongly skewed to become nearly shore-parallel. This offshore profile is maintained by the prevailing onshore transport of sediments.

The beaches are dominated by megacusps with a spacing of 700-800 m. These megacusps are closely associated with inner bars that commonly weld onto the beaches. The welding bars are much like ridge and runnel systems (swash bar complexes) of ocean beaches in form and behavior. Of course we have no tidal effects here so the dynamics are necessarily different. Migration of megacusps produce rapid changes in beach orientation and beach slope along the Gull Point beaches.

THINGS TO LOOK FOR

* Notice the wave height, wave approach direction, shoreline orientation and longshore current direction and speed.

* Observe the changes in sediment texture with respect to changes in beach morphology and distance from the shoreface. Notice also the heavy mineral concentrations.

* Observe the areas of erosion and deposition along the beach including any evidence of bars in the process of welding to the beach.

* Notice the washovers in areas where the primary dune ridge has been eroded away.

*** Notice the different stages of dune ridge formation including the succession of vegetation types that accompanies dune ridge growth**

REFERENCES CITED

Berg, D. W., 1965, Factors affecting beach nourishment requirements at Presque Isle Peninsula, Erie, Pennsylvania: Proceedings of the Eighth Conference on Great Lakes Research, Publication No. 13, Great Lakes Research Division, University of Michigan, p. 214-221.

Berg, D. W., and Duane, D. B., 1968, Effect of particle size and distribution on stability of artificially filled beach, Presque Isle Peninsula, Pennsylvania: Proceedings of the Eleventh Conference on Great Lakes Research, International Association for Great Lakes Research, p. 161-178.

Dreimanis, A., 1979, Late-Pleistocene Lakes in the Ontario and Erie Basins. 12th Conf. Great Lakes Res. Proc., p. 170-180.

Jennings, O. E., 1930, Perigrinating Presque Isle: Carnegie Magazine, v. 4, p. 171-175.

Lewis, C. F. M. et al., 1966, Geological and Palynological studies of Early Lake Erie deposits. 9th Conf. Great Lakes Res. Proc., p. 176-191.

Lewis, C. F. M., 1969, Late Quaternary events in Lake Huron and Lake Erie basins. 12th Conf. Great Lakes Res. Proc., p. 250-270.

Messinger, D. J., 1977, Form and change of a recurved sandspit, Presque Isle, PA: unpublished M. S. thesis, State University of New York College at Fredonia, 121 p.

Nummendal, D., Monitoring of shoreline changes, Presque Isle, PA, Annual Report for Contract no. DACW49-78-C-0020, U. S. Army Engineers District, Buffalo, 50 pp., 1978.

Nummedal, D., Monitoring of shoreline changes, Presque Isle, PA, Annual Report for Contract no. DACW49-79-C-0020, U. S. Army Engineers District, Buffalo, 52 pp., 1979.

Nummedal, D., Monitoring of shoreline changes, Presque Isle, PA, Annual Report for Contract no. DACW49-78-C-0055, U. S. Army Engineers District, Buffalo, 72 pp., 1980.

Nummedal, D., 1983, Sediment transport and morphodynamics of the Presque Isle shoreface: Louisiana State University, Department of Geology, Coastal Research Group, Technical Report 83-1, 33 p.

Nummendal, D. et al., 1974, Littoral processes and sedimentation in the Cattaraugus Embayment, NY: Final Report. Contract No. DACW49-74-C-0118, 246 p. (ref. Clemens, 1976).

Nummedal, D., and Sonnenfeld, D., 1983, A tracer study of sediment movement in the bar system at Presque Isle, Pennsylvania: Report to the U. S. Army Corps of Engineers, Buffalo, for contract DACW49-81-C-0019, 80 p.

Nummedal, D., Sonnenfeld, D. L., and Taylor, K., 1984, Sediment transport and morphology at the surf zone of Presque Isle, Lake Erie, Pennsylvania, in Greenwood, B., and Davis, R. A., Jr., eds., Hydrodynamics and sedimentation in wave-dominated coastal environments: Marine Geology, v. 60, p. 99-122.

Pope, J., and Gorecki, R. J., 1982, Geologic and engineering history of Presque Isle Peninsula, PA, in Buehler, E. J., and Calkin, P. E., eds., Geology of the Northern Appalachian Basin, Western New York: Guidebook, New York State Geological Association 54th Annual Meeting, Amherst, New York, p. 183-216.

Saville, T., 1953, Wave and lake level statistics for Lake Erie: Beach Erosion Board, Technical Memo No. 37, 24 p.

Sonnenfeld, D. L., 1983. Inner bar sediment dynamics, Presque Isle, PA. Unpublished M. Sc. thesis, Louisiana State University, Baton Rouge, LA.

Thomas, D. J., et al., 1987, Pleistocene and holocene geology on a dynamic coast: 52nd Annual Field Conference of Pennsylvania Geologists. Department of Environmental Resources, Bureau of Topographic and Geologic Survey, Harrisburg, PA, 88 p.

U. S. Army Corps of Engineers, 1953, Presque Isle Peninsula, Erie, Pennsylvania: Beach Erosion Control Study. House Document No. 231, 83rd Congress.

U. S. Army Corps of Engineers, Buffalo District, 1980. Presque Isle Peninsula, Erie, Pennsylvania: Final Phase 1 General Design Memorandum, Buffalo, NY.

Walton, J. C., 1978, Sediment characteristics and processes of beaches and bars, Presque Isle, Erie, Pennsylvania (6/22/74-8/10/74): unpublished M. S. thesis, State University of New York College at Fredonia, 157 p.

Williams, S. J., and Meisburger, E. P., 1982, Geological character and mineral resources of south central Lake Erie: U. S. Army Corps of Engineers, CERC Miscellaneous Report 82-9, 62 p.

A FEW OF OUR FAVORITE PLACES; AN ENVIRONMENTAL AND GEOLOGICAL EXCURSION IN CHAUTAUQUA COUNTY

Richard A. Gilman and Jack Berkley, State University of New York, College at Fredonia, N.Y. 14063.

INTRODUCTION

This trip will focus on a variety of geological features and processes of particular interest to students, rather than on results of current research on a single topic. General topics include environmentally important sites, geomorphology and glacial geology, and bedrock geology including unusual structural features. The setting for this trip is the a portion of the eastern shore of the Lake Erie Basin in southwestern Chautauqua County. Upper Devonian bedrock units (mostly shale and siltstone) are overlain by Wisconsin-age glacial drift, mostly morainal material on the "escarpment" (edge of the Allegheny Plateau), with post-glacial, pre-Lake Erie lake clay dominating in the glacially-scoured lake plain below. Post-glacial erosion has produced deeply incised fluvial valleys, some with rather spectacular gorges. Human economic activities center on agrarian enterprises, with the famous Lake Erie Grape Belt dominating production in the lake valley.

This field trip will take us to (in order of stops) [1] a "pop-up" structure in Canadaway Creek near the SUNY, Fredonia campus, [2] an active and [3] inactive fly ash waste disposal site, [4] Lake Erie State Park (lake erosion, bedrock features, glacial till), [5] a glacial lake Whittlesey beach deposit quarry, [6] a scenic overview from the "escarpment" of the Lake Erie basin (lunch), and [7] a scenic-educational hike into the Chautauqua Gorge to see bedrock and glacial features, including a pre-glacial buried valley. We strongly advise that all participants bring waterproof boots, or for the more adventurous, sneakers. This apparel will be essential particularly on STOP [7] which requires considerable stream wading. Fig. 1 is a map showing locations of all stops.

DESCRIPTION AND COMMENTARY ON INDIVIDUAL STOPS

STOP 1. POP-UP STRUCTURES

Pop-up anticlines were recorded as early as 1886 by G.K. Gilbert in otherwise flat lying Devonian strata of New York and Pennsylvania (Gilbert, 1886, 1888). He attributed these structures to "thermal expansion of surficial rocks following glacial unloading" (Sbar and Sykes, 1973). Since that time others have reported similar structures (Cushing et.al., 1910, Williams, H. R. et al.,1985, and Adams, J., 1982) in which the axis of the pop-up generally strike N-S or NW-SE. Sbar and Sykes (1973) reviewed the evidence favoring a contemporary ENE to
E-W maximum compressive stress for much of the Great Lakes region of North America (Fig. 2), and suggested that "the release of a vertical load in the presence of a large horizontal compressive stress of non-glacial origin, however, may be responsible for the formation of the pop-up...".

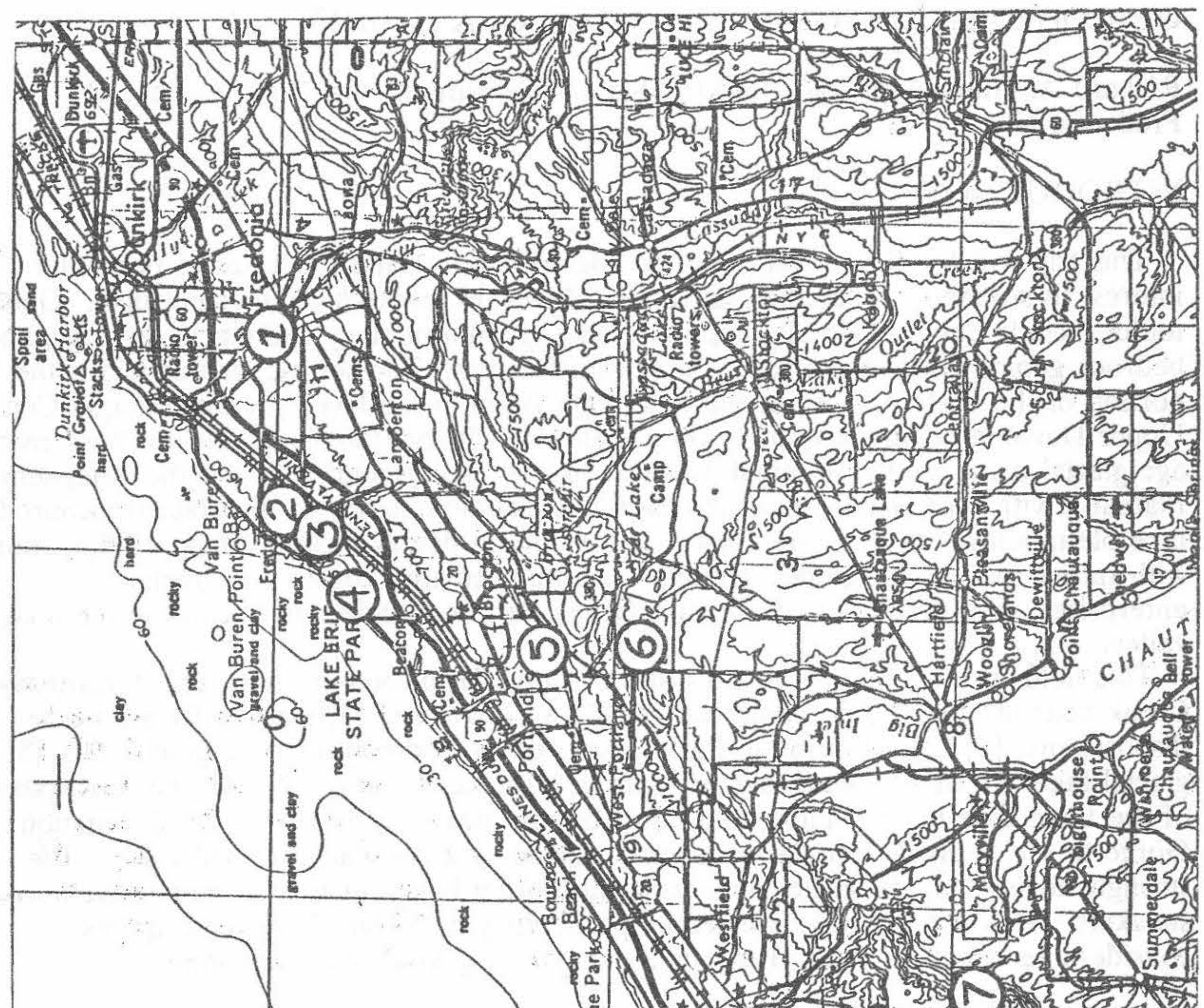


Figure 1. Map of a portion of southwestern New York State showing locations of stops on this field trip.

Pop-up anticlines in northern Chautauqua County generally have a NW-SE to N-S strike (STOP 7) but some have been found that strike E-W (STOP 1).

At this exposure notice that the shaley beds and even some siltstone beds are bent at the fold. Does this imply a certain degree of ductile deformation, and if so what was the mechanism of deformation and when did it occur? If the northwest trending structures are the result of the ENE regional compressive stress, how are the E-W trending pop-ups explained? Coates (1964) describes a 7 foot high, NW striking pop-up that developed overnight in a quarry floor in Ontario. Note that the STOP 1 fold is directly overlain by unconsolidated floodplain deposits and the constituent shale bedrock was, no doubt, subjected to fluvial erosion until "recently" (?). Because this fold represents a positive topographic feature it should be highly susceptible to erosion; note that where the fold crosses the present stream bed it has been planed off completely. Thus, if the Canadaway Creek valley is a post-glacial feature, the STOP 1 fold is probably a fairly recent feature (post-glacial), as an older fold (say, Paleozoic age) should have been eroded down to the current erosion level of the bedrock. We will compare the structural properties of this fold to folds of probable glacial origin at STOP 4 (Lake Erie State Park).

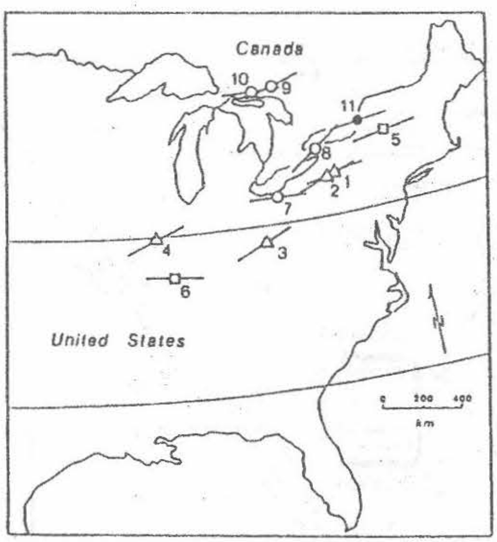


Figure 2. The status of information on intraplate stress in the eastern United States as of April 1972 based on Sbar and Sykes (1973) compilation. The strike of the horizontal component of the maximum compressive stress (S_H) is shown for hydraulic fracture tests (open triangles), fault-plane solutions (open squares), overcoring stress measurements (open circles), and post-glacial pop-ups (solid circle). From Gross, 1989.

STOP 2. NIAGARA MOHAWK FLY ASH & BOTTOM ASH DISPOSAL SITE
STOP 3. FRAME TRUCKING CO. RECLAIMED DISPOSAL SITE

Environmentally compatible disposal of toxic waste products is a major problem facing federal, state, and local governments. Western New York has gained national attention as a major dumpsite of hazardous materials, particularly in the Buffalo and Niagara Falls area (witness "Love Canal"). Chautauqua County is not immune to these problems. The county government is currently at odds with the state DEC over expansion of capacity at the Town of Ellery Landfill operation, a site in which the active dumpsite is near capacity levels. Closer to home, the Dunkirk-Fredonia area hosts a coal ash disposal site run by the Niagara Mohawk Power Corporation. We will visit the currently operational site and a nearby "reclaimed" site owned and formerly operated by the Don Frame Trucking Company (Fig. 3). This disposal cell was active from 1971 to 1985.

The Niagara Mohawk site occupies land that formerly included the Fredonia Airport on Van Buren Road (Fig 3). It primarily receives ash waste from the Dunkirk Steam Generation Plant located near Point Gratiot, Lake Erie, within the city limits of Dunkirk. This plant began electrical generation in 1949 using Pennsylvania bituminous coal as fuel. It generates 600 megawatts of power fired by about 5,000 tons of coal per

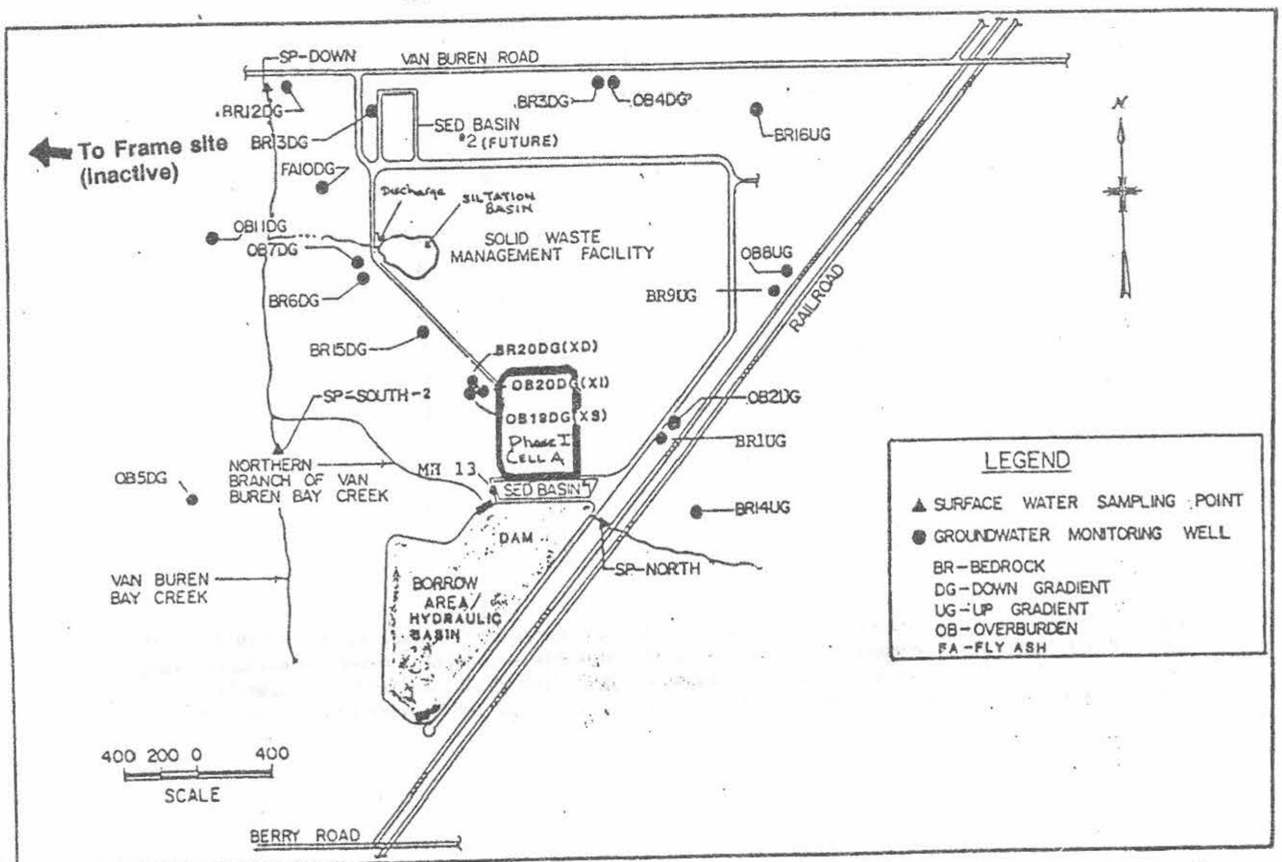


Figure 3. Map of Niagara Mohawk fly ash disposal area, Dunkirk. Cell A (center) is currently active dump site. The reclaimed "Frame site" (STOP 3) is located approximately in upper left of the map.

day. Originally the ash produced at the plant was mostly "bottom ash", the same "clinkers" or coal cinders produced by home, coal-fired furnaces. However, in 1973 new and more efficient electrostatic precipitators ("scrubbers") were constructed in the exhaust chimneys resulting in cleaner air, but creating addition solid wastes in the form of "fly ash". It is this fine grained fly ash that accounts for the majority of wastes currently dumped at the Dunkirk disposal site. An average of about 450 tons of fly ash deposited at the site every working day (Sunday excluded). Some bottom ash is also dumped at the site (avg. 1/90-2/90 was 130 tons), although much of this is bought by the Town of Pomfret for use as road aggregate. The disposal site also receives some wastes from an oil-fired plant in Oswego.

Of obvious environmental concern is the efficiency of safeguards designed to prevent aqueous leachates at the disposal site to contaminate local surface and groundwater supplies. This is controlled for the most part by clay and plastic bottom liners. Leachates are directed toward an artificial pond ("Sed Basin" in Fig. 3) that is frequently monitored for contaminants (see Table 1). Water from this pond is allowed to flow into a nearby creek after testing shows it to be free of significant contamination. Groundwater is also monitored by a series of test wells (Fig. 3) located around the perimeter of the site. Dust abatement is performed by watering trucks that obtain water from a larger artificial pond adjacent to the settling basin (Fig. 3).

TABLE 1. Example of groundwater analysis, Dunkirk Solid Waste Management Facility. Environmental Testing Facilities report 2/15/1990 for test well OB2UG.

METALS mg/L		ORGANICS ppb	
Al	2.58	Bromodichloromethane	<1.00
Sb	<1.00	Bromomethane	<1.00
As	<0.005	Vinyl Chloride	<1.00
Ba	<0.2	2-Chloroethylvinyl Ether	<1.00
Be	<0.005	Chloromethane	<1.00
B	0.64	Trichloroflouromethane	<1.00
Cd	<0.005	1,1,2-Trichloroethane	<1.00
Ca	61.0	1,1-Dichloroethane	<1.00
Cr	0.003	1,1-Dichloropropane	<1.00
Cu	<0.02	trans-1,3-Dichloropropane	<1.00
Fe	3.86	1,2,2,2-Tetrachloroethane	<1.00
Pb	<0.005	1,1,1-Trichloroethane	<1.00
Mn	0.037	Benzene	<1.00
Hg	<0.0002	Ethylbenzene	<1.00
Mo	<0.5	1,4-Dichlorobenzene	<1.00
Ni	<0.04	1,2-Dichlorobenzene	<1.00
Se	<0.002	o-Xylene	<1.00
Ag	<0.01	Bromoform	<1.00
Na	15.2	Carbon Tetrachloride	<1.00
Tl	<1.00	Chloroethane	<1.00
Zn	0.02	Chloroform	<1.00
Cr+6	<0.01	Trichloroethane	<1.00
		(...etc. All other organics = <1.00 ppb)	

NOTE: "<" *denotes less than minimum detection limits.*
Analyses by Environmental Testing Facilities, Inc. Dunkirk, NY.

Metzger (1974) noted that groundwater samples taken near the active fly ash disposal cell at that time (Frame site; STOP 3) showed that efforts to contain groundwater contamination were successful. However, he also showed that water samples collected near an older dump site (near the entrance on Van Buren Road) were contaminated with high concentrations of iron and manganese. It is safe to conclude that these metals were derived by leaching of the old fly ash materials by groundwater, and that other heavy metals (not analyzed) may also have shown higher-than-background abundances. Table 1 shows that containment techniques at the current cell are successful in keeping contamination restricted to the disposal site. All metal values are quite low and abundances for organics are all below detection limits. Hopefully, this situation will be maintained in subsequent years, but this remains to be seen.

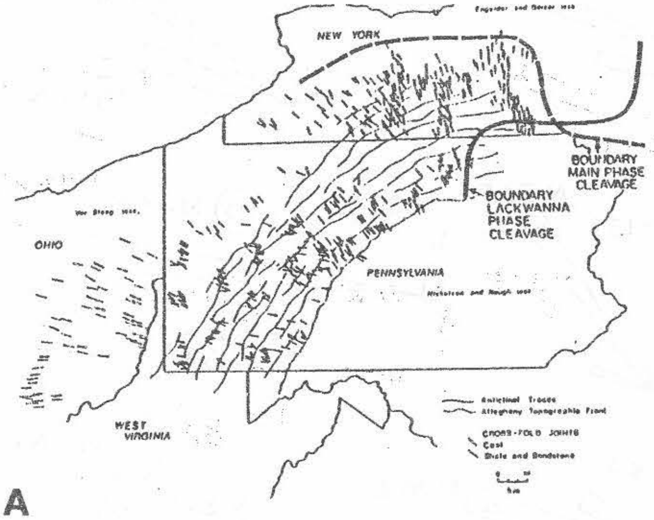
STOP 4. LAKE ERIE STATE PARK

The lake shore here offers us the opportunity to study several features of interest: (1) jointing developed in the South Wales member of the Canadaway Formation....we will also discuss additional studies on jointing in this area and across the Allegheny Plateau, (2) shoreline erosion...recent problems of severe erosion rates associated with high lake levels and how the U.S. Corps of Engineers resolved the problem at the bath-house, (3) a beautiful exposure of glacial till overlying shale bedrock along with lacustrine deposits overlying the till, and (4) septarian concretions.

JOINTS: In recent years joint patterns across the Allegheny Plateau have received considerable attention (Engelder, 1982, 1985; Engelder and Geiser, 1980; Gross, 1989). Fig. 4 shows results of these studies. Of particular interest is the noticeable change in dominant joint orientation from northwest to northeast that takes place in southern New York and northern Pennsylvania (Fig. 4C. There is also a clear need for additional data from western New York. This will in part be addressed by our discussions here at STOP 4. The strong ENE set of extensional joints (Fig. 4B) is interpreted as having formed in a neotectonic stress field with an ENE direction of maximum compressive stress (Engelder, 1982). As summarized by Gross (1989), "These features are thought to form near the earth's surface as the result of uplift and erosion and are controlled by the orientation of the contemporary tectonic stress field."

At Lake Erie State Park the South Wales member of the Canadaway Formation is well exposed at water level. It displays an abundance of joints oriented approximately N 50 W and N 30 E. Fig. 5C is a rose diagram of 134 joints measured at this location.

In preparation for this field trip RAG undertook a reconnaissance study of joint patterns in each member of the Canadaway Formation in northern Chautauqua County. Results are shown rose diagrams in Fig. 5. These diagrams show that in all cases there cases there is at least 1 strong NE joint set (N 25-35 E and/or N 65-70 E). In addition a N 30-40 W set and/or a N 50-60 W set is present in some units. However they are not consistently present in all exposures of these members. For example, the Northeast shale in Canadaway Creek has abundant NW joints, but very few are seen in Chautauqua Gorge (Fig. 5I & 5J).

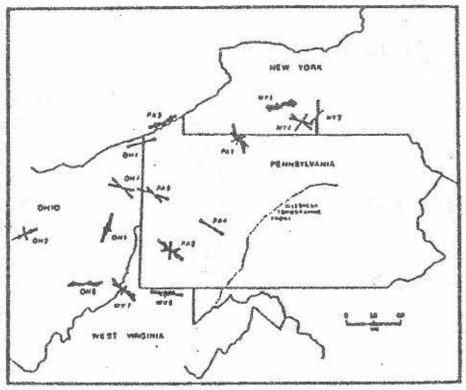


A

The distribution of cross-fold joints and cleavage within the Central Appalachians (after Engelder, 1985).

B

Rose diagrams of orientations of joints in drill cores from wells in Ohio, West Virginia, Pennsylvania, and New York. (From Terry Engelder, 1987, in *Fracture Mechanics of Rock*, B. Atkinson, ed., Academic Press.)



C The regional ENE joint set in the northeastern United States classified as neotectonic, after Engelder (1982). The orientation of the joints are shown as solid black lines.

Figure 4. Maps of joint orientations in northeastern U.S.

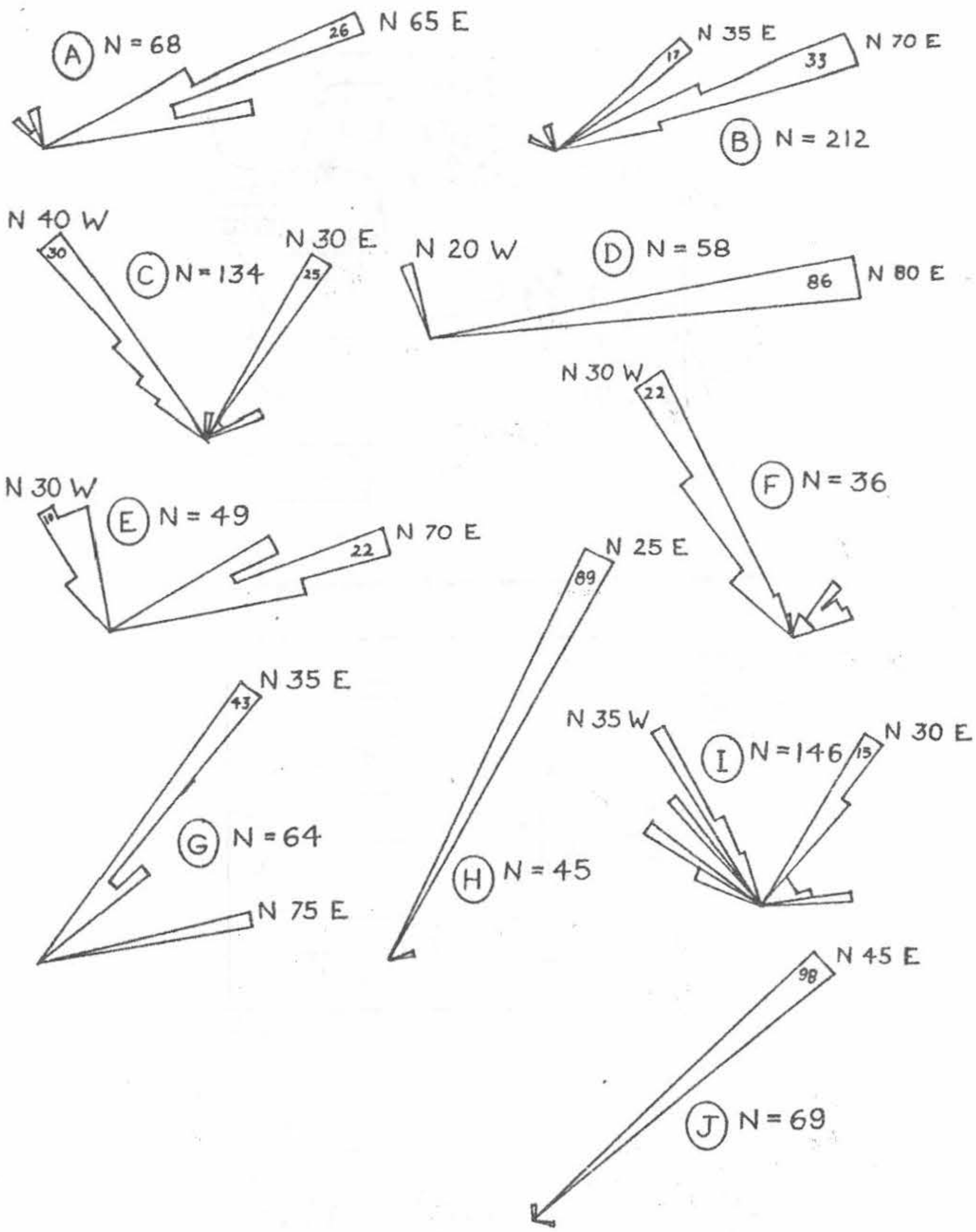


Figure 5. Rose diagrams of joints in members of the Canadaway Formation showing strong NE, and to a lesser extent, NW trends: [A] Dunkirk Shale at Point Gratiot, Dunkirk; [B] South Wales shale from Little Canadaway Creek, Portland; [C] South Wales shale from Lake erie State Park; [D] Gowanda shale from Canadaway Creek, Fredonia; [E] Gowanda shale Chautauqua Creek, Barcelona; [F] Laona siltstone from Chautauqua Creek, Barcelona; [G] Westfield shale from Walnut Creek, Forestville; [H] Shumla siltstone from Canadaway Creek, Shumla; [I] Northeast shale from Canadaway Creek, Arkwright; [J] Northeast shale from Chautauqua Gorge. N is the number of field measurements. Not all diagrams are to the same scale. Number inside the "rose petal" is the percentage of N.

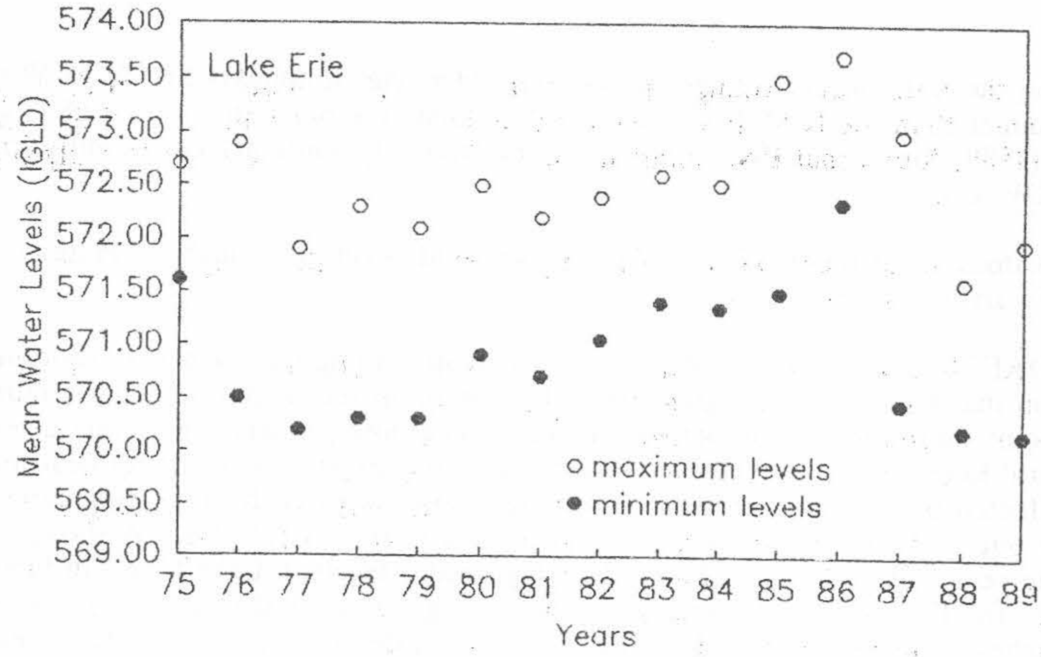
Using the criteria that younger joints abut older ones it appears that N 35 W joints are younger than the N 65 E set, and N 70 W joints are older than the N 65 E set. Gross (1989) found that ENE joints in the Lockport dolomite are consistently younger than NW sets.

How does all of these relate to earlier joint studies on the Allegheny Plateau?... that will be part of our discussion.

SHORELINE EROSION: Wave-induced erosion along the Lake Erie shoreline is a problem that has attracted a great deal of attention in recent years. From a human standpoint shoreline erosion destroys valuable lakeshore property and jeopardizes the structural integrity of buildings and other near-shore constructions. Sand beaches are also affected by increased erosion during high-water times, with sand being carried farther out in the basin during times of increased wave activity. Fig. 6A shows that the year 1986 experienced especially high water levels in Lake Erie. In fact that year set an all time record, exceeding previous highs measured in 1973 (573.51 feet). The highest level recorded in 1986 was 573.70 recorded during the summer. Levels have generally decreased since that time and were especially low during the drought of 1988 (Fig. 6A). In fact, the most recent Army Corp of Engineers lake level report suggests that lake levels will remain well below average during the summer of 1990. During times of high lake levels erosion rates increase because waves are more likely to impinge upon easily erodible glacial sediments that overlie bedrock. During low water periods waves contact bedrock which erodes less efficiently than glacial till. In addition, during low water periods wave energy is likely to be dissipated over a broad area of lake bottom, thus lessening the energy waves can expend on erosion (Metzger, 1974).

Fig. 6B shows the typical lake level pattern over any given year which is to rise during the spring and summer months and decrease during winter and fall. This is, perhaps, fortunate because winter storms can be particularly energetic. Storms at any time of the year cause strong prevailing winds to develop on the lake basin. This results in "setups", a process where water literally piles up on one side of the lake, usually the eastern or southern side. Storm setups combined with unusually high water levels can cause considerable erosion along the lake shore, and may destroy beaches. Sand carried out to deeper levels usually returns, however, as lake levels return to more normal values (Metzger, 1974). Fig. 7 shows that the Lake Erie shoreline in western New York is generally subject to light to moderate erosion.

A dramatic example of severe shoreline erosion directly affecting human activities can be observed at Lake Erie State Park. Erosion of virtually unconsolidated glacial till under lying the bathhouse was threatening to undercut this building's foundation. In 1986 the U.S. Army Corp of Engineers proposed an erosion abatement plan to save the building from probable structural damage resulting from this erosional undercutting. The bathhouse, situated on a 25 foot high bluff, was valued at about $300,000 by the state in 1986, so cost considerations dictated that constructing a new facility would be more expensive than saving the old building. The Corp of Engineers, therefore, devised a plan to construct a "rubblemound" revetment that would extend 250 feet along the toe of the bluff, and extend up the bluff to the bathhouse. The revetment project was subsequently approved for construction. It consists of a gravel fill layer overlain by large limestone boulders. A plastic filter fabric was laid down under this material to keep



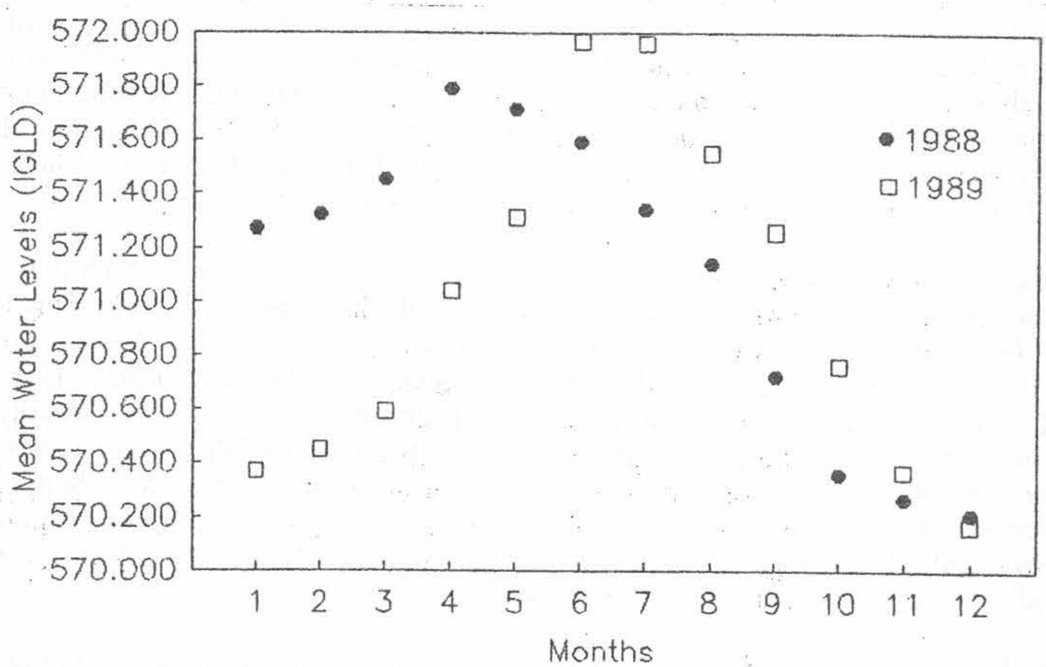


Figure 6. [A] Mean water levels in Lake Erie for the years 1975 to 1989. IGLD refers to "International Great Lakes Datum", water level at Father Point, Quebec measured in 1955. Note the record high levels in 1986; and present day low levels. [B] Monthly mean lake levels for Lake Erie in the years 1988 and 1989. Note the general tendency of lake levels to rise in spring and summer.

bluff materials from infiltrating the revetment, and a series of buried plastic pipes aids drainage. The estimated cost of the project prior to construction was $200,000. Fig. 8 shows a map view of the construction plan that was eventually completed in 1988.

GLACIAL DEPOSITS: Bedrock along the beach area is overlain by unconsolidated surficial deposits of glacial origin, deposited during the Wisconsin glacial advance. This material is well-exposed on the cliff face along the beach and consists of buff-colored, unstratified, badly sorted till, overlain in some areas by finely laminated, gray lake clay. The till consists of some recognizable local bedrock material especially in the lower horizons, but also includes pebbles through boulder-size rocks mostly of metamorphic or igneous origin. Some large blocks of relatively coherent bedrock were incorporated into the lower till layers in some areas. These blocks display local folding that may have been induced by glacial movement. We will visit one particularly well-exposed site and compare this folding to that observed in the pop-up structures observed at STOP 1.

Lake clay is exposed in some areas above the till, but has been removed by erosion elsewhere. It is thinly laminated, nearly pure gray clay in many areas but locally is mixed with sand, gravel or pebbles. The layering may represent a response to seasonal depositional conditions (varves), although this is subject to debate. Most varved clays show distinct color differences, apparently absent here. This clay was deposited by one of the precursor lakes to the present Lake Erie, Whittlesey or Warren. Judging from the low stratigraphic position of this clay relative to the till layer, the lake clay was probably deposited be the earlier lake, i.e., Whittlesey.

SEPTARIAN CONCRETIONS: A few hundred feet west of the till exposure, numerous 3 to 5 feet diameter septarian concretions are exposed at water level. According to Pettijohn (1975) the formation of septarian nodules appears to involve development of a nodular body, case hardening of the exterior with dehydration of the interior and generation of the shrinkage-crack pattern, and partial or complete filling of the cracks with precipitated mineral material, thereby producing the vein network of the nodule. Shrinkage and production of cracks imply a gel-like character of the original body.

Ultimately, development of concretions may be related to bacterial fermentation or organic matter during shallow burial. Such reactions apparently generate water, carbon dioxide and biogenic methane. This leads to an increase in the pH of the pore fluids allowing for carbonate precipitation. The carbonate cements which may include calcite and siderite, commonly develope as concretions that may be found intermittently along beds. Some of the nodules, the septarian concretions, have undergone dehydration with accompanying development of shrinkage cracks (Selley, 1988). These cracks are later filled in by precipitation of carbonate minerals (calcite and siderite) from infiltrating hydrous solutions. In addition, the relatively uncommon mineral, barite ($BaSo_4$), is a common accessory mineral in many septarian concretions.

STOP 5. GRAVEL PIT IN LAKE WHITTLESEY BEACH DEPOSITS

As the Wisconsin ice front retreated north of the high ground of the Portage Escarpment, meltwater that had previously drained southward was now trapped between

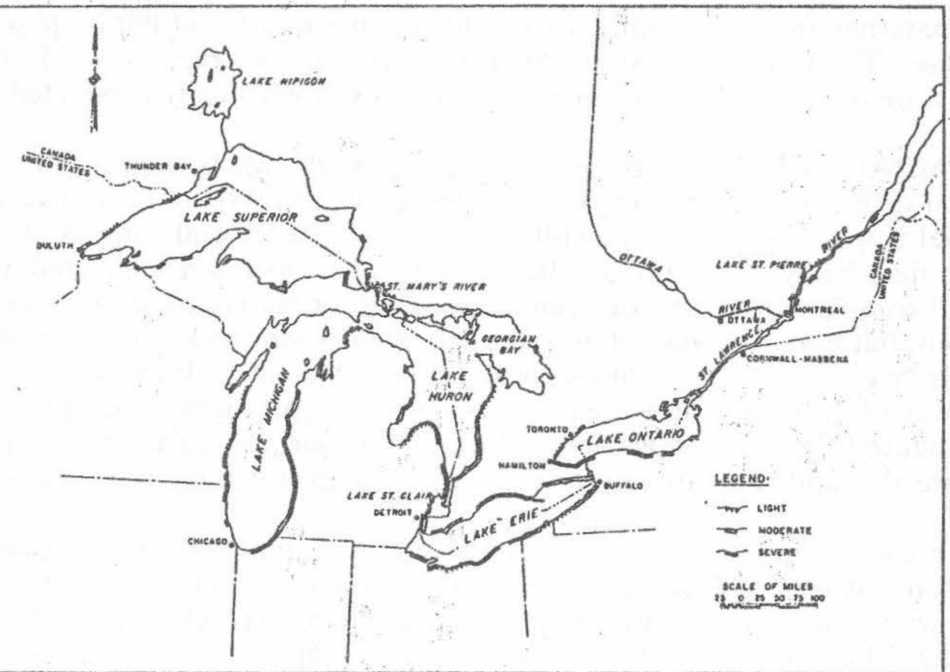


Figure 7. Map showing Great lakes and St. Lawrence River areas subjected to shoreline erosion.

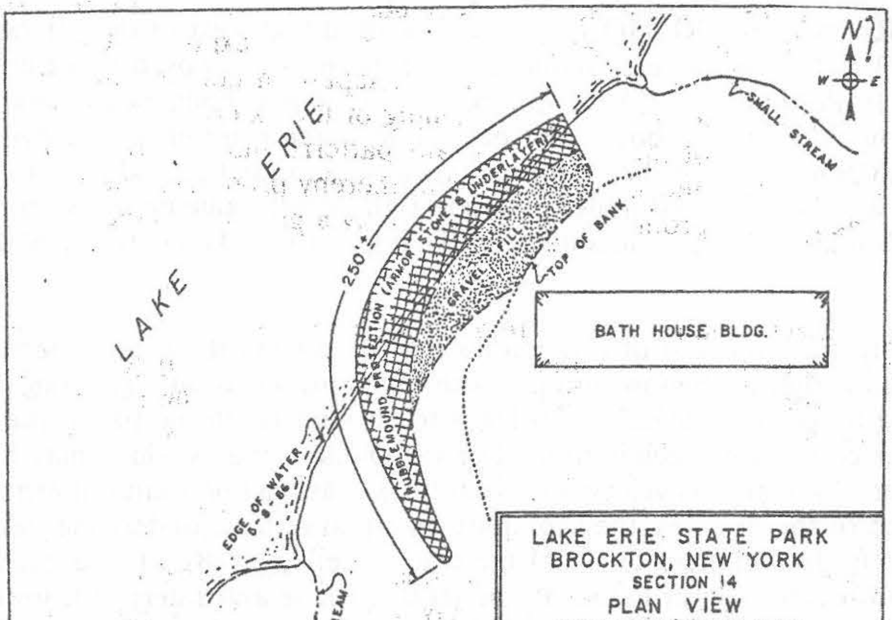


Figure 8. Map of the Lake Erie State Park revetment system.

the ice margin and the escarpment. This led to numerous small lakes and eventually to the development of three large lakes (the ancestral Great Lakes), Lakes Maumee, Whittlesey, and Warren. Two of these, Whittlesey and Warren, Formed beach terraces that are well preserved in Chautauqua County (Fig. 9). Prior to the deglaciation of the Mohawk and St. Lawrence river systems the drainage of both lakes Whittlesey and Warren was westerly. As the northeast outlet channels became ice free Lake Erie was established at it's present level. The sequence of deglaciation events for Chautauqua County is shown in Fig. 10.

In the Westfield-Fredonia area, Route 20 lies along the Lake Warren beach and Webster Road along the Lake Whittlesey beach.

STOP 5 is in a Lake Whittlesey beach deposit where the following features should be noted; the style and dip of bedding, composition of the pebbles in the gravels, and the structures and textures within the beds.

Due to post glacial isostatic rebound the elevations of these beach terraces rise fifty feet toward the east between the villages of Ripley and Silver Creek, a distance of about 40 miles.

STOP 6. THAYER ROAD OVERLOOK

The Lake Erie Basin was likely a low area rimmed on the south by a cuestaform ridge in pre-glacial times (Dreimanis, 1969;Gravenor, 1975). The sharp rise in topography in front of this overlook is called the Portage Escarpment that although modified by glacial activity is believed to have formed by subaerial erosion prior to Pleistocene glaciation. Bedrock of the escarpment consists of the Canadaway Formation (Fig. 11). Unlike the Niagara Escarpment to the east, the Portage Escarpment is not capped by the resistant strata but is dominated by shales with interbedded thin siltstones. Muller and Fahnestock (1974) asked, "If the topography be truly cuestaform, where are the resistant capping strata responsible for the scarp which is 700 to 1100 feet high in eastern Chautauqua County?"

The overlook is situated on top of the Lake Escarpment Moraine, a moraine complex that Muller (1963) correlates with the Valley Heads Moraine in the Finger Lakes region. This is the result of a stationary ice front at about 14,000 B.P. The moraine is a prominent feature from Northeast, PA to east of Gowanda, NY.

STOP 7. CHAUTAUQUA GORGE HIKE

Note: The hike will take about 3 hours and much of the time we will be walking in the stream. Consequently you should have footwear that you don't mind getting wet and that has good traction. These rocks are extremely slippery when wet...USE CAUTION.

As the ice margin retreated northward from the Portage Escarpment, north draining streams such as Chautauqua Creek were established. Some of these streams carved deep gorges across the escarpment, in most cases carving their channels into the

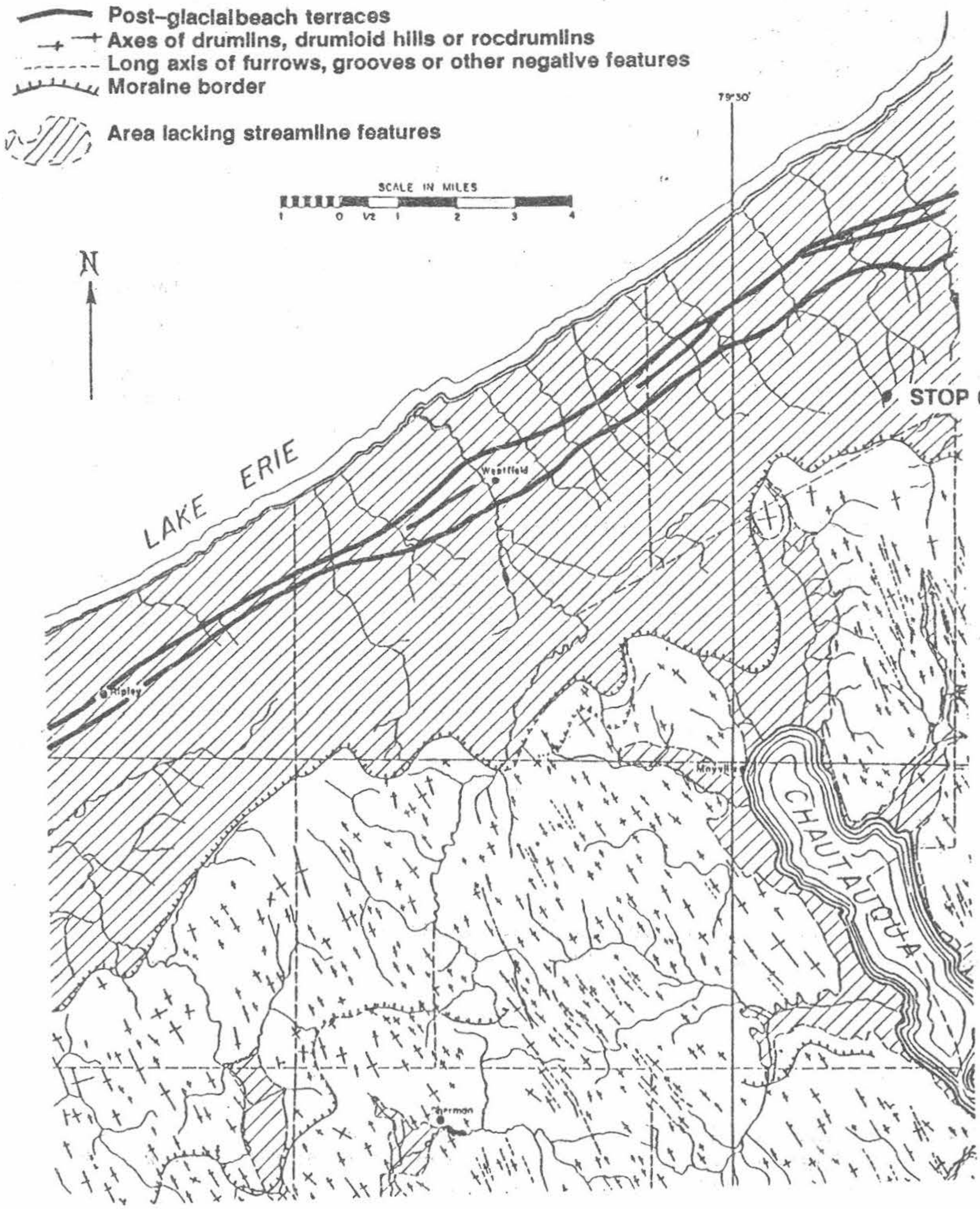


Figure 9. Map of beach terraces and other glacial features.

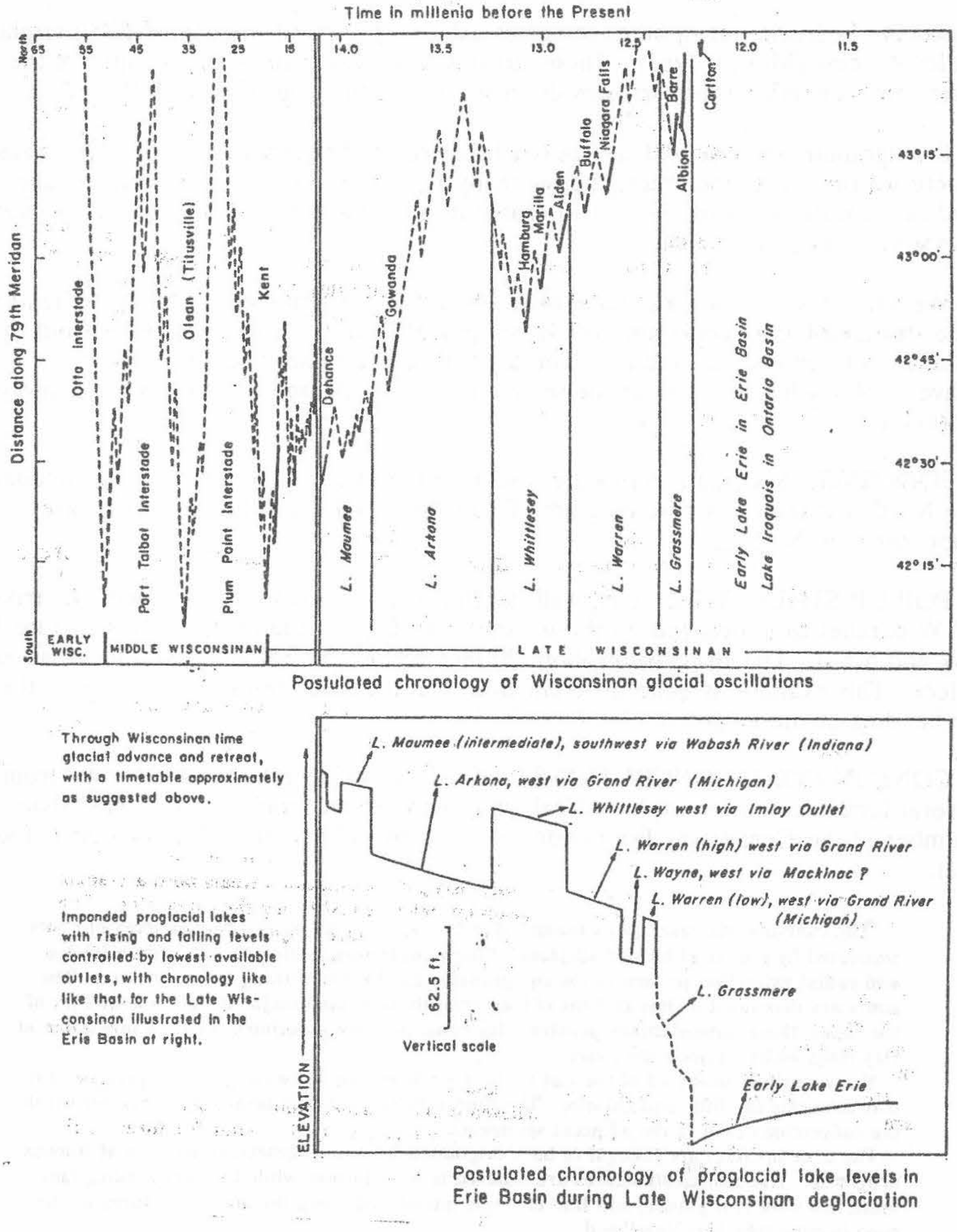


Figure 10. Postulated chronology of late Wisconsin deglaciation. From Muller, 1975.

Devonian bedrock, but in other instances exhuming parts of pre-glacial (or interglacial) valley systems (Muller, 1963). These streams head less than 10 miles south of the Lake Erie shore. Farther south streams drain southward into the Allegheny River.

Stratigraphic units we will see on our hike are the Dexterville (siltstone), exposed where we first enter the creek; the overlying Ellicott, seen high on the steep valley walls; and as we walk downstream we will cross into the Northeast (shale)that lies beneath the Dexterville (Figs. 11 & 12).

We will walk approximately one mile downstream examining the following features: The stratigraphic influence on joint development, pop-up structures, fossils from the Ellicott, cone-in-cone concretions, and a pre-glacial channel filled with kame gravels. We will also consider the sedimentology of this interval of the Canadaway Formation.

JOINTING: Systematic joints are absent in the Dexterville member but abundant in the Northeast shale. A rose diagram of 70 joints is shown in Fig. 5J. Most are approximately N 45 E.

POP-UP STRUCTURE: One well developed pop-up will be examined. It strikes N 27 W parallel to a localized northwest joint set. Dips at the crest of the structure are 15 SW and 6 NE. The structure is about 20 feet across and has an amplitude of about 2 feet. This example is quite different from the anticline seen at stop 1 in that there is no bending at the hinge.

CONE-IN CONE CONCRETIONS: Cone-in Cone concretions are known from several locations in western New York and Pennsylvania from the Northeast shale member of the Canadaway Formation. A study by Gilman and Metzger (1967) found that:

> **"The cone-in-cone concretions examined in this study occur as two hemispheres of cones separated by a layer of bedded siltstone. Microscopic examination reveals both a bladed and radial extinction pattern in the equigranular calcite constituting the concretion. The cones are disrupted by two systems of fractures: those termed major form the boundary of the cones, those termed minor penetrate the cones and are associated with the formation of clay rings which encircle the cones.**
>
> **X-ray analysis shows all of the $CaCO_3$ in the concretions to be calcite. The predominate clay minerals are illite and chlorite. The concentration of mixed layer clay is greater within the concretions than in the adjacent sediments.**
>
> **The cone-in-cones are thought to have originated from a syngenetic concretion of fibrous aragonite. The conical structures are believed to have formed while the surrounding clay materials were still plastic, and that clay was introduced along the major fractures as the cone-in-cone structure developed."**

A cross section of a typical concretion is shown in Fig. 13.

PALEONTOLOGY AND SEDIMENTOLOGY: Faunal communities of the Dexterville and Ellicott members were studied by Burrier (1977). His table of fauna from various lithologies within the Chadakoin Formation is shown in Table 2. He concluded that these strata represent regressive and transgressive facies of delta front and prodelta environments. Miller and Lash (1984) consider the Northeast shale to be

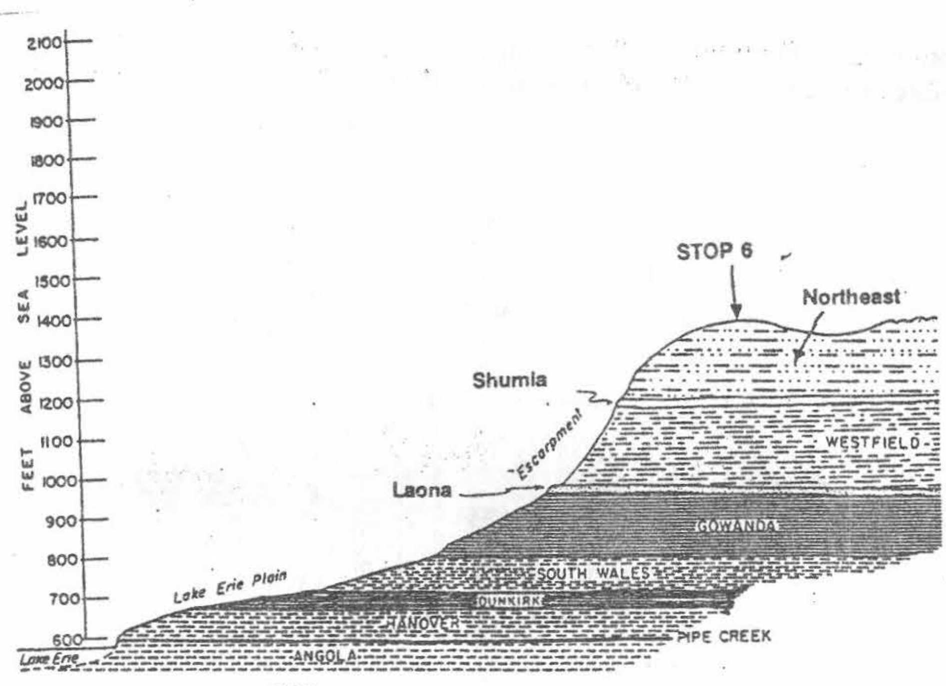


Figure 11. Cross section of the Portage Escarpment, southwestern New York State. From Tesmer, 1963.

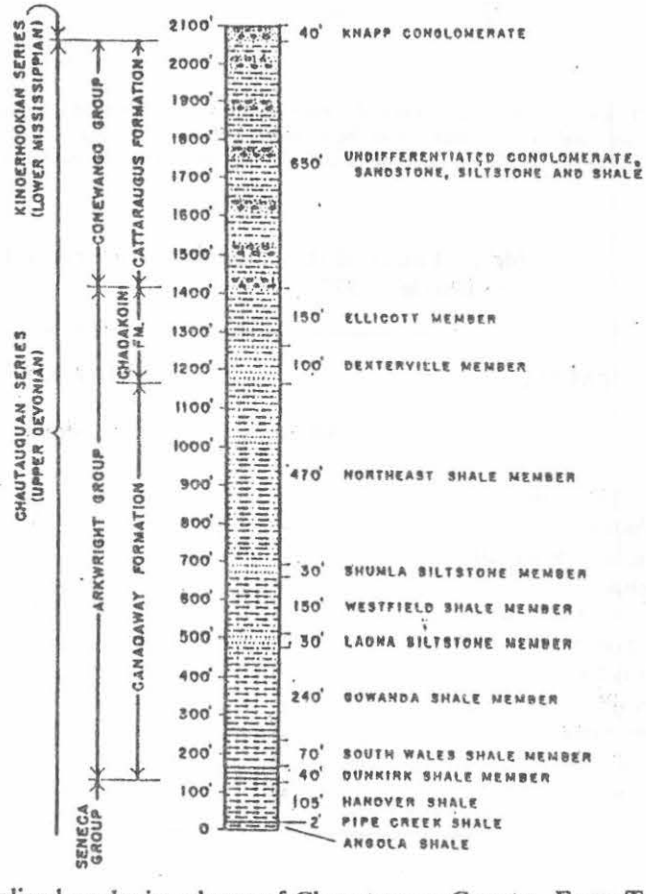


Figure 12. Generalized geologic column of Chauatauqua County. From Tesmer, 1963.

a thickening -, thinning - upwards sequence of turbidites deposited from unconfined sheet-like turbulent flows, perhaps originating from storm activity on the shelf to the east.

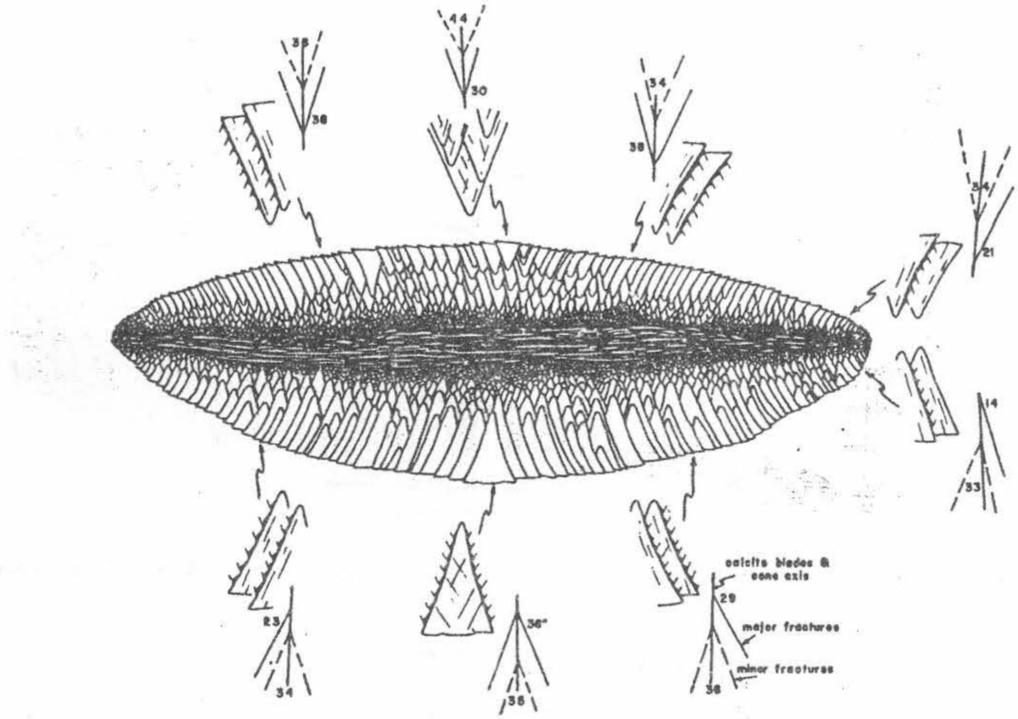


Figure 13. Drawing of a cross section of a cone-in-cone concretion. Also diagramed are the cone axes (coincident with the calcite blades) and the fracture systems as measured in several places in the concretion. Cone axes are perpindicular to bedding in surrounding shale and the medial silt. From Gilman and Metzger, 1967.

Table 2. Fauna of the Chadakoin formation, Chautauqua Gorge. From Burrier, 1977.

FAUNA	SEDIMENT TYPE		
	Siltstone	Shale	Coquina
Camarotoechia	*X*	*X*	*X*
Productella	*X*	*X*	*X*
Pugnoides duplicatus	*X*		*X*
Cyrtospirifer	*X*	*X*	*X*
Leptodesma	*X*	*X*	
Bryozoans	*X*		*X*
Nucloidea	*X*		
Goniophora	*X*		
Grammysia	*X*		
Ambocoelia	*X*		*X*
Mytilarca	*X*	*X*	*X*

PRE-GLACIAL VALLEY: The topographic map (Fig. 14) of the part of Chautauqua Creek known as *THE GULF* shows a pronounced change in valley shape; a narrow steep walled bedrock gorge in the region where we will start our hike, to a wide flat floored valley about a mile downstream. The north end of this wide valley segment is again constrained between narrow bedrock walls. Post-glacial erosion by Chautauqua Creek has carved a channel that in part cuts across a pre-glacial valley that was subsequently filled with glacial deposits. Muller (1963) interprets these as kame deposits. The extent and orientation of the older channel is not certain.

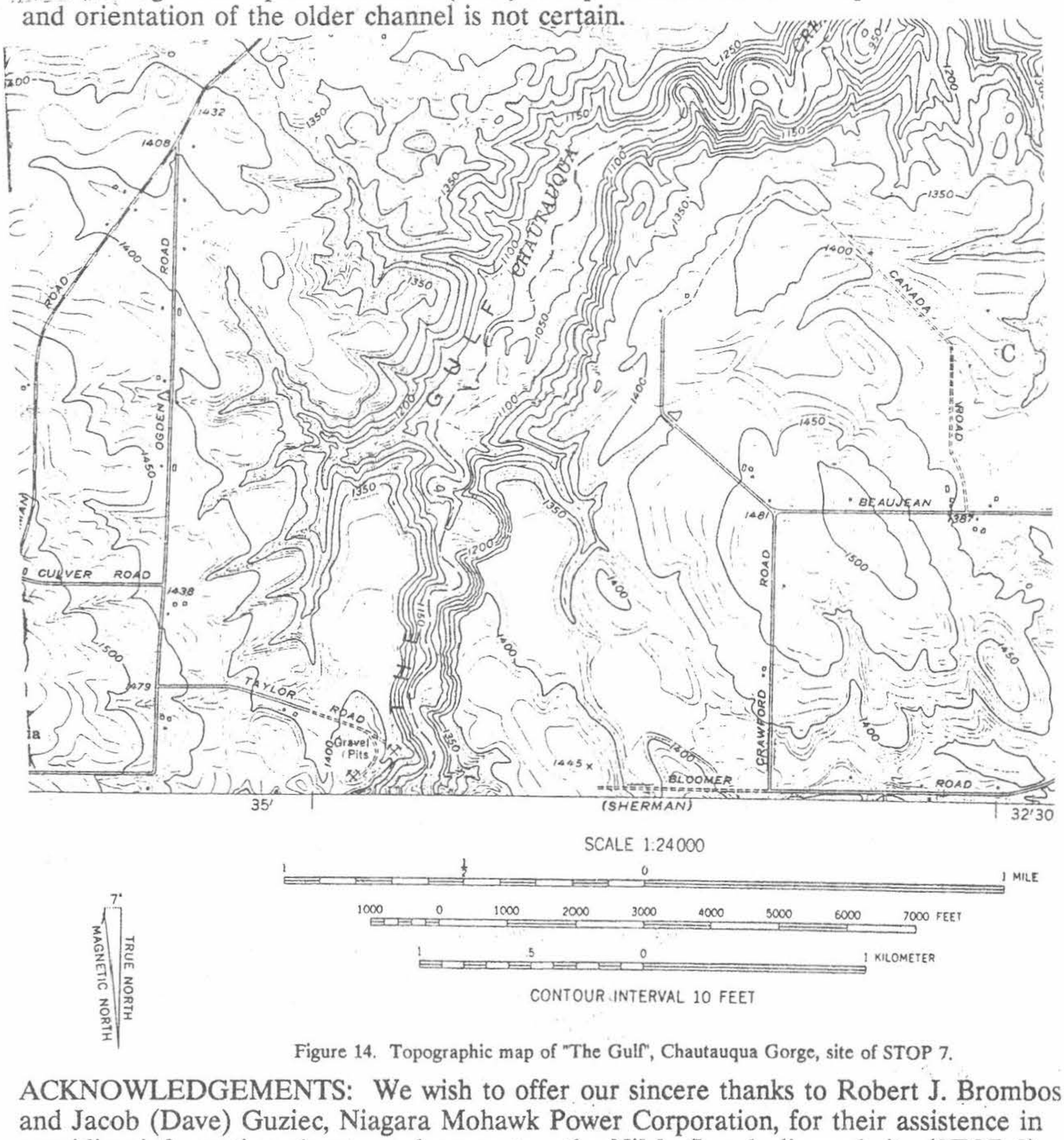


Figure 14. Topographic map of "The Gulf", Chautauqua Gorge, site of STOP 7.

ACKNOWLEDGEMENTS: We wish to offer our sincere thanks to Robert J. Brombos and Jacob (Dave) Guziec, Niagara Mohawk Power Corporation, for their assistence in providing information about - and access to - the NiMo fly ash disposal site (STOP 2). We also thank Mr. Don Frame for allowing us access to his reclaimed disposal site (STOP 3). We also received invaluable information from Mr. Tony Eberhart and Carl Czora, Buffalo Branch, U. S. Army Corp of Engineers, for which we are grateful.

REFERENCES CITED

Adams, J., 1982, Stress-relief buckles in the McFarland quarry, Ottawa: Can. J. Earth Sci., v. 19, p. 1883-1887.

Burrier, D., 1977, The paleoecology of the Chadakoin Formation of Chautauqua County, New York: Unpublished M.S. thesis, SUNY, Fredonia.

Cadwell, D. H., ed., 1988, Surficial Geologic Map of New York: New York State Museum-Geological Survey, Map and Chart 40.

Coates, D. F., 1964, Some cases in engineering work of orogenic stress effects: in Judd, W. R., ed., State of Stress in the Earth's Crust, New York, Elsevier, P. 679-688.

Cushing, D. H., Fairchild, H. L., Ruedemann, R., and Smyth, C. H. Jr., 1910, Geology of the Thousand Islands region: New York State Museum and Science Service Bull. v.145, 185pp.

Dreimanis, A., 1969, Late-Pleistocene lakes in the Ontario and Lake Erie basins: Internat. Assoc. Great Lakes Research, 12th Conf. Great Lakes, Proc., p. 170-180.

Engelder, T., 1982, Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America?: Tectonics, v. 1, p. 161-177.

---- 1985, Loading paths to joint propagation during a tectonic cycle: An example from the Appalachian Plateau, U.S.A.: Jour. Struct. Geol., v. 7, p. 459-476.

---- 1986, The use of joint patterns for understanding the Alleghanian Orogeny in the Upper-Devonian Appalachian basin, Finger Lakes district, New York: in Guidebook, 58th Annual Meeting, New York State Geol. Assoc., Ithaca, p. 129-143.

Engelder, T. and Geiser, P. A., 1980, On the use of regional joint sets as trajectories of paleostress fields during the development of the Appalachian Plateau, New York: Jour. Geophys. Res., v. 85, p. 6319-6341.

Gilbert, G. K., 1886, Recent geological anticlines: Am. Jour. Sci., ser. 3, v. 32, p. 324.

----1888, Post-glacial anticlinal ridges near Ripley, New York and Caledonia, New York: Am. Assoc. Sci. Proc., v. 40, p. 249-250.

Gilman, R. A. and Metzger, W. J., 1967, Cone-in-cone concretions from western New York: Jour. Sed. Pet., v. 37, p. 87-95.

Gravenor, C. P., 1975, Erosion by continental ice sheets: Am. Jour. Sci., v. 275, p. 594-604.

Gross, M. R., 1989, Fractures in the Lockport dolomite of western New York and southern Ontario: unpublished M. S. Thesis, The Pennsylvania State University, 98 pp.

Hancock, P. L. and Engelder, T., 1989, Neotectonic Joints: Bull. Geol. Soc. Am., v. 101, p. 1197-1208.

Metzger, W. J., 1974, Selected problems of environmental geology in Chautauqua County, New York: in Guide Book, 46th Ann. Meeting, New York State Geol. Assoc., Fredonia, NY, p. E-1 - E-16.

Miller, M. A. and Lash, G. G., 1984, Turbidite cycles in the Upper Devonian Northeast shale, western New York - evidence for smoothing of depositional topography: Northeast section, Geol. Soc. Am., Abstracts with Programs.

Muller, E. H., 1963, Geology of Chautauqua County, New York, Part 2, Pleistocene geology: New York State Museum and Science Service Bull. 392: Albany, New York.

Pettijohn, F. J., 1975, Sedimentary Rocks: Harper and Row, New York, 3rd edition, 628 pp.

Sbar, M. L. and Sykes, L. R., 1973, Contemporary compressive stress and seismicity in eastern North America: An example of intra-plate tectonics: Bull. Geol. Soc. Am., v. 80, p.1231-1264.

Selley, R. C., 1988, Applied Sedimentology: Academic Press, London, 446 pp.

Tesmer, I. H., 1963, Geology of Chautauqua County, New York, Part I, Stratigraphy and Paleontology: New York State Museum and Science Service Bull. 391: Albany, New York.

Williams, H. R., Corkery, D. and Lorek, E. G., 1985, A study of joints and stress-release buckles in Paleozoic rocks of the Niagara Peninsula, southern Ontario: Can J. Earth Sci., v. 22, p. 296-300.

ROAD LOG

TOTAL MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0		Proceed to Temple Street entrance to the college; turn right (north).
0.2	0.2	Turn left into the parking lot of St. Paul's Church.
		<u>STOP 1.</u> Follow the path to Canadaway Creek to look at pop-up anticline in the Gowanda shale.
		Return to cars and turn left (north) on Temple Street.
0.4	0.1	Intersection: Proceed ahead on Temple Street.
1.0	0.6	Thruway overpass
1.4	0.6	Turn left on Willow Road.
1.9	0.5	Turn left on Route 5.
4.3	2.4	Turn left on VanBuren Road (Fireside Restaurant on corner)
4.8	0.5	Turn right into Niagara Mohawk fly ash disposal site.
		<u>STOP 2.</u> (permission to visit this facility may be obtained from Mr. Jacob (Dave) Guziac, NiMo, 106 Point Dr. North, Dunkirk, NY 14048 716-366-2885).
		Follow gravel road to DON FRAME disposal site.
		<u>STOP 3.</u> Reclaimed fly ash disposal site.
		Follow gravel road to Route 5 (reset odometer)
0.0	0.0	Turn left (west) on Route 5.
1.7	1.7	Turn right into Lake Erie State Park; keep right to beach area parking lot.
2.2	0.5	<u>STOP 4.</u> Lake Erie shoreline; environmental and geological features
		Return to Route 5.
2.7	0.5	Turn right (west) on Route 5.
3.3	0.6	Turn left (south).
3.9	0.3	Thruway overpass
4.2	0.3	Railroad underpass, DANGER: WATCH FOR STOP LIGHTS!

5.2	1.0	Intersection with Route 20 at Brocton: Continue straight ahead. Rought 20 lies along the top of the Lake Warren beach ridge; this will be discussed at the next stop.
5.9	0.7	View of Lake Whittlesey beach terrace at 10 o'clock
6.1	0.2	Turn right (west) on Webster Road.
6.7	0.6	Turn right into gravel pit
		STOP 5. Lake Whittlesey beach deposit.
6.9	0.2	Turn right (west) on Webster Road.
7.7	0.8	Turn left (south) on Fay Road.
8.2	0.5	Turn left (east) on Elliocott Road.
8.5	0.3	Turn right (south) on Thayer Road.
9.4	0.9	At about this point we leave the lake plain and proceed up the escarpment.
9.9	0.5	Turn left into Thayer Road Overview.
		STOP 6. LUNCH.
10.2	0.3	Turn right (north) on Thayer Road.
11.7	1.5	Turn left (west) on Ellicott Road.
12.6	0.9	Turn north on Cemetary Road.
12.8	0.2	Intersection (stop). Proceed straight ahead.
13.3	0.5	Turn left (west) on Route 20: Village of Portland. Route 20 follows along the lake Warren beach as we proceed toward Westfield.
17.9	4.6	Entering Westfield.
18.9	1.0	Stop light: continue ahead.
19.2	0.3	Stop light: continue ahead crossing bridge over Chautauqua Creek.
19.5	0.3	Turn left (south) on Chestnut Street.
22.8	3.3	Bear left at Ogden Road
24.2	1.4	Turn left on Taylor Road (white house on the corner).
24.8	0.6	Park at the end of the road, across from gas well.
		STOP 7. Hike into Chautauqua Gorge. Requires wading in water.
		Return to Westfield, then to Fredonia via Route 20.

TOTAL ROUND TRIP APPROXIMATELY 53 MILES.

SUBMARINE EROSION AND CONDENSATION IN A FORELAND BASIN: EXAMPLES FROM THE DEVONIAN OF ERIE COUNTY, NEW YORK.

By

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

and

GORDON C. BAIRD
Department of Geosciences
SUNY College at Fredonia
Fredonia, New York 14063

INTRODUCTION

Geologists have long recognized the importance of unconformities in the stratigraphic record. Sloss (1963), for example, used six widespread unconformities to subdivide the North American stratigraphic succession into five major sequences or megasequences. This seminal work was, in some ways, the forerunner of sequence stratigraphic analysis which has recently resulted in a revitalization of stratigraphy (see articles by Vail and Mitchum, 1977; Vail et al., 1977; Posamentier et al., 1988; Van Wagoner et al., 1988; Baum and Vail, 1988; see Fig. 1). It is generally accepted that the stratigraphic record is highly incomplete and that much time is represented at some bed contacts. However, it is not always clear where these surfaces exist within stratigraphic sections. Many erosion surfaces, particularly in marine mudstone successions, are very subtle and can only be recognized with detailed observations. Nonetheless, many such surfaces are traceable over vast areas, forming the basis for a new dynamic view of stratigraphic processes within epicontinental sea and foreland basin settings.

Traditionally, geologists have associated erosion surfaces with major shallowing events that produced subaerial exposure. By definition, the sequence boundaries recognized by seismic stratigraphers (see Van Wagoner et al., 1988) represent, at least in part, subaerial erosion surfaces (Fig. 1). Obviously, subaerial processes of stream entrenchment and freshwater solution of carbonates are key agents in producing the major unconformities. A number of agents may also produce extensive truncation of older sediments in submarine environments; major storm waves, deep bottom currents, and internal waves may act to erode or winnow sediments on the shallow seafloor.

Condensed sections are relatively thin, time-rich intervals that are often widespread and may correlate laterally with much thicker sedimentary successions. These intervals, which range from millimeters to a few meters in thickness, are characterized by multiply reworked fossils and clasts, commonly showing a mixture of biofacies; fossils heavily corroded and may be fragmented. They may show evidence of prefossilization i.e., a phase of early diagenesis within sediment followed by exhumation. Diaclasts, such as hiatus concretions, reworked pyrite, and reworked fossils may be common within condensed beds. Biostratigraphically condensed intervals may show a mixing of zonal fossils; they are often enriched with respect to zonally important index fossils, such as conodonts. Chemically resistant skeletal elements such as phosphatized fish bone and teeth, are common in condensed beds, at least in post-Silurian time. The presence of certain early diagenetic minerals, such as phosphate, hematite, and glauconite is also indicative of sedimentary condensation.

Sedimentary condensation in siliciclastic dominated sedimentary environments results from two major processes: (1) winnowing and bypass of fine-grained sediments due to persistent or episodic current action; and (2) sediment starvation in marine basins as a result of nearshore sediment entrapment. In carbonate depositional systems, condensation may result from a restriction or near elimination of carbonate production in the absence of input of extrabasinal sediment.

Surfaces of erosional truncation and condensed sections occur predictably within sedimentary cycles (Fig. 1). Erosion surfaces should be associated with rapid relative sealevel drop. Conversely, condensed horizons are associated with rapid relative sealevel rise and transgression, which commonly produces offshore sediment starvation. The rate of fall may exceed local subsidence leading to exposure of formerly inundated areas and subaerial erosion. Furthermore, the lowering of wave base and of storm wave base may produce erosional truncation of sediments, particularly muds, which accumulated initially below the reach of wave erosion. This may result in erosion and redeposition of muds and silts by storm-generated gradient currents into deeper portions of the basin. The gradient currents themselves may produce erosive effects in some areas, particularly in regions of slopes, where minor bypass channels may develop that permit more rapid basinward sediment transfer of fine-grained sediments across or down gently sloping ramps and into basinal regions. Consequently, relatively thick and "time-poor" successions in mid-basin depocenters should be expected to coincide with erosion on the basin margin. Condensed sections would not be predicted to occur near the bases of sedimentary sequences associated with sealevel fall.

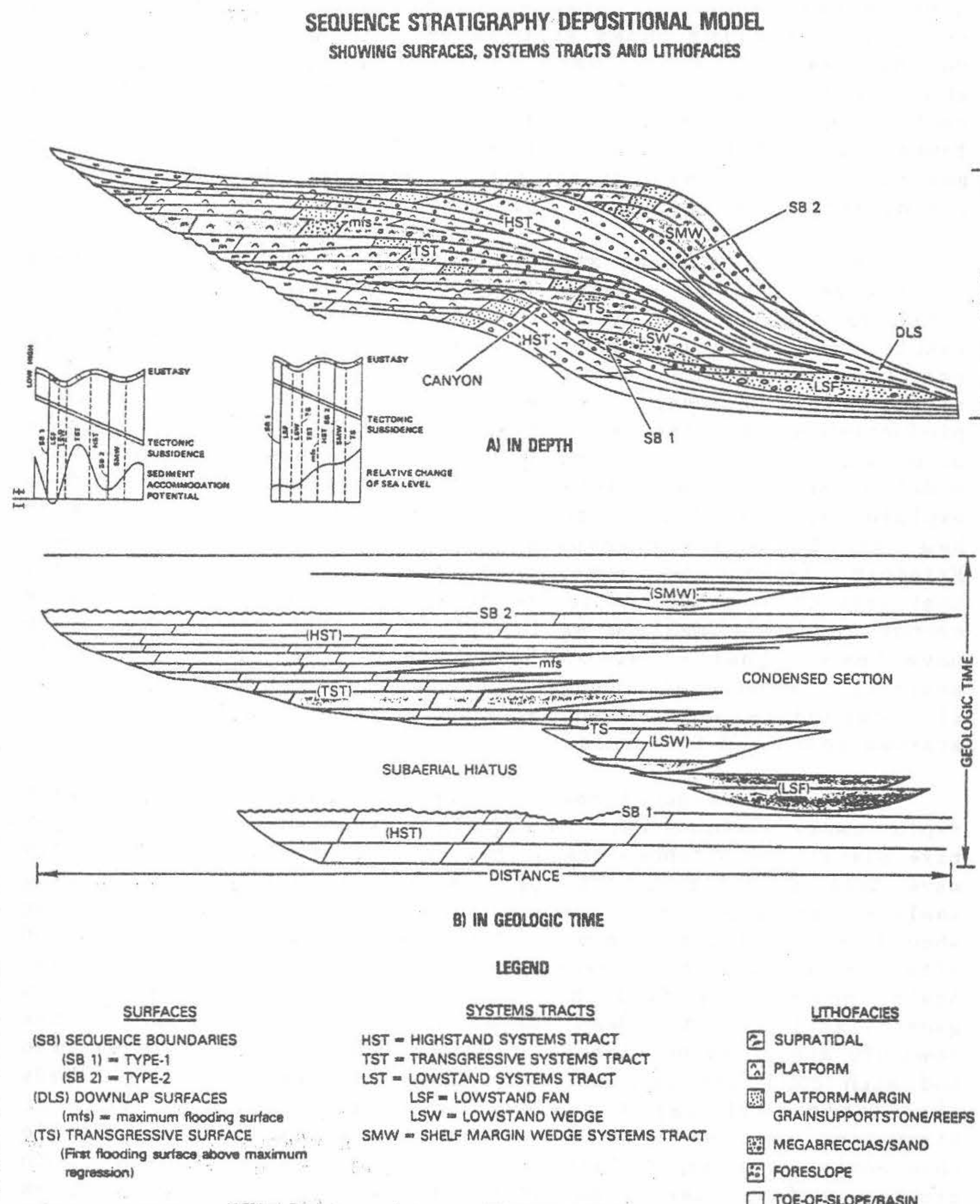


Figure 1. Idealized sequence stratigraphy model showing key surfaces and systems tracts. Sequence boundary 1 (SB-1) is a type 1 (incised) boundary, SB-2 is a type 2 sequence boundary. Note that the sequences are separated by an unconformity representing exposure during times of maximum rate of sea level fall - this unconformity can be traced to continuity offshore. This correlative conformity is a time boundary. From Baum and Vail (1988).

Nonetheless, as we illustrate in this paper, condensed beds, closely resembling those associated with sediment-starvation during sealevel rise, were formed during early phases of shallowing cycles. Such condensed beds at the bases of coarsening-up cycles often appear as "precursors," in terms of fauna, to the fossil assemblages which characterize the regressive maximum of a given sedimentary cycle. Precursor beds of this sort are difficult to explain in traditional models.

The sequence model also predicts that rapid relative rises in sealevel will be associated with sedimentary condensation in offshore areas. Relative sediment starvation will commonly result from rapid transgression, which drowns rivers, producing estuaries and bays that serve as collecting traps for siliciclastic sediment. In areas of shallow water where carbonate production may be high, siliciclastic sediment starvation could be associated with the development of limestone beds. Indeed, this model of nearshore siliciclastic alluviation has been proposed to explain thin, continuous carbonate units in the Devonian rocks of New York State and elsewhere (see McCave, 1973; Johnson and Friedman, 1969). We have argued (Brett and Baird, 1985, 1986) that most thin carbonates in the Middle Devonian of New York State cannot represent maximum marine flooding, but rather appear to have been deposited at or immediately after major drops of sealevel. Nonetheless, certain phosphatic bone-rich or conodont-rich carbonate beds do appear to represent genuine sediment-starved condensed intervals.

The sequence-model does not predict erosion associated with rapid transgression except in shoreface areas where ravinement may take place. In offshore areas, the raising of wave base and storm wave base should have the opposite effect of that seen during sealevel lowering (i.e., areas formerly affected by winnowing should be deepened to the point that they are no longer eroded and offshore areas should experience little or no wave erosion. Again, however, we find that there are many exceptions to this generalization, and indeed, major erosional truncation surfaces commonly appear to be associated with rapid relative sealevel rise and with condensed sections. In many instances, condensed beds associated with rapid sealevel rise and relative sediment starvation, may be truncated by erosion surfaces that underlie apparently deep water facies, particularly black shale. Such erosion surfaces may be marked by unusual lag deposits, such as lenticular bodies of reworked sedimentary pyrite, fish bones, and conodonts (Baird and Brett, 1986; Baird et al., 1988).

In the present paper, we illustrate a variety of condensed and erosional intervals within Middle to lower Upper Devonian rocks of Erie County, New York. We find that most erosion surfaces and condensed horizons are associated with rapid changes in sealevel; these include both rapid drops associated with the falling inflection point (on a sinusoidal sealevel fluctuation curve) and rapid sealevel rises associated with the rising inflection point (see Posamentier et al., 1988; Van Wagoner et al., 1988). However, not all such surfaces are associated with major fluctuations in sealevel. For example, it is possible for a condensed interval to occur within a carbonate simply by poisoning of the "carbonate factory" by sedimentary events such as the influx of volcanic ash.

GEOLOGIC SETTING

Tectonics and Paleogeography

The Middle and lower Upper Devonian strata of western New York State consist of carbonates and fine-grained siliciclastics belonging to the Onondaga Group, Hamilton Group and Genesee Formation (for summary discussions see, Rickard, 1981; Woodrow et al., 1988; Fig. 2). These sediments accumulated in shallow shelf to deeper basinal environments along the northwestern margin of the Appalachian foreland basin.

Precise paleogeographic positioning of North America during the Middle Devonian has been rendered uncertain by the pervasive Permo-Carboniferous overprinting of paleomagnetic signatures. Kent (1985) suggested that the Appalachian Basin was positioned at or near the paleoequator. However, more recent reconstructions place the area between 10 and 15^{o} S latitude (Van der Voo, 1988). A humid, possibly, monsoonal, climate would have prevailed. Devonian seas in the Appalachian Basin were of normal salinity. Through much of Middle and Late Devonian time these seas were density-stratified with anoxic waters in deepest portions of the Appalachian basin. Thermohaline density-stratification may have resulted from restriction of circulation and/or influx of fresh water from the Catskill deltaic complex (Ettensohn, 1985; Woodrow, 1985).

Acadian orogenic activity during the Middle and Late Devonian probably resulted from oblique convergence between Laurentian and the Avalon terranes (Ettensohn et al., 1988). Thrust loading was associated with the Acadian crustal shortening and northwestward migration of the Appalachian foreland basin. Erosional denudation of the Acadian source terranes resulted in gradual and progressive infilling (to overfilling) of the Appalachian Basin, thus producing the Catskill deltaic complex (Ettensohn, 1987).

General Stratigraphy

The early Middle Devonian (Eifelian) Onondaga limestone rests with a major erosional unconformity - the so-called Wallbridge unconformity (Dennison and Head, 1975)- on upper Silurian dolostones. This is a second order unconformity recognized by Sloss (1963) as the boundary between the Tippecanoe and Kaskaskia sequences. The surface is irregular with a relief of up to a meter, and reflects a karstified surface developed on the Silurian carbonate during a major, second order eustatic drop in relative sea level (Johnson et al., 1985). Minor amounts of late Early Devonian Oriskany Sandstone occur locally in pockets along the unconformity.

The Onondaga Limestone comprises about 30-50 meters of shallow shelf carbonates. The Edgecliff Member is characterized by crinoidal grainstones and bioherms, while wackestone and packstone with abundant nodules and beds of chert, typify the Clarence and Moorehouse Members.

The Onondaga carbonates accumulated in near-wavebase to deeper subtidal environments prior to the onset of the major phase of the Acadian Orogeny. Widespread metabentonitic ash layers within the upper beds of the Onondaga herald the onset of renewed tectonism in southeastern source areas. The upper Onondaga displays a relative deepening-upward succession. Upper beds of argillaceous, dark, micritic limestone are abruptly overlain by black, fissile and organic-rich shales of the lower Hamiton Group (Marcellus, Skaneateles formations). Hamilton siliciclastics represent the first major pulse of terrigenous sediments from the Catskill deltaic complex.

Upper Hamilton beds (Ludlowville, Moscow Formations) in western New York are fossiliferous grey, commonly calcareous mudstones that record general, but not progressive, shallowing. A series of major and minor cycles are recognized in western New York (Brett and Baird, 1985, 1986); where best developed each cycle displays a gradation from dark gray calcareous mudstones to fossiliferous pelletal wackestone-grainstone deposits. These limestones, deposited at or immediately following major shallowing events, form widespread marker units in the Hamilton Group that have been used to map component formations; in ascending order, these are the Stafford Limestone at the base of the Skaneateles Formation; the Centerfield Limestone at the base of Ludlowville Formation, and the Tichenor Limestone at the base of Moscow Formation (Figs. 2, 3).

Subsymmetrical shale-limestone-shale cycles in the Hamilton of western New York have been interpreted as 4th and 5th-order shallowing-deepening cycles (Brett and Baird, 1985, 1986). Carbonate beds appear to correlate approximately with the siltstone and sandstone caps of coarsening-upward cycles in central New York. The relatively sharp basal contacts of major limestone or sandstone units also can be considered to represent sequence or at least subsequence boundaries. Toward the basin margin the largest-scale cycles appear to become asymmetrical deepening-upward sequences.

The top of the Hamilton Group is marked by a major unconformity in western New York. The angular unconformity documented at this level (Brett and Baird, 1982) reflects major tectonic arching of the northwestern margin of the Appalachian

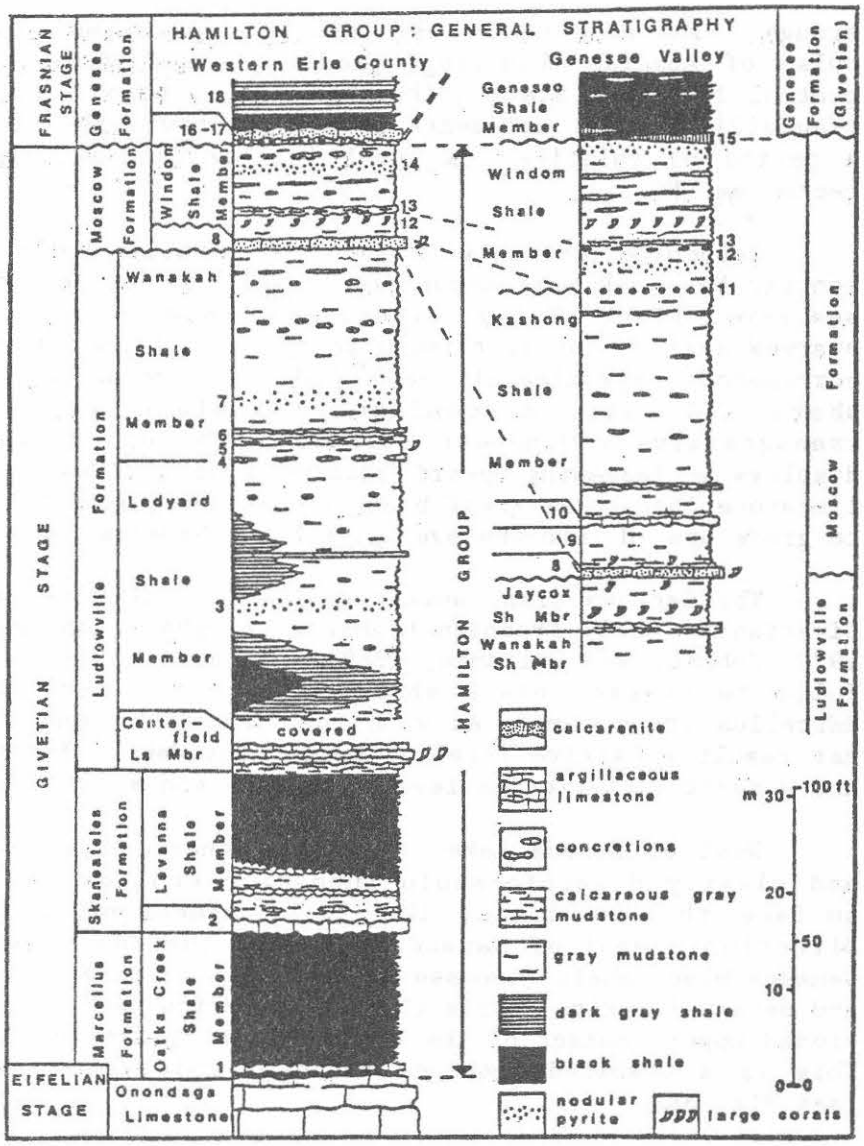


Figure 2. Stratigraphy of the Middle Devonian Hamilton Group and part of the overlying Genesee Formation. Upper Hamilton and basal Genesee beds in the Genesee Valley are correlated with condensed equivalent deposits in Erie County New York. Numbered units include: 1) basal Marcellus fossil-rich beds (concealed west of LeRoy, Genesee Co.); 2) Stafford Limestone Member; 3) Alden Pyrite bed; 4) Mt. Vernon Bed 5) Pleurodictyum beds; 6) Murder Creek trilobite bed; 7) lower Wanakah pyrite beds; 8) Tichenor Limestone Member; 9) Deep Run Shale Member; 10) Menteth Limestone Member; 11) Sub-Windom discontinuity-related phosphatic lag bed; 12) Bayview Coral Bed; 13) Smoke Creek trilobite Bed; 14) Penn Dixie Pyrite Bed; 15) Leicester Pyrite Member; 16-17) North Evans bone bed and Genundewa Limestone; 18) West River Shale Member.

trough. The same tectonism, possibly associated with a renewed pulse of Acadian thrusting, produced a synsedimentary bulge in central New York State ("Chenango Valley high") that served to trap siliciclastic sediments in eastern New York (Heckel, 1973). A period of relative sea level fall coincided with this minor tectonism (Fig. 3).

Carbonate sediments of the upper Givetian Tully Formation of central New York were deposited in the ensuing period of gradual sea level rise. Tully carbonates formed on a siliciclastic-starved shelf. The relatively sharp base of the Tully limestones corresponds approximately to a 3rd order sequence boundary. A sharp mid Tully disconformity surface may represent a transgressive ravinement surface. The upper Tully clearly displays a deepening-upward pattern from shallow, near-wavebase limestone into outer-shelf black limestones which, in turn, appear to grade upward into the overlying black Genesee Shale.

The Genesee black anoxic sediments constitute a major late Givetian to early Frasnian highstand (Taghanic onlap of Johnson, 1970; Johnson and Friedman, 1969). In many ways the change from Tully to Genesee black shales mimics the earlier Onondaga-Marcellus transition. As with the latter, it appears to be the net result of active foreland basin subsidence (Ettensohn, 1985) and a major eustatic sea level rise (Johnson et al., 1985).

West of Seneca Lake, the Tully/Genesee contact is a sharp and clearly disconformable submarine erosion or dissolution surface that truncates the Tully Limestone in a westward direction. West of Canandaigua Lake the Tully is absent and Genesee black shale (Geneseo Member) with associated reworked bone and detrital pyrite debris (Leicester Pyrite Member) occurs on the eroded upper contact of the Windom Shale (Brett and Baird, 1986). This is a combined sequence boundary and transgression surface (see Fig. 3B).

The Geneseo black and dark gray maximum-highstand shale facies is followed by an upward-shallowing trend through the gray silty Penn Yan Member. A fairly major sea level drop is marked by the Crosby Sandstone in the Finger Lakes region. Crinoidal bone-rich lag beds of the North Evans Limestone in Erie County (Figs. 2, 3; STOPS 2, 3, 8) may also correspond to this regression event. If so, the Crosby-North Evans event may represent a major subsequence boundary. Subsequent marine flooding (rapid deepening) produced condensed pelagic limestones of the Genundewa Limestone, followed by dark gray shales of the West River Member (see Brett and Baird, 1982; Baird et al., 1988).

DISCONTINUITY SURFACES AND CONDENSED DEPOSITS ASSOCIATED WITH INITIAL SEALEVEL DROP: "PRECURSOR BEDS"

As noted the Middle Devonian Hamilton Group has been subdivided into a series of shallowing-deepening cycles, marked in western New York (see Brett and Baird, 1985, 1986). Many of these cycles display abrupt bases along which relatively condensed, shell-rich, oxic facies are juxtaposed sharply on dark gray to black offshore, dysoxic shale deposits. This appears to be a case of abrupt shallowing and certainly a change from dysoxic to fully oxic environments sometimes with a diverse benthic fauna. These, sharply-based "precursor" beds appear to form the beginnings of several shallowing cycles and are traceable for up to 200 km between the Erie County region and the central Finger Lakes. These beds are normally a few centimeters to about a half meter thick and are abruptly overlain by dark gray mudstones or shales again containing a dysaerobic fauna. These beds then grade upward into the main central limestones (or corresponding siltstone or sandstone) of each cycle. Thus, for example, at the base of the well-known Centerfield cycle a relatively sharp separates a gray shell-rich bed (Peppermill Gulf Bed) from the underlying black Leiorhynchus-bearing Levanna Shale and provides a reference horizon at which the Centerfield lower boundary may be drawn (Gray, 1984, in press). A comparable bed (Stafford-Mottville "A") occurs at the base of the widespread Stafford-Mottville limestone interval. Minor cycles such as the lower Wanakah Pleurodictyum beds (see Miller, 1986, this volume) also display sharp precursor beds which are typically condensed shell-rich layers.

Precursor beds may contain faunal elements otherwise associated with the caps of the shallowing cycles. These common elements may simply record the existence of a fauna which is relatively intolerant of turbidity and therefore developed primarily during times of siliciclastic sediment shut-off. However, such faunas are usually better developed and more diverse in the cap beds of regressive or shallowing cycles.

Precursor beds are particularly well-developed in central and west-central New York and tend to die out toward the east into the thicker unconformity-capped siliciclastic, coarsening-upward cycles in central and eastern New York State. In western New York these are recorded as thin, calcareous, shell-rich markers, within shales; they are not particularly prominent in sections but they represent important condensed intervals.

Any model for the precursor beds must explain the facts that these horizons display sharp to disconformable bases, relative condensation, and an abrupt change from dysoxic to oxic, presumably deeper to shallower water. Finally, we must account for the fact that these beds appear to be connected closely in time with the development of more major shallowing cycles. In short, these are condensed and sometimes erosionally- based beds associated with initial sealevel drop or the beginnings of a shallowing cycle. However, it is also probable that such beds reflect a smaller-scale, shallowing-deepening cycle superimposed on the larger one. If so, some precursor beds may be thought of as accentuated caps of heretofore unrecognized small-scale cycles.

Keith Miller (personal communication) suggests that the abruptness of the bases of these horizons may result from the constructive interference of two different sealevel fall processes. For example, a small-scale Milankovitch band-induced eustatic sealevel fall might be superimposed on a larger scale cycle.

In the sequence stratigraphy model the precursor beds lie between early and late highstand, i.e., at the change from aggradational deep-water facies to progradational, upward-shallowing successions. The association of condensed shell-rich beds with sealevel drop is counter intuitive and demands an explanation. These beds are typically followed by thick sections of silty, sparsely fossiliferous mudstone that appear to record major influxes of siliciclastics associated with the continuing sealevel drop. The initial period of offshore sediment starvation (which is not as extensive as that for typical transgressive sediment-starved beds; see below), may record disequilibrium conditions associated with the rapid change in sealevel. For example, Posamentier et al. (1988) argue that rapid drops in base level may cause rivers to regrade to an equilibrium profile. During this period of readjustment of stream beds a considerable amount of sediment may actually be deposited in subaerial environments, a phenomenon referred to as subaerial accommodation; sediment is being trapped near the lower reaches of streams, resulting in a brief period of sediment-starvation in offshore areas. Erosion associated with the bases of precursor beds may, in part, record lowering of storm-wave base or increased scouring by gradient currents. Erosive surfaces of the bottoms of these beds are most accentuated in areas that appear to have had steep northwest-dipping slopes as in the case of shell-nodule-rich diastemic beds in the Ludlowville Formation in the Cayuga Lake region (see Baird, 1981; Baird and Brett, 1981).

EROSION AND CONDENSATION ASSOCIATED WITH REGRESSION MAXIMA (SEQUENCE BOUNDARIES)

A second type of sharp erosional surface and fossil lag bed is commonly associated with the caps of shallowing-upward cycles. This type of bed is well exemplified by the Tichenor Limestone (see Fig. 4; STOPS 2, 7). In western New York, these are discontinuity-floored coarse crinoidal and shell-rich gravel deposits. They appear to represent some of the shallowest (highest energy) deposits in the entire Devonian succession.

Such surfaces occur at the bases of thin, high-energy limestone beds in the Hamilton Group of Erie County which record

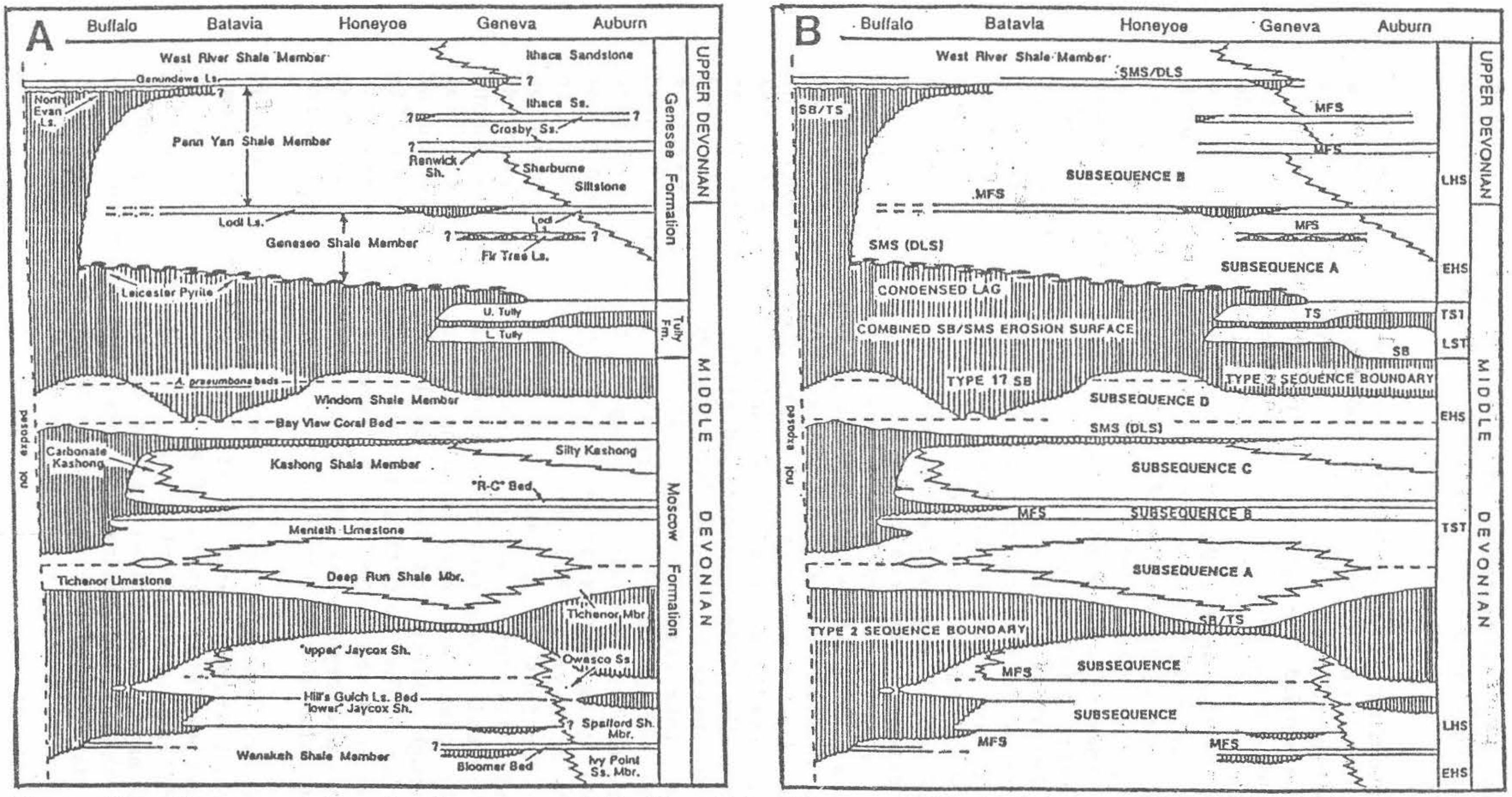


Figure 3A. Chronostratigraphy for uppermost Ludlowville, Moscow, Tully and Genesee Formations across western New York State. The most important changes from Rickard, (1975) involve major revisions of uppermost Ludlowville and lower Moscow strata (see Baird, 1978, 1979; Mayer 1989; Mayer et al. 1990, in prep.), and reinterpretation of the relationship of Leicester and North Evans lag deposits to surrounding beds and unconformities (see Brett and Baird, 1982; Baird and Brett, 1986a, b).

3B. Sequence stratigraphic interpretation of the upper Ludlowville, Moscow, Tully and Genesee Formations. The symbols include: CI=condensed interval; EHS=early highstand; LHS=late highstand; LST=lowstand; MFS=marine flooding surface; SB=sequence boundary (type 2 implies planar type erosion surgace); SMH/DLS= suface of maximum starvation or down lap surface; SB/SMS=combined sequence boundary and maximum starvation surface erosion surface (see text); TS=transgressive surface. The subsequence designation refers to sequence-like units of lesser temporal magnitude and lacking major erosional boundaries (see Brett et al. article, this guidebook for further explanation).

the immediate post-regressive maxima of cycles (Figures 3-5). The basal surfaces of such beds bevel underlying strata in a westward direction.

The most prominent erosion surface is the lower contact of the Tichenor Limestone which forms the base of the Moscow Formation by definition of Baird (1979). The Tichenor is a prominent and widespread, though very thin (10-40 cm) crinoidal packstone-grainstone unit (Fig. 4). The Tichenor occurs close to the regressive center (lowstand) of the largest-magnitude shallowing event in the entire Hamilton Group (Fig. 5). In the Seneca Lake region the Tichenor is nearly conformable with underlying calcareous silty mudstones of the upper Jaycox member and with very similar facies of the overlying Deep Run Member (Baird, 1979). These facies appear to represent slightly lower energy mud deposits that accumulated below normal wave base whereas the Tichenor pack- and grainstones reflect persistent physical reworking of carbonate gravels and sands in relatively high energy near-wave-base environments. Tichenor crinoidal debris accumulated immediately after a period of maximum shallowing. The actual regressive maximum may have been coarse siltstone or sandstone. Remnants of a sub-Tichenor siltstone have been discovered in one locality (Big Hollow Creek) between Seneca and Cayuga Lakes. The highest portions of the sub-Tichenor coarsening-upward cycle have everywhere been removed by an interval of submarine or possibly subaerial erosion. Hence, the Tichenor actually overlies an erosion surface which bevels the underlying upper Ludlowville mudstone and siltstones. The amount of section removed beneath this surface increases both east and west of the Seneca Lake region; in such areas, the upper portions of the Jaycox Member (3 to 10 m of section) have been removed (Fig. 4a). Progressive westward overstep of distinctive marker beds in the underlying Jaycox Shale member by the Tichenor has been demonstrated (Mayer, 1989; Mayer et al., in press). In extreme western sections along Lake Erie the Tichenor rests on Wanakah Shales with the Jaycox and uppermost Wanakah beds removed. The pattern of erosion suggests truncation of older sediments along the flanks (east and west dipping ramps) of the central Finger Lakes trough (Baird and Brett, 1981) during a period of extreme shallowing. The fact that the base of the Tichenor is unconformable, even in the basin center, reveals both the severity of the regression and removal of the eroded sediments by apparent southward current transport, along the axis of the foreland basin.

The Tichenor, itself, is a condensed bed consisting of multiply reworked, broken and abraded pelmatozoan ossicles, corals, and other skeletal debris (Fig. 4). In places, two or more graded beds can be observed within the Tichenor suggesting amalgamation of major tempestites (storm deposits) during which much or all of the relict sediment blanket was reworked. Very crude conodont data suggest that the Tichenor may be slightly diachronous, becoming younger outward from the axis of the Finger Lakes trough.

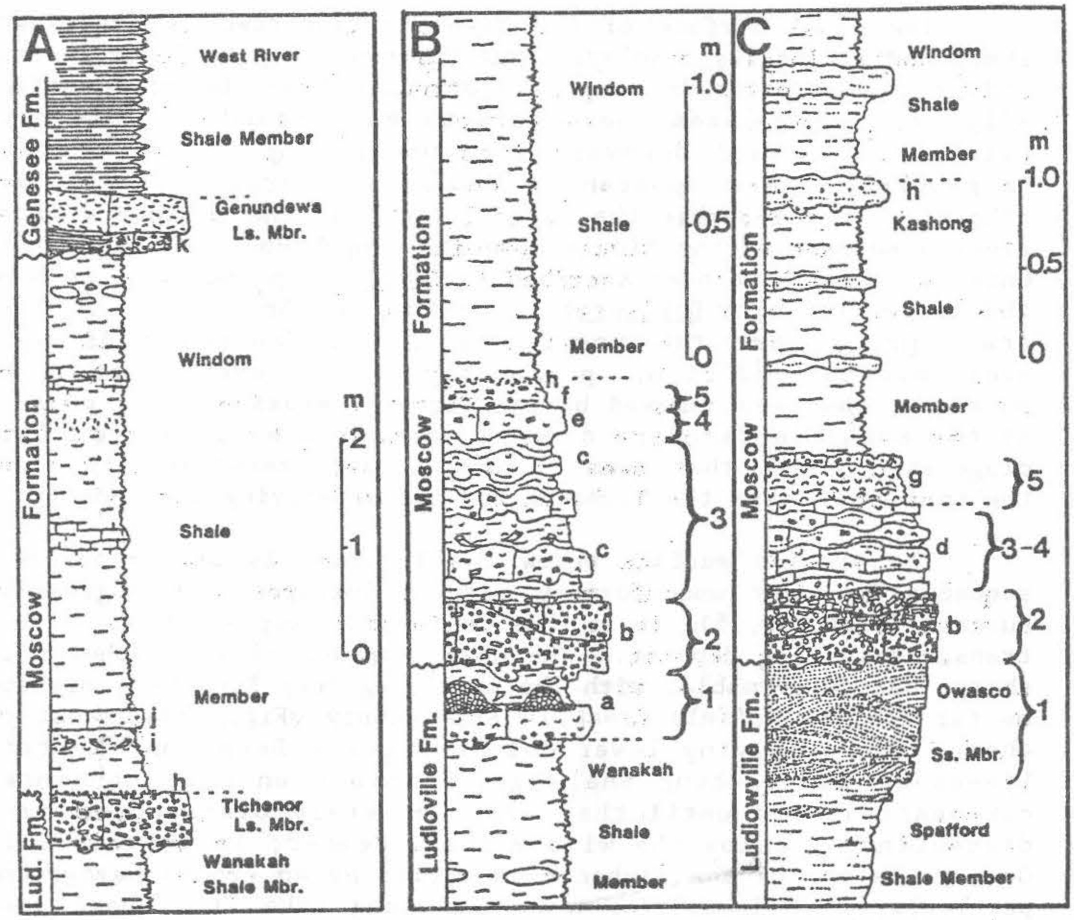


Figure 4. Stratigraphy of Moscow Formation and adjacent units. A. Stratigraphic section on Eighteenmile Creek east of (upstream from) Route 5 bridge (see STOP 2); B-C, Comparison of coeval condensed uppermost Ludlowville (Jaycox Member) and lower Moscow (Tichenor Member-Kashong Member) deposits on opposite flanks of broad depocenter in Genesee Valley-Cayuga Valley region; section B is on Buffalo Creek at Bullis Road (STOP 7) in Erie County; section C is generalized uppermost Ludlowville-lower Moscow interval for localities in the Owasco Lake Valley. Numbered divisions 1-5 denote coeval condensed chronostratigraphic units which are traceable across the depocenter (see Figure 2). The corresponding condensed units display similar facies except that Division 1 unit in west (lower Jaycox limestone bed) is represented by the Owasco sandstone east of the trough (see text). Numbered units include: a) Hill's Gulch Bed and overlying mudstone unit of Jaycox Member; b) Tichenor Limestone Member; c) condensed muddy carbonate beds equivalent to Deep Run Shale Member; d) condensed muddy carbonate beds equivalent to both Deep Run and Menteth members (see Figures 2, 3); e) Menteth Limestone Member; f) condensed calcareous mudstone facies of lower Kashong Shale Member; g) condensed facies of lower Kashong Member ("R-C" Bed of Baird, 1979); h) sub-Windom discontinuity surface marked by phosphatic debris; i) Bayview Coral Bed; j) Smoke Creek trilobite bed; k) North Evans bone-conodont bed.

The basal surface of the Tichenor Limestone is everywhere sharp and it locally displays large burrows(?) up to 10 cm across, and up to one meter in length, protruding from the sole surface (Fig. 4). In places these burrows may extend, as discrete exichnial "tubes," downward into underlying shales. These "megaburrows" are apparently devoid of scratch marks, but otherwise they resemble the large burrows on the analogous basal erosion surface of the Middle Devonian Hungry Hollow limestone in Ontario; these have been ascribed to Cruziana (probably burrows of the large trilobite Dipleura) by Landing and Brett (1987). Such prods prove that the underlying Jaycox-Wanakah muds were overcompacted and firm; presumably, up to several meters of mudstones had been removed by pre-Tichenor erosion. The details of the basal surface are commonly obscured by crusts of late diagenetic pyrite that seem to have formed preferentially along the contact between the Tichenor and the underlying mudstone.

The erosion surface below the Tichenor is analogous to a sequence boundary unconformity with a juxtaposed transgressive surface (Figs. 3b, 5); it is thus believed to represent an initial transgressive lag deposit. The upper contact of the Tichenor is sharp, but conformable, with the overlying Deep Run Shale perhaps as far west as Buffalo Creek in Erie County (Fig. 4b). West of there, the overlying lower Moscow units - Deep Run, Menteth Limestone and Kashong Shale grade into even more condensed carbonate layers until they are completely overstepped by a discontinuity below the Windom Shale Member; in western Erie County the top of the Tichenor is marked by an eroded hardground particularly at localities near and along Lake Erie Shore (see Fig. 4a; STOP 2). Hence the upper surface of the Tichenor is a disconformity representating depositional pinchout of muds onto the shallow western platform. Only the relatively major Windom deepening event allowed this disconformity to be overlapped and preserved (Figs. 3, 5).

The lower Moscow Formation (Tichenor, Deep Run Menteth and Kashong members) is approximately equivalent to a transgressive systems tract; it records a slight upward-deepening through a series of smaller-scale shallowing-upward units (Fig. 5). This interval is a highly condensed section in Erie County as a result of winnowing and bypass. Indeed, a few centimeters of limestone in the western Erie County sections may be the only vestige of this entire package which exceeds a net of 35 m (110 feet) of mudstone in the Genesee Valley-Canandaigua Lake Region (see Baird and Brett, 1981; Fig. 3). The very widespread sub-Windom phosphatic bed at the top of the Kashong Shale may represent a surface of maximum starvation (Fig. 4).

In addition to the sub-Tichenor unconformity, regressive erosion surfaces are seen at various other levels in the New York Devonian. For example, the base of the Jaycox Member which immediately (and unconformably) underlies the Tichenor in Erie County is represented by a thin (15-22 cm) crinoidal and coral-rich limestone unit (Hills Gulch bed) which mimics (and has been

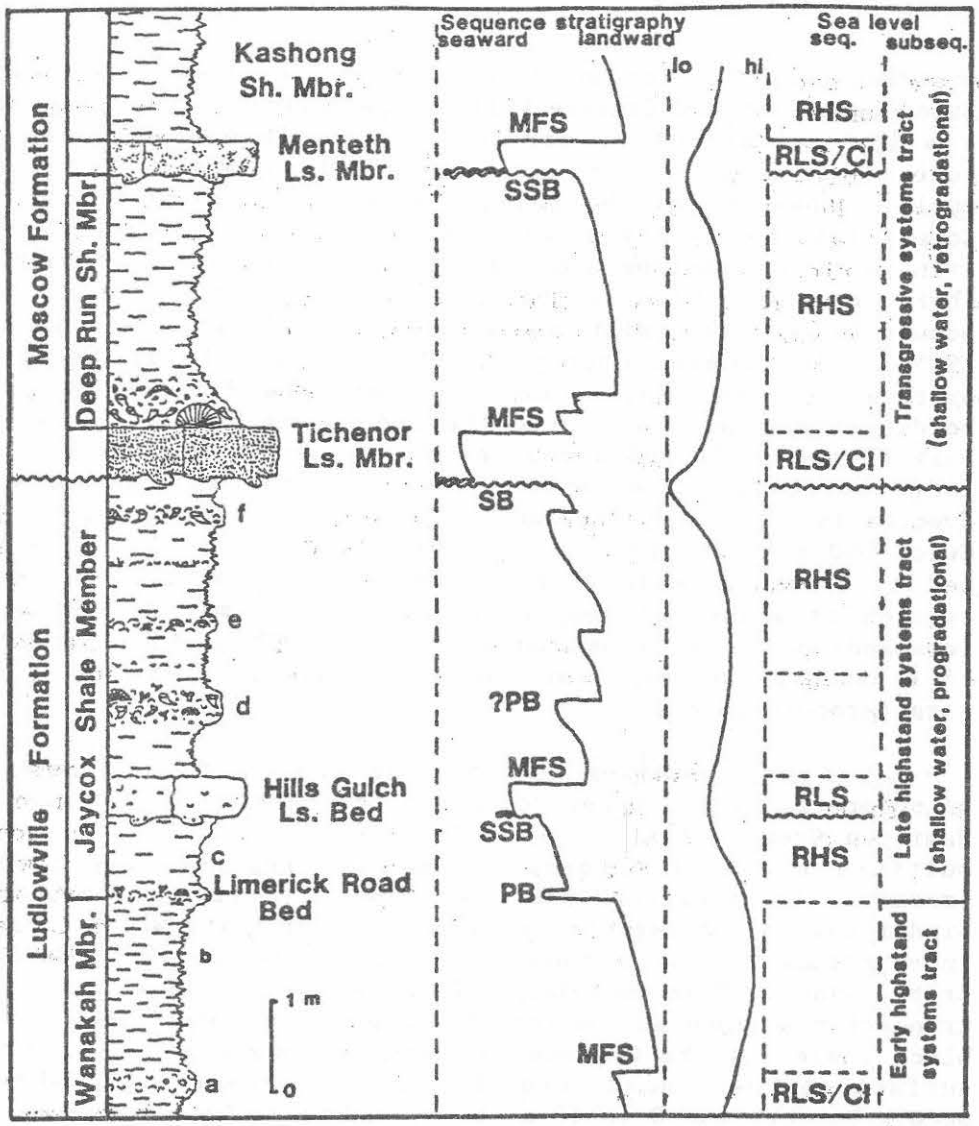


Figure 5. Sequence Stratigraphy Model applied to a portion of the upper Hamilton Group. Upper Ludlowville Formation divisions (uppermost Wanakah Member, Jaycox Member) and lower Moscow units (Tichenor, Deep Run, Menteth, and basal Kashong Members) are shown to be the result of rises and falls of sea level. Major regression event at base of Moscow Formation (discontinuity flooring Tichenor Limestone), traceable across most of New York State, probably marks a sequence boundary (SB). Other discontinuities (flooring Hill's Gulch and Menteth limestones) mark subsequence boundaries (SSB). PB = Precursor (PB) beds (see text; STOP 6). Remaining terms include: MFS=marine flooding surface; RHS=relative highstand; RLS=relative lowstand; CI=condensed interval. Units include: a) Bloomer Bed; b) upper Wanakah mudstone with diminutive dysaerobic fauna; c) barren, blocky mudstone unit; d) Green's Landing Coral Bed; e) Sponge-Megastrophia Bed; f) Cottage City Coral Bed.

mistaken for) the Tichenor itself. About 70 miles to the east of Erie County, in the Genesee Valley, the Hills Gulch bed appears as a silty calcareous, moderately fossiliferous mudstone that conformably overlies soft gray shale and appears to cap a small cycle. However, as the bed is traced westward from Livingston County (Genesee Valley area) it becomes a compact, silty, coral-crinoid rich packstone and, near Bethany, it begins to display a sharp, diastemic lower surface. Coarse crinoid and coral debris occurs in vaguely graded layers within the Hills Gulch bed (Mayer, 1989). In Genesee County the Hills Gulch horizon displays overstep of lower Jaycox beds and near the Erie-Genesee county border it begins to cut into the underlying Wanakah Shale. The base of the Hills Gulch bed resembles that of the Tichenor with large burrow prods extending down into the underlying shale. Eventually, the sub-Tichenor unconformity oversteps the Hills Gulch bed both at and west of Cazenovia Creek. The Hills Gulch bed is instructive in that it demonstrates that the pattern of erosion is evident in major sequence bases, may develop at the lowstands of lesser subsequences as well. Thus, the same pattern of erosional overstep near the basin margin is evident at many stratigraphic levels.

A final example of the regressive-type (lowstand) unconformity is the upper contact of the Moscow Formation of the Hamilton Group (Figs. 3, 6). In Erie County this is a complex multiple event unconformity, in part representing lowstand erosion. Previous work has shown that the upper contact of the Windom Shale is a regionally angular unconformity which increases in magnitude to the northwest (Brett and Baird, 1982; Baird and Brett, 1986). From Canandaigua Lake westward to Erie County, this truncation surface at the top of the Windom floors latest Givetian black shales of the Genesee Formation. However, this truncation surface can be traced into the Finger Lakes region, where it occurs beneath the 0 to 10 m-thick, middle to late Givetian Tully Limestone (see Heckel, 1973), which intervenes between the Moscow and Genesee Formations (Fig. 3).

It is evident, therefore, that some or much of the Windom truncation surface was developed prior to deposition of the Tully. Basal Tully carbonates are massive micrites representing shallow subtidal (and possibly lagoonal) shelf conditions as suggested by the occurrence of rare stromatolites (Heckel, 1973). Where the Tully is in contact with the Windom truncation surface the contact, like that of the Tichenor, is marked by large tubular burrows some of which extend downward 10-20 cm into the Windom Shale.

In western New York the entire Tully deposit has been removed by a still younger erosion surface; this latter discontinuity is of a very different nature and is associated with a period of sediment-starvation during rapid sea level rise to highstand conditions of the Genesee Formation (see Figs. 3, 6). Nonetheless, this latest unconformity is combined with the pre-Tully sequence boundary unconformity in western New York.

Sub-Tichenor, sub-Hills Gulch, and sub-Tully unconformities all involved the removal of stiff compact mud. We cannot rule out that some of this erosion, at least in basin margin areas, may have been subaerial. However, at least later stages of sediment removal were surely submarine. Dislodgement and resuspension of the firm muds may have been aided by active burrowing by organisms adapted to excavation of compact mud. Vestiges of this fauna are seen in the megaburrows at the bases of limestone beds. Abrasive scour of the seafloor may have been aided by tractional and saltational transport of skeletal sands and gravels during storms. In some cases, it is evident the preexisting burrows or furrows were present and widespread prior to deposition of the lowest calcarenite sediments. In fact, basal sediments of the limestone may have initially accumulated as tubular tempestites, i.e., infillings of deep burrows (see Wanless, 1988).

DISCONTINUITY SURFACES ASSOCIATED WITH RAPID SEA LEVEL RISE AND SEDIMENT STARVATION (SEE STOPS 1, 4, 8)

<u>Black shale-roofed discontinuities</u>

A third, and largely unrecognized, type of erosion surface is associated with the rapid-deepening phases and the bases of black shales (Figs. 3, 6-8). The lower contact of the black shales or lag beds is sharp and typically erosional and, indeed, regional truncation of the underlying strata is common. The erosion surfaces in such cases are nearly planar to undulatory and razor sharp. No burrowing of the underlying sediments is evident, as would be predicted, because the erosion apparently took place under anoxic or at least dysoxic conditions (see Figs. 6, 7). There is evidence for dissolution of carbonates at the surface and the overlying thin lag beds commonly contain chemically resistant particles such as phosphate nodules, conodonts, fish bones and teeth and in several instances, diaclasts of reworked early diagenetic pyrite (Fig. 7). Pelmatozoan ossicles may be present but other carbonate detritus is rare or absent in the lag beds. Lag beds typically range from thin sheets only millimeters thick to pods or lenses up to 20 cm thick and several meters across (Fig. 6). Internally, the lag beds may be laminated or vaguely cross laminated and may display thin partings of black shale. Small lenses of debris may occur, up to several centimeters above the erosional base, interbedded with black shales. Pyritic lag beds, typified by the Leicester Pyrite, have been described in detail elsewhere (Baird and Brett, 1986; Baird et al., 1988).

Black shale/limestone contacts (STOP 8)

Sharp, lower surfaces are characteristic of most Devonian black shale units; in many instances there is evidence for erosion and dissolution of underlying sediments and of lag beds above the erosion surfaces. Typically, the underlying beds are condensed outer shelf limestone facies. One such contact of black shale on carbonate, not seen on the present field trip, is the contact of the Union Springs (or Bakoven) black shale on the upper Onondaga Limestone (Seneca or Moorehouse members). This contact is rarely seen, but it is nearly always sharp and marked by concentrations of fish teeth and phosphate nodules on the upper surface of the Onondaga; where the succession is more nearly continuous, as in the Cayuga-Seneca Lake area, alternating 10-25 cm-thick, micritic, styliolinid ("ribbon") limestones and black shale beds (i.e., Seneca Member), underlie the black shale. Although these beds appear transitional between massive Onondaga carbonate and, overlying black, organic-rich shale, there is still a scattering of fish bones, quartz grains, and rare phosphatic granules above the uppermost ribbon limestone of the Seneca Member. East and west of this area, the basal contact of the Marcellus black shales is increasingly erosional and marked by abundant fish bones, conodonts, quartz pebbles and phosphatic nodules. Both east and west of the central Finger Lakes trough the transitional black micritic limestone/shale beds of the Seneca Member are cut out by the erosion surface, such that the phosphatic bone bed comes to rest on the beveled upper surface of the underlying Moorehouse member.

A similar but more cryptic contact, occurs at the upper surface of the Tully Limestone in the southern Cayuga Lake region. Here, again, the upper beds of the Tully (Fillmore Glen beds of Heckel, 1973) appear transitional, with decimeter-scale dark styliolinid micritic limestones alternating with dark gray to black shales through about one meter of section. The upper surface of the highest limestone bed displays a thin lag of styliolinids and silty pyritic debris.

As with the upper contact of the Onondaga Limestone, the transitional Fillmore Glen beds-and then successively lower units of the Tully Limestone, are truncated by the erosion surface to the west. Near Gorham, the lowest bed of the upper Tully Member (Bellona Coral Bed) is beveled and a thin lag of bones, crinoidal grains and minor reworked pyrite burrow tubes (Leicester Pyrite Member) rests sharply on the planar upper surface.

Still farther west, at Canandaigua Lake, the entire Tully is beveled and the overlying black Genesee Shale then rests sharply on the upper Windom Shale. As noted above, some upper Windom beds had already been truncated beneath the sub-Tully sequence boundary erosion surface. Hence, the unconformity between the Windom (upper Hamilton Group) and Geneseo is actually a compounded erosion surface marking the juxtaposition of the basal Tully, and upper Tully unconformities. How much additional erosion of Windom

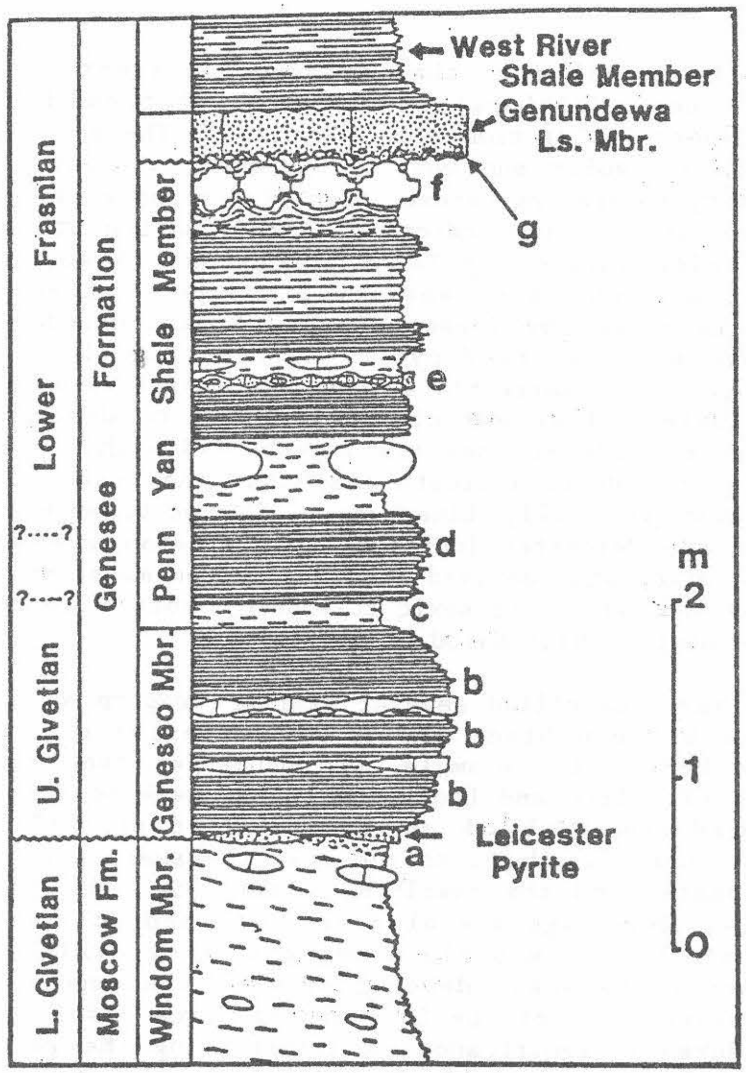


Figure 6. Stratigraphic section at waterfalls on Cayuga Creek south of (upstream from) Clinton Road bridge (see STOP 8). Important features in this outcrop include: manifest angularity of sub-Genesee unconformity within the outcrop, the Middle-Upper Devonian (Frasnian-Famennian) boundary (see interval of Question marks), and both the Leicester Pyrite and North Evans lag deposits. Lettered units include: a) brachiopod-rich bed (equivalent to Fall Brook Coral Bed?) which is overstepped (truncated) within the outcrop; b) calcareous black shale and concretionary black limestone rich in diminutive Devonochonetes and current-aligned Styliolina,; c) grey mudstone layer equivalent to Lodi Limestone Bed (to east); d) black shale unit probably equivalent to Renwick Shale (to east); e) Styliolina-rich concretionary layer ("Linden Bed" of Kirchgasser and House, 1981; Kirchgasser et al., 1988); f) grotesque concretions beneath North Evans-Genundewa limestone ledge; g) North Evans bone-conodont Bed.

beds can be ascribed to the last erosion event is not clear. However, some additional scour of the bottom during latest Givetian (post-Tully) time did take place. The sub-Tully erosion surface is irregular and heavily burrowed, whereas the combined unconformity separating Geneseo from Windom Shale is nearly planar and knife sharp. This indicates a modification of the old sub-Tully erosion surface by later processes. Also, the contact between the Windom and Genesee, like that between Tully and Genesee, is marked by lenses of fish bones, conodonts, pebbles and, above all, reworked pyrite debris (Fig. 6). The pyrite lenses have been referred to as the Leicester Pyrite Member of the Genesee Shale. They are clearly related to deposition of the black Geneseo muds as they are interbedded with the black shale (Fig. 6). Conodonts indicate also that the lenses are slightly younger than the Tully Limestone. The occurrence of reworked pyrite in the Leicester indicates that erosion and concentration of pyrite diaclasts, derived from the Windom mudstone, took place under low oxygen to (dysoxic-to-anoxic) bottom conditions (see Baird and Brett, 1986; Baird et al., 1988).

We have identified several similar erosion surfaces at the junctions between black shales and underlying concretionary carbonate beds. For example, we documented two thin limestone intervals (Fir Tree and Lodi) within the Genesee Formation which are beveled beneath black shales (Baird et al., 1988). Again, thin lenticular lag deposits occur at the sharp contacts between the carbonates and the overlying shale (Fig. 7). The pyrite-, conodont-, bone lags are also are interbedded with the basal laminae of the black shale proving that reworking of pyritic debris and black shale deposition were contemporaneous. One further observation of the Lodi and Fir Tree discontinuities is of considerable significance in interpreting the genesis of the erosion surfaces beneath black shales. In both cases the contact between the concretionary carbonates and the overlying shales is abrupt but gradational in areas where grey silty shales overlap the surface. However, as the overlying units grade, in presumed downslope direction, into black, laminated shales the contact becomes increasingly sharp and then erosional. Thus, the concretionary carbonates are truncated in an apparent downslope direction, being utterly missing in basinal section where both underlying and overlying black shales are in direct contact. These shales are separated by a cryptic discontinuity marked by a thin lag of fish debris, conodonts or silt. However, no major lenses of reworked pyrite are present, as flooring black shales contain little or no nodular pyrite.

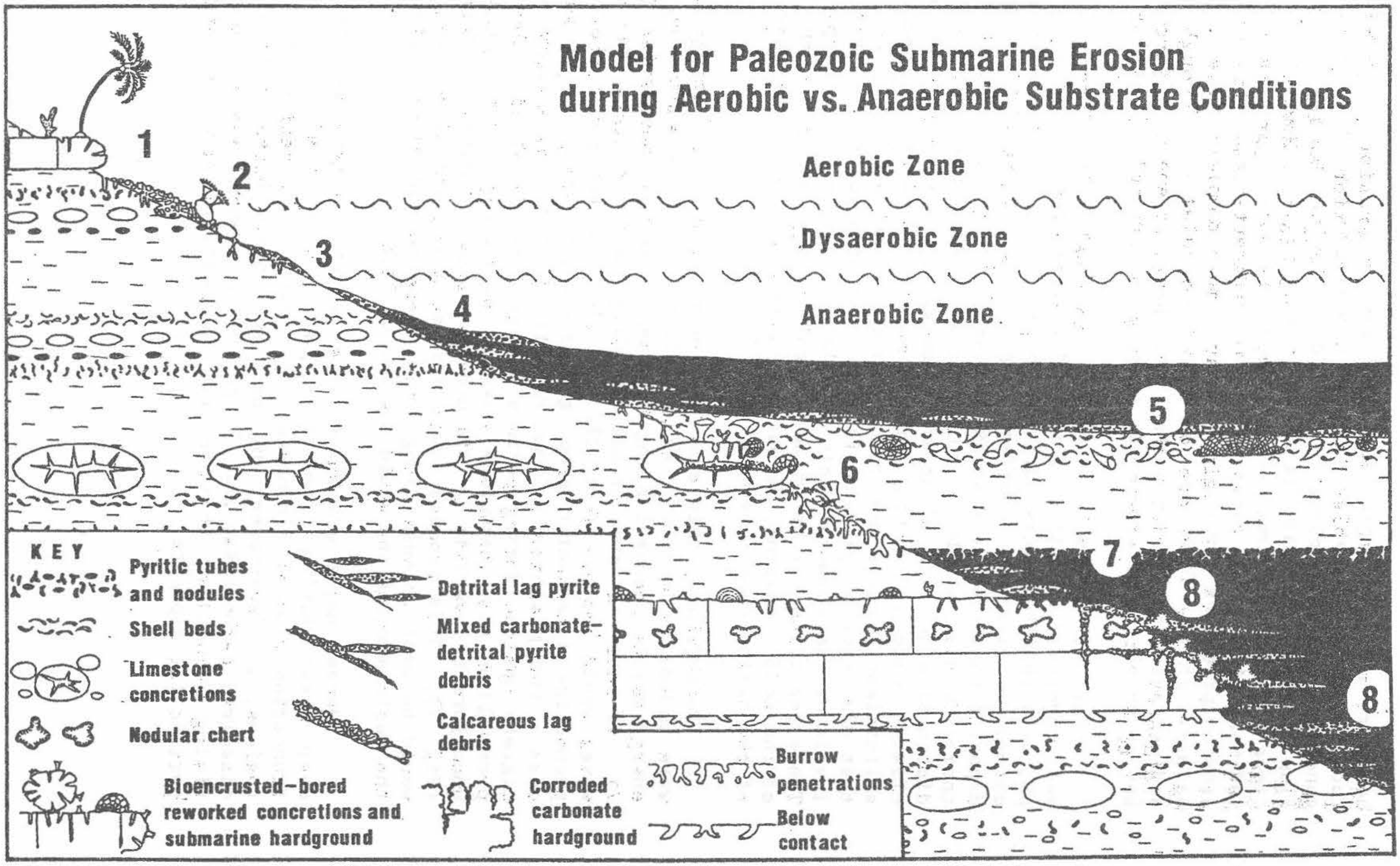
Any model for the genesis of black shale-roofed disconformities must explain the following features: 1) The erosion occurs during apparent intervals of sea level rise rather than fall; 2) they are nearly planar or runnelled surfaces overlain by dark gray or black laminated shales; 3) The surfaces are frequently underlain by concretionary argillaceous micritic carbonate or calcareous mudstones with distinctive deeper water biofacies (auloporid corals, small brachiopods, plus pelagic forms such as styliolinids and cephalopods); 4) surfaces are sharply overlain by lag deposits of geochemically resistant allochems such as phosphate, quartz, bone and conodonts, but lacking most carbonates; 5) reworked pyrite is commonly present, at least, where pyritic mudstones -serving as source beds - underlie the unconformities; pyrite should only be stable under anoxic conditions; 6) surfaces are sharp where overlain by black shales but seem to fade into conformity where overlain by gray dysaerobic deposits in the presumed upslope direction (Fig. 7).

From the standpoint of sequence stratigraphy the major black shale-roofed disconformities are not sequence boundaries but appear to correspond to surfaces of maximum starvation or downlap surfaces of seismic stratigraphers. As noted above, both the Onondaga and Tully limestones appear to comprise the bases of large-scale sequences. The upper nodular to tabular, argillaceous carbonates and interbedded shales appear to record a condensed section terminated by the abrupt change to deep-water black shales. Hence, the carbonates themselves comprise a transgressive systems tract; this sharp upper contact is a sediment-starved surface recording an increased rate of deepening to early highstand (deepest-water) conditions. During the rapid transgressive phase the carbonate factory was abruptly terminated and the offshore areas were also starved of siliciclastic sediment due to coastal entrapment. In the following highstand interval siliciclastics began to prograde seaward, providing a sedimentary wedge that downlaps onto the surface of starvation.

Rapid sea level rise also produced an elevation of the pycnocline (zone of density stratification) causing flooding of formerly shallow areas with deeper and slightly denser bottom water. Relative sediment-starvation and high productivity in surface waters enabled abundant organic matter to accumulate below the pycnocline, producing widespread oxygen deficiency on the seafloor. During shallow water intervals anoxic water was confined to basin centers but in major transgressions it expanded onto the basin margin ramps and shelves.

This pattern fits with the typical sequence model for development of surfaces of maximum starvation. A critical adjunct to the model, however, is that substantal submarine erosion may occur during times of rapid sea level rise in addition to sedimentary condensation. The thin concretionary beds that underlie the black shales reflect the onset of sedimentary condensation. Once sediment starvation reaches a critical point,

Figure 7. Generalized model for the exhumation and hydraulic dissolutional concentration of detrital pyrite on black shale roofed submarine discontinuities. A hypothetical submarine discontinuity cutting into a variety of oxic-dysoxic facies is shown being overlapped by both oxic and anoxic deposits. Discontinuity surfaces and associated lag deposits exposed respectively to oxic and anoxic waters are radically different owing to differing chemical stabilities of pyrite and carbonate in the two opposing settings (see Baird and Brett, 1986a, b; Baird et al., 1988). Features shown include: 1) Oxygenated discontinuity surface on sea bed with encrusted and bored hardground; 2) oxygenated seafloor covered by lag of coral-encrusted reworked calcareous concretions and associated shell and carbonate debris with absence of detrital pyrite; 3) dysoxic submarine substrate marked by burrows and reduced lag fraction composed partly of residual undissolved carbonate and partly of detrital pyrite and phosphatic component; 4) anoxic submarine substrate marked by absence of burrowers and bioencrusters and by complete dissolution of carbonate lag debris. Detrital pyrite is stable in this setting, but carbonate is prone to dissolution producing starved-lens concentrations of pyrite grains, bone debris, and conodonts; 5) black shale-roofed discontinuity truncating coral-rich shell bed; note that no carbonate debris survived to be concentrated on this surface despite the abundance of subjacent shell material; 6) grey mudstone-roofed erosional contact showing effects of burrowing, boring, and bioencrustation processes under oxygenated conditions. Again, pyrite is unstable and carbonate stable in this setting; 7) contact between black shale unit and overlying grey mudrock. Bioturbation associated with the onset of oxic conditions has mixed and blurred this contact. Note the dramatic stratinomic changes at the discontinuity contact timed with this change; 8) effect of prolonged exposure of limestone unit to oxygen-deficient waters; the carbonate undergoes solution producing surface pitting and solution along joints with attendant formation of a chemical lag placer of chert nodules. Detrital pyrite lenses are diachronously shingled reflecting progressive burial of the discontinuity by black basinal muds.


Model for Paleozoic Submarine Erosion
during Aerobic vs. Anaerobic Substrate Conditions
Aerobic Zone
Dysaerobic Zone
Anaerobic Zone
1
2
3
4
5
6
7
8
8
KEY
Pyritic tubes and nodules
Shell beds
Limestone concretions
Nodular chert
Bioencrusted–bored reworked concretions and submarine hardground
Detrital lag pyrite
Mixed carbonate-detrital pyrite debris
Calcareous lag debris
Corroded carbonate hardground
Burrow penetrations
Below contact

however, non-deposition may give way to erosion. One way of looking at this problem is that minor erosion, due to storm-generated gradient currents, and deep-flowing density currents, plus chemical corrosion is always taking place. However, during times of sediment input this erosion is more than balanced by sedimentation. At times of total sediment-starvation ambient erosion processes may begin to play a more dominant role and the balance shifts from one of slow accumulation through non-sedimentation to erosion without any change in energy. Hence, the key to this type of erosion is stoppage of new sediment input.

Part of the explanation for erosional loss of section may well involve dissolution of the older carbonates by corrosive (low pH, or carbonate undersaturated) bottom water. We have noted that carbonate grains are often conspicuously absent in the lag deposits that overlie disconformities (see Fig. 7). Nonetheless, there is also evidence for increased erosional energy in a distinctive pattern along basin margins since relatively heavy clasts, such as pyrite, were apparently dislodged from firm siliciclastic clays and concentrated on the sea bed. We argue that the erosion process is also concentrated along particular levels of the slope-to-basin profile and that it dies out upslope. These positions may be marked by thin pyrite lenses; in the case of the Lodi limestone, we observed pyrite lag beds on either side of the basin center (Baird et al., 1988).

Furthermore, this erosive process occurs in settings which were primarily anoxic. We postulate that wave and/or current energy may be focussed along internal water mass boundaries (Fig. 7). If interval waves propagated along the pycnoclines then these waves would impinge along the sea floor at the point where the pycnocline intersected sloping basin-margin ramps. In such a case a topographically narrow belt of erosional energy would migrate upslope during periods of rapid sea level rise when the pycnocline also expanded upward, producing a transgressive unconformity overlain by diachronous, anoxic, laminated deposits (Fig. 7). This process would produce a pattern of erosion that would be most intense along lower slope to mid slope regions of the basin margins and would die out upslope.

Hence, major erosion surfaces appear to have developed in deep water during times of relative sea level rise. The truncation probably involved: 1) sediment-starvation in distal regions; 2) development of corrosive anoxic bottom waters that dissolved carbonates; and, 3) possibly, internal waves along the rising pycnocline that eroded the seafloor and concentrated pyritic lag deposits.

Black shale - roofed furrowed contacts (see STOP 4)

Occasionally, shale-roofed discontinuities display numerous, low-relief erosional channels. This phenomenon is well displayed at two levels in Erie County as well as a few other contacts. In these cases, again erosional scours are associated with minor but abrupt deepening events. The first example, involves the junction of a black shale with an underlying gray mudstone unit in the Levanna Shale Member (lower Hamilton Group, Skaneateles Formation) at Buffalo Creek (Stop 4; Fig. 8). The contact is sharp and is marked by a series of small, low-relief channels or gullies each about 1/2 to 1 m across and with a total relief of about 35 cm which resemble submarine furrows (see Flood, 1983). These gullies are subparallel, evenly spaced, and oriented approximately N-S, perpendicular to the inferred paleoslope.

The basal contacts of the channels are delineated by very thin erosion lag deposits of crinoid debris, trilobite fragments and fish bones and teeth. A particularly unusual feature of the channels is that the contact between the gray and overlying dark gray shales is sharp at the rims, but appears slightly blurred in the floors of the gullies. As it is obvious that an erosion surface separates the gray and black shales, the apparent gradation between the two facies in channel bottoms is paradoxical. We suggest that during earliest phases of dark mud deposition in the channels, conditions remained marginally oxic. Burrowing organisms were able to mix dark, slightly organic-rich muds downward into underlying, still soft, gray muds.

The erosional channels themselves may have been produced as bypass chennels on a gently south-sloping ramp; vortical debris-laden currents developed during sustained unidirectional flow may have have initiated a ridge-and-furrow topography; later, currents carrying erosive agents such as shell debris, accentuated the erosional runnels scouring them somewhat deeper. Finally, the furrowed surface was blanketed by dysoxic muds; channel-filling alternated with episodes of minor scour resulting in the phenomenon of channels nested within the larger channels (Fig. 8c).

A second example of furrowing has been described in detail from the Wanakah Shale by K.B. Miller (1988; see discussion in article by Miller, this guidebook). In this instance, a series of broader and deeper channels (with widths up to 30 m and depths up to 1 to 2 m) were cut into a gray fossiliferous mudstone. In Erie County the channels are filled with a slightly darker, blue gray fissile to platy, barren mudstone. Farther east, in the Finger Lakes region, these channels still occur at the same horizon, but they are infilled with a very dark gray to black fissile shale. Miller carefully considered various alternatives and concluded that the Wanakah "channels" were erosional forms rather than lows between mud bars or ripples.

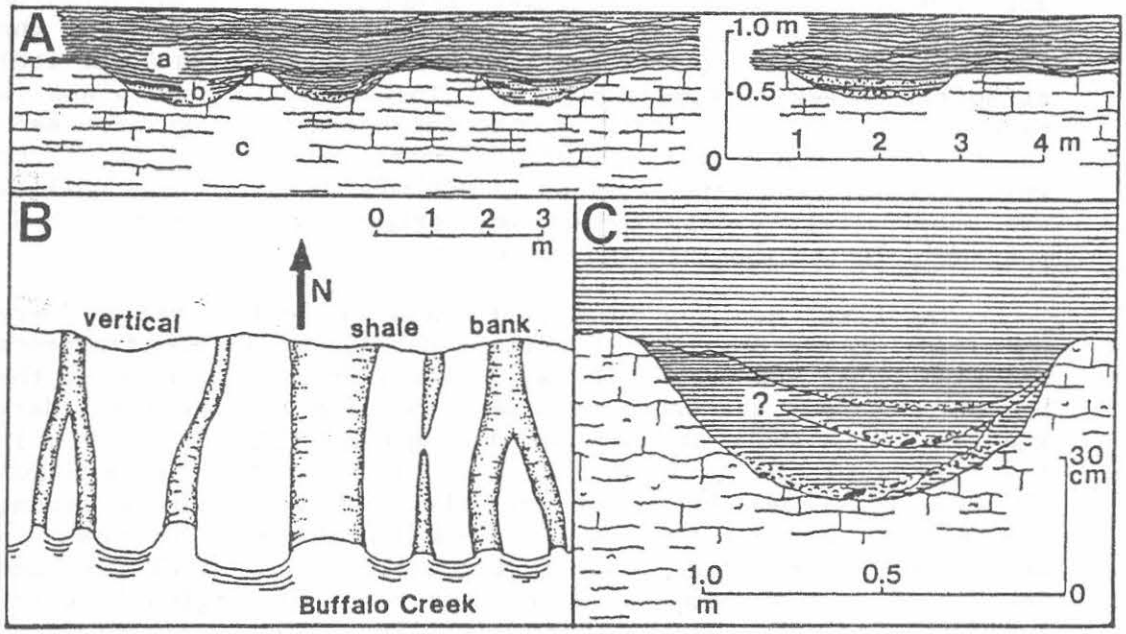


Figure 8. Submarine discontinuity within Levanna Shale Member on Buffalo Creek at Union Road (see STOP 4). A) along-bank profile of a series of erosional runnels (troughs) into calcareous mudstone which are filled with brownish-black shale of upper unit; B) vertical ("map") view of channels on exposed creek bed bordering shale bank. Note southward bifurcation of some channels suggestive of southward current-flow; C) complex history of episodic scouring and filling of mud within channels. Sharp scoured contacts with associated lag debris of fish bones and shells grade laterally to extinction (continuity) over very short distances. Lettered units include: a) black, laminated shale with Leiorhynchus, Styliolina, and palynomorphs; b) brown-black shale filling in troughs with associated scoured contacts and brachiopod-trilobite fish bone lag debris; c) calcareous gray to dark grey blocky to chippy mudstone with Ambocoelia, Devonochonetes, and Phacops rana.

The Wanakah channels are somewhat accentuated by the development of early diagenetic concretions which follow the contours of the channel profiles (the concretion horizons appear to move up and down along the outcrop face as they outline the otherwise cryptic channels). Again, a minor lag of shell debris is present locally along the bases and margins of the channels and furrows. Here, as in the Levanna example, the development of the channeled (furrowed) mud surface appears to have taken place during interval of sediment-starvation during rapid relative sea level rise (see Miller: FIELD TRIP: this volume).

CONDENSED INTERVALS ASSOCIATED WITH BENTONITES

Finally, some metabentonite beds appear to have at their bases thin, condensed bone- lag deposits. This may indicate that rather than representing a catastrophic pulse of ash influx the general interval of increased volcanism and ash may have had an influence on the sedimentary systems in producing a sediment-starved surface, over which ash, presumably from many episodes of volcanism, gradually accumulated. Some metabentonites appear to be a type of condensed deposit analogous to bone beds; indeed many erosion-related bone-lag deposits are enriched in pyroclastic debris (see Conkin and Conkin, 1984; Conkin et al., 1980) which one would expect within "time-dense" marine lag units (see STOP 5).

At present, it is unclear whether the sediment-starvation interval was produced by a minor sea-level rise that coincided with volcanic activity or by a curtailment of carbonate production. The latter would perhaps result from intense episodes of ash input.

CONCLUSIONS

Significant discontinuities occur, predictably, at three positions within Middle and Upper Devonian marine shale and carbonate sequences of western New York. Pronounced erosional unconformities, commonly with irregular, burrowed contacts, occur on the bases of shallow water carbonate beds corresponding to rapid lowering of relative sea level. These contacts (e.g., basal Tichenor, basal Tully) correspond to sequence and subsequence boundaries. In some cases they are composites of lowstand submarine (and possibly subaerial) erosion surfaces and transgressive ravinements.

A second type of important discontinuity occurs at the bases of major highstand deposits, roofed by black shales. These are typically planar disconformities (no burrowing and only minor very low-relief furrowing), marked by concentrations of phosphate, pebbles, glauconite, conodonts, bones and/or reworked pyrite. Such surfaces are associated with condensed sections formed during times of rapid sea level-rise; as such, they correspond to accentuated surfaces of maximum sediment-starvation or downlap surfaces of sequence models. Erosional processes may include deep storm waves, density currents, or interval waves propagated along pycnoclinal water mass boundaries.

The third category of discontinuity and condensed interval occcurs at the boundary between early deepest water highstand deposits and progradational late highstand deposits. As such they appear to be associated with the inflection point between latest phases of sea level rise and early stages of regression. They are overlain by shallowing-upward intervals of gray mudstone to silty calcareous mudstone that may culminate in a sequence or subsequence bounding unconformity below shallow water transgressive deposits. These mid-highstand beds commonly display reworked concretions and relatively shallow water biofacies. They appear to initiate cycles of rapid shallowing and mimic some characteristics of cycle-capping carbonates (limestones above sequence or subsequence boundaries) - hence, we term them "precursor" beds. Such beds are not predicted by most sedimentary models. We infer that they may represent brief sediment-starved intervals during periods of sediment disequilibrium resulting from sea level lowering. Submarine erosion probably involves a combination of bioturbation and storm-generated gradient currents.

Finally, some diastems may be associated with ash beds and may or may not be related to sea level fluctuations. Some ash layers may be condensed beds.

All four types of discontinuities and associated condensed beds provide outstanding regional and roughly isochronous markers that enable detailed physical correlations of strata. They also subdivide the stratigraphic column into bounded genetically related bundles. Mapping of these discontinuities facilitates interpretation of sea level variations and basin dynamics.

ACKNOWLEDGMENTS

We wish to thank various people whose discussions or field work have influenced our ideas on sequence stratigraphy in the Devonian including former and present students Lee Gray, Mike Savarese, Keith Miller, William Goodman, Steve Mayer, and Dave Griffing. We also thank various property owners for making our stops possible.

This manuscript benefited from critical reviews by Curt Teichert and David Lehmann. Above all, we extend our sincere thanks to Heidi Jacob for patiently and skillfully processing our manuscripts.

This research was funded in part by National Science Foundation Grant EAR 8816856; early work in Erie County was supported by NSF EAR 8313103 and a grant from the donors to the Petroleum Research Fund, American Chemical Society.

REFERENCES

Baird, G.C., 1978, Pebbly phosphorites in shale: a key to recognition of a widespread submarine discontinuity in the Middle Devonian of New York. Jour. Sed. Petrol., v. 48, p. 545-555.

Baird, G.C., 1979, Sedimentary relationships of Portland Point and associated middle Devonian rocks in central and western New York. N.Y. State Mus. Bull., v. 433, p. 1-23.

Baird, G.C., 1981, Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian. Lethaia, v. 14, p. 105-122.

Baird, G.C. and Brett, C.E., 1981, Submarine discontinuities and sedimentary condensation in the upper Hamilton Group: examination of paleoslope deposites in the Cayuga Valley. N.Y. State Geol. Assoc. 53rd Ann. Mtg. Guidebook; Binghamton, N.Y., p. A1-A33.

Baird, G.C. and Brett, C.E., 1983, Regional variation and paleontology of two coral beds in the Middle Devonian Hamilton Group of western New York. Jour. Paleontol., v. 57, p. 417-446.

Baird, G.C. and Brett, C.E., 1986a, Erosion on an anaerobic seafloor: significance of reworked pyrite deposits from the Devonian of New York State. Palaeogeog., Palaeoclim., Palaeoecol., v. 57, p. 157-193.

Baird, G.C. and Brett, C.E., 1986b, Submarine erosion on the dysaerobic seafloor: Middle Devonian corrasional disconformities in the Cayuga Valley region. N.Y. State Geol. Assoc. 58th Ann. Meeting: Guidebook, Ithaca, N.Y., p. 23-80.

Baird, G.C., Brett, C.E., and Kirchgasser, W.T., 1988, Genesis of black-shale roofed discontinuities in the Devonian Genesee Formation, western New York, p. 357-375. In McMillan, N.J., Embry, A.I. and Glass, D.J., eds., Devonian of the World, v. II. Can. Soc. Petrol. Geol. Mem. 14, v. II.

Baird, G.C., Brett, C.E. and Frey, R.C., 1989, "Hitchiking" epizoans on orthoconic cephalopods: preliminary review of the evidence and its implications. Senckenbergiana lethaea, v. 69, p. 439-465.

Baum, G.R. and Vail, P.R., 1988, Sequence stratigraphic concepts applied to Paleogene outcrops, Gulf and Atlantic Basins. Soc. Econ. Pal. Min. Spec. Pub., v. 42, p. 309-328.

Brett, C.E., 1974, Contacts of the Windom Member (Moscow Formation) in western Erie County, New York. N.Y. State Geol. Assoc. 46th Ann. Meeting Guidebook, Fredonia, N.Y., p. C1-C22.

Brett, C.E. and Baird, G.C., 1974, Late Middle and Early upper Devonian disconformities and paleoecology of the Moscow Formation in wester Erie County. In N.Y. State Geol. Assoc. 46th Ann. Meeting Guidebook, Fredonia, N.Y., p. C23-C29.

Brett, C.E. and Baird, G.C., 1982, Upper Moscow-Genesee stratigraphic relationships in western New York: evidence for regional erosive beveling in the Late Middle Devonian. N.Y. State Geol. Assoc. 54th Ann. Meeting Guidebook, Buffalo, N.Y., p. 217-245.

Brett, C.E. and Baird, G.C., 1985, Carbonate shale cycles in the middle Devonian of New York: An evaluation of models for the origin of limestones in terrigenous shelf sequences. Geology, v. 13, p. 324-327.

Brett, C.E. and Baird G.C., 1986, Symmetrical and upward shallowing cycles in the Middle Devonian of New York State and their implications for the punctuated aggradational cycle hypothesis. Paleoceanography v. 1, no. 4, p. 431-445.

Brett, C.E., Speyer, S.E. and Baird, G.C., 1986, Storm-generated sedimentary units: tempestite proximatelity and event stratification in the Middle Devonian Hamilton Group of New York, p. 129-156. In Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Pt. 1, N.Y. State Mus. Bull. 457, 156 p.

Conkin, J.E. and Conkin, B.M. 1984, Paleozoic metabentonites of North America: Part 1: Devonian Metabentonites in the eastern United States and southern Ontario: their identities, stratigraphic positions, and correlatioon. Univ. Louisville Studies in Paleontol. and Stratig., v. 16, 135 p.

Conkin, J.E., Conkin, B.M., and Lipchinsky, L.Z., 1980, Devonian black shale in the eastern United States; Part 1. Southern Indiana, Kentucky, northern and eastern highland rim of Tennessee and central Ohio. Univ. Louisville Studies in Paleont. and Stratig., v. 12, 63 p.

Dennison, J.M., 1983, Internal stratigraphy of Devonian Tioga ash beds in Appalachian Valley and Ridge Province. Geol. Soc. America, Abstracts with Programs, v. 15, no. 6, p. 557.

Dennison, J.M. and Head, J.W., 1975, Sea level variations interpreted from the Appalachian Basin Silurian and Devonian. Amer. Jour. Sci., v. 275, p. 1089-1120.

Dennison, J.M. and Textoris, D.A., 1978, Tioga metabentonite time-marker associated with Devonian shales in Appalachian Basin, in Schott, G.L., Overby, W.K., Jr., Hunt, A.E., and Komar, C.A., eds., Eastern Gas Shale Symposium, 1st Proceedings: U.S. Dept. Energy Spec. Pap., MERC/SP-77/5, p. 166-182.

deWitt, W. and Colton, G.W., 1978, Physical stratigraphy of the Genesee Formation (Devonian) in western and central New York. U.S. Geol. Surv. Prof. Pap., v. 1032-A, 22 p.

Ettensohn, F.R., 1985, Controls on the development of Catskill Delta complex basin facies, p. 63-77. In Woodrow, D.L. and Sevon, W.D., eds., The Catskill Delta. Geol. Soc. Amer. Spec. Pap. 201.

Ettensohn, F.R., 1987, Rates of relative plate motion during the Acadian orogeny based on the spatial distribution of black shales. Jour. Geol., v. 95, p. 572-582.

Ettensohn, F.R., Miller, M.L., Dillman, S.B., Elam, T.D., Geller, K.L., Swager, D.R., Markowitz, G., Woock, R.D. and Barron, L.S., 1988, Characterization and implications of the Devonian-Mississippian black-shale sequence, eastern and central Kentucky, U.S.A.: pycnoclines, transgression, regression and tectonism, p. 323-346. In McMillan, N.J., Embry, A.F. and Glass, D.L., eds., Devonian of the World, Canadian Soc. Petrol. Geol. Mem. 14., v. II.

Flood, R.D., 1983, Classification of sedimentary furrows and a model for furrow initiation and evolution: Geol. Soc. Amer. Bull., v. 94, p. 630-639.

Grasso, T.X., 1986, Redefinition, stratigraphy and depositional environments of the Mottville Member (Hamilton Group) in central and eastern New York: In Brett, C.E., ed., Dynamic Stratigraphy and depositional environments of the Hamilton Group (Middle Devonian) in New York State, Part I: N.Y. State Museum Bull., v. 457, p. 5-31.

Gray, L.M., 1984, Lithofacies, biofacies and depositional history of the Centerfield Member (Middle Devonian) of western and central New York State. Unpubl. Ph.D. diss., Univ. of Rochester, 158 p.

Gray, L.M., in press, The paleoecology, origin and significance of a regional disconformity at the base of the Ludlowville Formation (Middle Devonian) in central New York. In Landing, E. and Brett, C.E., eds., Dynamic Stratigraphy and depositional environments of the Hamilton Group (Middle Devonian) of New York State. N.Y. State Mus. Bull.

Heckel, P.H., 1973, Nature, origin, and significance of the Tully Limestone. Geol. Soc. Am. Spec. Pap. 138, 244 p.

Huddle, J.W., 1981, Conodonts from the Genesee Formation in western New York. U.S. Geol. Survey Prof. Pap., p. 1032-A.

Hussakoff, L. and Bryant, W.L., 1918, Catalog of the fossil fishes in the museum of the Buffalo Society of Natural Sciences, Buffalo Soc. Natural Sci. Bull., v. 12, p. 1-198.

Johnson, J.G., 1970, Taghanic onlap and the end of North American Devonian provinciality. Geol. Soc. America Bull., v. 81, p. 2077-2106.

Johnson, K.G. and Friedman, G.M., 1969, The Tully clastic correlatives (Upper Devonian) of New York State: a model for recognition of alluvial dune?, tidal, nearshore (bar and lagoon) and offshore sedimentary environments in a tectonic delta complex: Jour. Sed. Petrol., v. 39, p. 451-485.

Johnson, J.G., Klapper, G. and Sandberg, C.A., 1985, Devonian eustatic fluctuations in Euramerica. Geol. Soc. Amer. Bull., v. 96, p. 567-587.

Kent, D.V., 1985, Paleocontinental setting for the Catskill Delta. In Woodrow, D.L. and Sevon, W.D., eds., The Catskill Delta. Geol. Soc. Amer. Spec. Pap. 201, p. 9-13.

Kirchgasser, W.T. and House, M.R., 1981, Upper Devonian goniatite biostratigraphy. In Oliver, W.A., Jr. and Klapper, G., eds., Devonian biostratigraphy of New York. I.U.G.S., Subcomm. Devonian Stratigr., Washington, D.C., v. 1, p. 39-55.

Kirchgasser, W.T., Baird, G.C. and Brett, C.E., 1988, Regional placement of the Middle/Upper Devonian (Givetian-Frasnian) boundary in western New York State, p. 113-118. In McMillan, N.J., Embry, A.F. and Glass, D.J., eds., Devonian of the World, Canadian Soc. Petrol. Geol. Mem. 14, v. II.

Landing, E. and Brett, C.E., 1987, Trace fossils and regional significance of a Middle Devonian (Givetian) disconformity in southwestern Ontario: Jour. Paleontol. v. 61, p. 205-230.

Loutit, T.S., Hardenbol, J., Vail, P.R. and Baum, G.R., 1988, Condensed 66sections: the key to age determination and correlation of continental margin sequences. SEPM Spec. Pub. 42, p. 183-213.

McCave, I.N., 1973, The sedimentology of a transgression: Portland Point and Cooksburg Members (Middle Devonian): New York State. Jour. Sed. Petrol., v. 43, p. 484-504.

Mayer, S.M., 1989, Stratigraphy and paleontology of the Jaycox Shale Member, Hamilton Group of the Finger Lakes region of New York State [unpubl. masters thesis]. SUNY College at Fredonia, 121 p.

Mayer, S.E., Brett, C.E. and Baird, G.C., 1990, New correlations of the Upper Ludlowville Formation, Middle Devonian: implications for upward-coarsening regressive cycles in New York. Abstr., Geol. Soc. America v. 22, no. 22, p. 54.

Meyer, W.F., 1985, Paleodepositional environments of the Stafford Limestone (Middle Devonian) across New York State [unpubl. masters thesis]. SUNY College at Fredonia, 67 p.

Miller, K.B., 1986, Depositional environments and sequences, "Pleurodictyum Zone", Ludlowville Formation of western New York, p. 57-77. In Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) of New York State. N.Y. State Mus. Bull., v. 457, 157 p.

Miller, K.B., 1988, A temporal hierarchy of paleoecologic and depositional processes across a middle Devonian epeiric sea. Unpubl. Ph.D. diss., Univ. of Rochester, 243 p.

Posamentier, H.W., Jervey, M.T., and Vail, P.R., 1988, Eustatic controls on clastic deposition-conceptual framework. Soc. Econ. Pal. Min. Spec. Pub., v. 42, p. 109-124.

Rickard, L.V., 1975, Correlation of the Silurian and Devonian rocks in New York State. N.Y. State Mus. and Sci. Serv. Map and Chart Series 24, p. 1-16.

Rickard, L.V., 1981, The Devonian system of New York State. In Oliver, W.A., Jr. and Klapper, G., eds., Devonian Biostratigraphy of New York, Pt. I, Intl. Union of Geol. Sci., Subcom. of Devonian Stratigraphy, Washington, D.C., p. 5-22.

Rickard, L.V., 1984, Correlation of the subsurface Lower and Middle Devonian of the Lake Erie region. Geol. Soc. America Bull., v. 95, p. 814-828.

Sass, D.B., 1951, Paleoecology and stratigraphy of the Genundewa Limestone of western New York [unpubl. masters thesis]. Rochester, University of Rochester, 113 p.

Savrda, C.E. and Bottjer, D.J., 1987, The exaerobic zone, a new oxygen-deficient marine biofacies. Nature, v. 327, p. 54-56.

Sparling, D.R., 1988, Middle Devonian stratigraphy and conodont biostratigraphy, north-central Ohio. Ohio Jour. Sci., v. 88, p. 2-18.

Sloss, L.R., 1963, Sequences in the cratonic interior of North America. Geol. Soc. Amer. Bull., v. 74, p. 93-113.

Tucker, M.E., 1973, Sedimentary and diagenesis of Devonian pelagic limestones (Cephalopodenkalk) and associated sediments of the Rhenohercynian Geosyncline, West Germany, Neues Jahrb. Geologie Paläont., Abh., v. 142, p. 320-350.

Vail, P.R., Mitchum, R.M., 1977,Seismic stratigraphy and global changes of sea-level, part I: overview, <u>In</u> C.E. Payton, ed., Seismic stratigraphy -- applications to hydrocarbon exploration. Am. Assoc. Petrol. Geol. Memoir 26, p. 51-52.

Vail, P.R., Mitchum, R.M., Jr., Todd, R.G., Widmier, J.M., Thompson, S., III, Sangree, J.B., Bubb, J.N. and Hatlelid, W.G., 1977, Seismic stratigraphy and global changes of sea level. Am. Assoc. Petrol. Geol. Mem. 26, p. 49-212.

Van der Voo, R., 1988, Paleozoic paleogeography of North America, Gondwana and intervening displaced terranes: comparison of paleomagnetism with paleoclimatology and biogeographical patterns. Geol. Soc. Amer. Bull. v. 100, p. 31-324.

Van Wagoner, J.C. Posamentier, H.W., Mitchum, R.M., Vail, P.R., Sarg, J.F., Loutit, T.S., and Hardenbol., J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions, SEPM Spec. Pub., v. 42, p. 39-46.

Vogel, K., Golubic, S. and Brett, C.E., 1986, Endolith associations and their relation to facies distribution in the Middle Devonian of New York State, U.S.A. Lethaia, v. 20, p. 263-290.

Wanless, H.R., 1986, Production of subtidal tubular and surficial tempestites by Hurricane Kate, Caicos Platform, British West Indies. Jour. Sed. Petrol., v. 58, p. 730-750.

Woodrow, D.L., 1985, Paleogeography, paleoclimate and sedimentary processes of the Late Devonian of New York State, U.S.A. Lethaia, v. 20, p. 263-290.

Woodrow, D.L., Dennison, J.M., Ettensohn, F.R., Sevon, W.T., and Kirchgasser, W.T., 1988, Middle and Upper Devonian stratigraphy and paleogeography of the central and southern Appalachians and eastern mid-continent, U.S.A. In McMillan, N.J., Embry, A.J. and Glass, D.J., eds., Devonian of the World, v. I, Can. Soc. Petrol. Geol. Mem. 14, v. I.

Yochelson, E.L. and Lindeman, R.H., 1986, Considerations on systematic placement of styliolines (incertae sedis., Devonian). In Hoffman, H. and Nitecki, M.H., eds., Problematic Fossil Taxa, Oxford Univ. Press., p. 48-56.

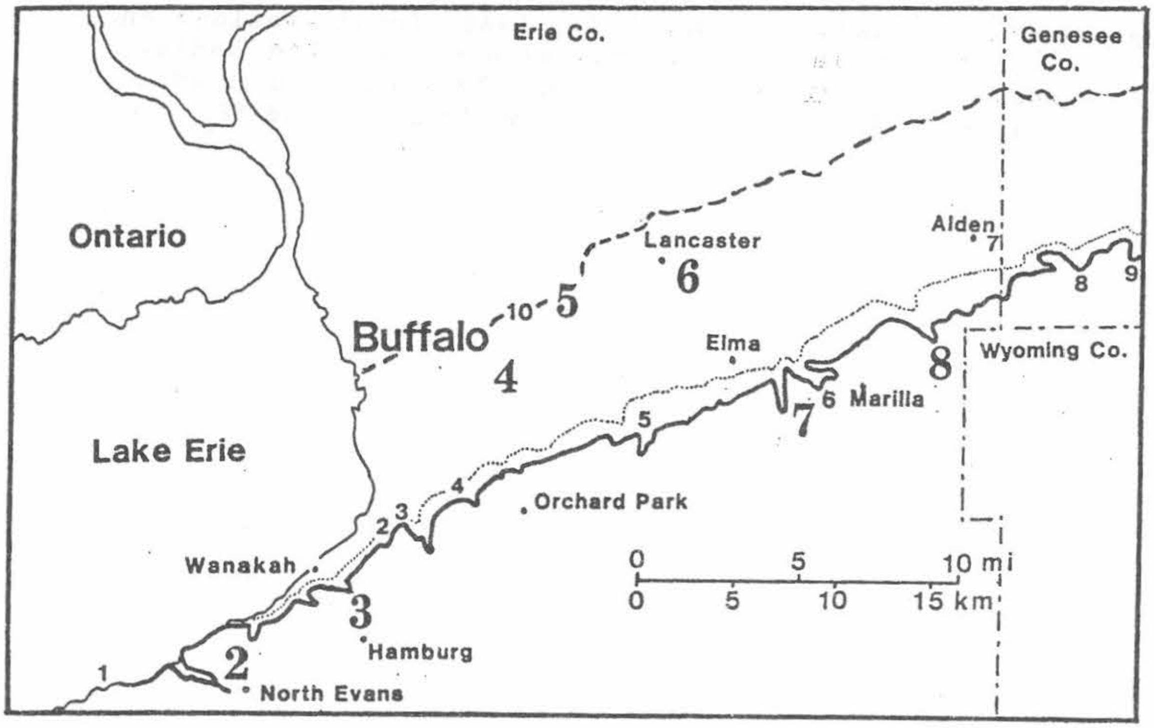


Figure 9. Field trip stops in Erie County, New York. (STOP 1: Point Gratiot at Dunkirk, Chautauqua County, not shown). Outcrop belt of the Middle Devonian Hamilton Group and localities examined. Dashed line denotes contact of Onondaga Limestone and Marcellus Formation; stippled line is base of Moscow Formation; heavy dark line is base of Genesee Formation. Key reference sections include: 1) Lake Erie bluffs at Highland-On-The-Lake; 2) unnamed creek south of Big Tree Road; 3) Bayview Shale Pit (formerly Penn Dixie Quarry); 4) Smoke Creek south branch near Blasdell; 5) Cazenovia Creek at Elma; 6) Little Buffalo Creek at Marilla; 7) Spring Creek at Alden; 8) Elevenmile Creek near Darien Center; 9) Murder Creek at Darien; 10 Federal Crushed Stone Corp. Quarry in Cheektowaga. Field trip stops (large numbers) include: Eighteenmile Creek (2); abandoned shale pit at Cloverbank (3); Buffalo Creek at Union Road (4); Cayuga Creek in Depew (5); Cayuga Creek at Como Park (6); Buffalo Creek at Bullis Road (7); Cayuga Creek at Clinton Road (8).

ROAD LOG FOR DEVONIAN DISCONTINUITIES

(see Figure 9)

TOTAL MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0		Leave Fredonia Campus at the Temple Street exit. Turn right (N) onto Temple St.
0.4	0.4	Y-intersection; bear right (N) on Brigham Road.
0.8	0.4	Cross New York State Thruway
1.5	0.7	Enter City of Dunkirk
1.6	0.1	Pass Al-Tech Corporation Factory (on right).
2.7	1.1	Junction of Brigham Road with Lake Shore Drive (Route 5) turn left (SW) onto Lake Shore Drive.
2.9	0.2	Junction of Lake Shore Drive (Route 5) with North Point Drive. Turn right (N) onto North Point Drive.
3.1	0.4	Turn right onto Cedar Beach Parking area. Niagara Mohawk coal-powered generating station straight ahead (to NE). Proceed north on foot for approximately 500 feet along beach to northeast-facing lake shore outcrop near lighthouse museum at Point Gratiot.

STOP 1. TYPE SECTION OF DUNKIRK SHALE MEMBER, POINT GRATIOT AT DUNKIRK.

This outcrop is excellent for viewing the sharp contrast between the grey-green bioturbated, but non-shelly, mudstone of the Hanover Shale Member and the black, organic-rich, laminated facies of the overlying Dunkirk Member. A conformable, bioturbated contact between Hanover and Dunkirk can be seen below the 15 to 18 cm thick (6 to 7 inches) basal Dunkirk black bed which dips southward below the water at this section. However, a sharp, erosional discontinuity with associated reworked debris marks the contact between the 12-13 cm (5 inch) thick grey-green bioturbated bed above the 18 cm (7 inch) basal bed and the overlying continuous black shale succession.

Close examination of this contact shows that abundant reworked (detrital) pyrite is present along it with lesser amounts of carbonized wood, conodonts, fish bones, and pyritic goniatite steinkerns (Baird and Brett, 1986a); much of this material is probably derived from the 12-13 cm (5 inches), grey-green mudstone unit which is conspicuously rich in pyrite cemented and overgrown burrow tubes, although the abundance of wood suggests that the basal few inches of the overlying black shale is unusually condensed, recording the descent of numerous logs from the water surface over a long period of time. This lag unit also records bioturbation, which at this section, was sufficient to flux the broken pyritic tubes into random orientations.

As this discontinuity is traced northeastward to the vicinity of Java Village, Wyoming County, more and more beds progressively appear beneath it such that this break passes essentially to extinction (see Fig. 5; Baird and Lash; FIELD TRIP SAT. A, this volume). Instead several black shale beds alternating with bioturbated and non-bioturbated grey-green mudstone make up an intervening sequence several meters-thick. Most of the thin black shales, however, display minor diastemic basal contacts with the grey-green beds marked by minor detrital pyrite. Collectively, these diastems downcut underlying units as they are traced southwestward such that higher discontinuities downcut through lower ones until only a single 5 inch (12-13 cm) grey-green bed remains at this locality. This illustrates the pervasiveness of submarine erosion associated with the transgressive grey-to-black facies change beneath even very thin black shale units. It also illustrates the distinctive character of condensed sedimentary deposits involving alternations between dysoxic and anoxic facies. For additional information about this outcrop, see Baird and Lash; FIELD TRIP LOG SAT A: STOP 2 (this volume).

3.5	0.4	Reboard vehicles and return to Lake Shore Drive (Route 5). Turn left (NE) onto Lake Shore Drive. Dunkirk Harbor Complex visible (to left) for next mile.
4.6	1.1	Junction of Lake Shore Drive with Main Street (Route 60). Turn right (5) onto Main Street.
5.2	0.6	Junction (Y-intersection bifurcation) of Main Street and Route 60. Bear left onto Route 60.
5.6	0.4	Cross Norfolk and Western Railroad tracks.

5.7	0.1	Leave Dunkirk, New York
6.7	1.0	Thruway overpass
7.0	0.3	Thruway entrance at red light. Turn left to enter Thruway. Bear right to go towards Buffalo.
15.1	8.1	Cross Walnut Creek. Hanover-Dunkirk contact is at base of bridge. Stebbins Road bridge (being repaired to right) collapsed in 1987, killing one person, when an overweight vehicle caused the bridge to plunge.
16.2	1.1	Cross Silver Creek, Dunkirk Shale at base of bridge.
19.4	3.2	Dunkirk Shale above Hanover Shale in Thruway cut immediately beyond (downhill from) the Silver Creek exit.
19.8	0.4	Cross Cattaraugus Creek floodplain. Sand and gravel pits have been opened on this bottom in the vicinity of the Thruway.
20.5	0.7	Cross Cattaraugus Creek. Begin to cross Seneca Indian Reservation.
27.4	6.8	Cross Big Sister Creek. Angola Shale Member of West Falls Formation exposed below bridge.
27.7	0.3	Exit for Thruway rest area and restaurant facilities. Keep going straight.
29.7	2.0	Turn off Thruway at Eden exit.
30.4	0.7	Thruway exit junction with Eden-Angola Road. Turn right (w) towards Angola.

32.4	2.0	Junction of Eden-Angola Road with U.S. Route 20. Continue straight on Eden-Angola Road.
34.7	2.3	Turn right onto Old Evans Center Road.
35.1	0.4	Junction of Old Evans Center Road with North Main Rd. Bear right onto North Main Rd.
35.3	0.2	Junction of North Main Road with Route 5. Turn right (NE) onto Route 5.
		Enter village of Highland-On-The-Lake.
38.7	3.4	Cross Eighteenmile Creek.
38.8	0.1	Turn right onto first side street north of Route 5 bridge. Park vehicles and proceed down path at end of street to large cutbank section on north side of Eighteenmile Creek 200 ft. east of (upstream from) Route 5 bridge.

STOP 2. MIDDLE AND UPPER DEVONIAN MARINE DEPOSITS ALONG EIGHTEENMILE CREEK.

This large outcrop shows to maximum advantage the Tichenor Limestone, the overlying Windom Shale Member, and the major discontinuity at the top of the Windom (marked by North Evans bone-conodont-rich debris) which marks the base of the Upper Devonian series in this area (Figs. 3, 4A). We will see the North Evans to better advantage at the next stop (see STOP 3) and will, thus, focus on the Tichenor and basal Windom shale at this stop.

The Tichenor is a 0.7 m (2 ft.)-thick, widespread crinoid-brachiopod grainstone-packstone unit which is also noteable for large corals (Heliophyllum, Eridophyllum, Favosites), large bivalves (Plethomytilus, Actinopteria) and large fragments of the trilobite Phacops rana. As is obvious from the texture, this is a medium to high energy deposit which may have accumulated within the depth-range of fairweather wavebase. It is also extremely condensed and time-rich; in the western Finger Lakes region, deposits equivalent to the western Tichenor (mostly Deep Run Shale Member) reach a thickness of almost thirty five meters. Through a combined affect of westward stratigraphic thinning and westward lateral facies change from mudstone to shelly carbonate, the Deep

Run is replaced by very thin clean skeletal carbonate in Erie County. The base and top of the Tichenor at this locality are sharp, reflecting erosional processes. The base marks a major disconformity seperating deposits of the Ludlowville Formation from the overlying Moscow Formation; the Jaycox Shale Member of the Ludlowville Formation is progressively bevelled below this contact from the Genesee Valley westward such that only the base of the Jaycox is visible at Buffalo Creek in eastern Erie County and none is visible west of the north branch of Smoke Creek in south Buffalo. This bevelling was produced by a major regression event which caused synchronous submarine (and possibly subaerial) erosion in central and eastern New York State (see McCave, 1973; Baird, 1979; Baird and Brett, 1981) and this contact may be a bona fide sequence boundary. Examine overturned Tichenor blocks for large, abrasion-enlarged, hypichnial trace fossil markings produced by large arthropods; these resting-(or dwelling) excavations into firm Wanakah marine muds were later filled-and casted by the Tichenor skeletal sands.

The top of the Tichenor is also marked by a submarine discontinuity, perhaps of similar hiatal magnitude. Submarine erosion that preceded deposition of the Windom, removed most of (or all of) three units (Deep Run-equivalent carbonate beds, Menteth Member, Kashong Member) which overlie the Tichenor east of Buffalo. This erosion produced a submarine hardground at this locality and at other western Erie County Tichenor sections). Examine this surface closely for auloporid corals and crinoid holdfasts which encrust this contact and for the presence of phosphatic nodules and steinkerns reworked from below (see Brett, 1974; Brett and Baird, 1974 for additional description of this surface).

The overlying Windom Member records transgressive deepening to quiet, open-shelf conditions minimally affected by storm wave-reach. Fifty centimeters (16 in.) above the base in a unit (Bayview Coral Bed) which is rich in large brachiopods, particularly Spinatrypa, and which locally yields large rugose corals, including Cystiphylloides and Heliophyllum along this bank. Above it is a calcareous mudstone unit (Smoke Creek Bed) which is well known for complete enrolled and outstretched specimens of the trilobite Phacops rana (see Brett and Baird, 1982 for additional information on this unit).

		Return to vehicles. Turn right onto Route 5 and proceed northeast towards Wanakah, N.Y.
41.2	2.4	Cross Weyer Creek. Fossiliferous Middle Devonian Wanakah Shale exposed mainly downstream (north) of Route 5 bridge.

41.5	0.3	Enter town of Wanakah, New York
41.6	0.1	Junction of Route 5 and Old Lake Road. Proceed straight (NE) on Route 5.
42.6	1.0	Cross Amsdell Road Amsdell Creek, further up Amsdell Road, has been an important source for conodonts and fish fossils from the North Evan bone-conodont bed.
42.8	0.2	Enter town of Cloverbank.
43.1	0.3	Turn right (SE) onto Cloverbank Road by the Cloverbank Hotel.
43.4	0.3	Cross Amtrak railroad tracks
43.5	0.1	Park on shoulder pull-off just south of the railroad tracks. Proceed on foot 0.3 mile southwest along a path bordering tracks to the abandoned Lehigh Cement Co. Shale quarry which is southeast of the tracks.

STOP 3. NORTH EVANS BONE BED AND GENUNDEWA LIMESTONE AT THE CLOVERBANK QUARRY.

This abandoned shale pit, developed by the Lehigh Cement Co. as a source of claystone for the mill (now defunct) in Lackawanna, contains a stratigraphic succession from the Middle Devonian in the Upper Devonian. The Quarry walls about 16 m (50ft) high expose the Upper Devonian Middlesex-Cashaqua Shale. The quarry is floored mainly by the shaley upper contact of the Genundewa limestone and the West River Shale. In a narrow pit the Genundewa is breached and a section extends down some 2.5-3 m (8-10ft.) to, or just below, the Amsdell (Praeumbona) bed of the Windom shale. The Windom here is in contact with the North Evans Limestone and the contact is an erosion surface showing a characteristic "rip-up" horizon. We will focus on lag deposits (North Evans Bone-Conodont Bed) associated with an erosional contact between the grey Windom Member (Givetian) and the overlying Genundewa Limestone of Late Devonian (early Frasnian) age. For additional information on these units, see: Sass, 1951; Huddle, 1981; Brett, 1974; Brett and Baird, 1982; Baird and Brett, 1986a.

The North Evans Limestone, or "Conodont bed" of older literature, is typically a 0.5 to 6 inch-thick packstone-grainstone unit composed of reworked crinoid ossicles, conodonts (including many mixed from older zones), fish teeth and armor fragments, as well as calcitic brachiopod, coral, and trilobite fossils derived from the underlying Windom or from intervening beds now removed. Conspicuous within the North Evans at this locality and also at the abandoned Penn Dixie Shale Pit at Bayview, New York (see Brett and Baird, 1982: STOP 2) are reworked limestone concretions from the underlying Windom; These are often stacked like shingles and they typically have a green exterior patina of glauconite as well as surface dissolution pits produced following exhumation on the seafloor.

The overlying Genundewa Limestone is composed almost entirely of Styliolina fissurella, a problematic millimeter-long conoidal calcitic shell which may be of tentaculitid affinities or the shell of an unknown protistan organism (Yochelson and Lindemann, 1986). Wood debris, goniatitid and orthocerid cephalopods, and the bivalves Buchiola and Pterochaenia comprise the remainder of a low diversity biota. The Genundewa appears to be a classic example of the pelagic Cephalopodenkalk facies of the European and North African literature (Tucker, 1973). As such, it records very slow sedimentation in an offshore, sediment-starved, dysoxic environment.

Although the North Evans rests on fossiliferous Middle Devonian deposits at this locality and would appear to record a simple transgression over the oxic, shell-rich Windom facies, a more complex story emerges when the same bed is viewed in eastern Erie County (see STOP 7); at that section, the North Evans overlies brown-black anoxic facies of the Upper Devonian Penn Yan Shale Member. Hence, an apparent regression with attendant seafloor erosion is recorded after deposition of the Penn Yan and prior to final accumulation of the North Evans lag. The North Evans-Genundewa succession is, thus, a transgressive interval culminating in the deposition of brown-black anoxic deposits of the overlying West River Shale Member which is visible here above the Genundewa Limestone. The North Evans and Genundewa beds probably correspond ,respectively, to the transgressive surface and overlying condensed interval in the sequence stratigraphy model (see text); they probably correspond, in part, to other sediment-starved condensed units, such as pebbly greensands and bone beds identified by sequence workers in younger sedimentary systems (see Loutit et al. 1988; Baum and Vail, 1988).

The North Evans lag deposit was apparently reworked into the dysoxic Genundewa outer shelf-slope setting because the encrinitic-bone-conodont hash layer grades upward into Genundewa facies and it underlies the Genundewa throughout Erie County except for exposures along Lake Erie near Highland-On-The-Lake where a brown-black shale subfacies of Genundewa overlies the North Evans (Brett and Baird, 1982). Hence the North Evans is rich in carbonate debris with only minor amounts of detrital pyrite; compare the North Evans with the pyrite-dominated lag that we saw at STOP 1 and, particularly, what we will see in the

Leicester Pyrite deposit at STOP 8. Exhumed Windom calcareous shell material was apparently dissolved in the largely anoxic Geneseo basin setting to produce the residual detrital pyrite lag deposit of the Leicester (Baird and Brett, 1986a). Conversely, detrital pyrite was differentially stable in this anoxic or minimally dysoxic setting. Carbonate undersaturation and/or low pH conditions of the basal Geneseo substrate environment are clearly not evident for the North Evans-Genundewa setting to anywhere the same degree. Only where a black shale subfacies of Genundewa directly overlies North Evans along the Lake Erie shore, does the North Evans lag deposit begin to become rich in detrital pyrite, poor in carbonate debris, and laterally discontinuous (lenticular) in outcrop, the way the Leicester always appears.

Fossils to look for in the North Evans are crushing teeth (tritors) of ptyctodonts, an extinct fish order, placoderm dermal armor and jaw fragments, tricuspate cladodid shark teeth, as well as reworked Windom calcareous fossils, including the trilobite Phacops rana.

43.9	0.4	Return to vehicles. Return to Route 5 and turn right (NE) onto Route 5.
44.8	0.9	Cross Morse Creek. One of many shore area localities in the richly - fossiliferous Middle Devonian Wanakah Shale Member.
45.3	0.5	Excellent view of Lake Erie, Buffalo-Lackawanna skyline for next 0.6 mile.
45.7	0.4	Enter town of Athol Springs.
47.7	2.0	Exit Route 5 onto traffic circle by Ford automotive plant. Bear right to get onto Route 179.
47.8	0.1	Turn right onto Route 179.
48.8	1.0	Cross Route 62 (South Park Ave.). Keep going straight.
50.0	0.2	Cross Mile Strip Road and enter New York State Thruway at Blasdell entrance.
50.5	0.5	Mudstone (probably Wanakah) exposed along Thruway entrance ramp.

51.4	0.9	Cross Smoke Creek, Ledyard Shale Member exposed to right (upstream) from Thruway.
52.4	1.0	Proceed through Thruway Toll plaza.
53.3	0.9	Cross Cazenovia Creek. Middle Devonian black Levanna Shale Member exposed upstream (to right) from Thruway bridge.
53.7	0.4	Exit from Thruway onto Route 400. Proceed east on this expressway to Union Rd. exit.
55.4	1.7	Exit from Route 400.
55.7	0.3	Junction of route 400 exit ramp with Union Road. Turn left (N) onto Union Road.
56.5	0.8	Pull off from Union Road to the right onto a dirt road and small parking area in vacant lot immediately before bridge over Buffalo Creek and immediately after passing junction of Indian Church Road on the left. Proceed north, on foot to creek. Cross creek if water is not high.

STOP 4. SUBMARINE DISCONTINUITY WITHIN MIDDLE DEVONIAN LEVANNA SHALE MEMBER ALONG BUFFALO CREEK.

Along this cutbank exposure one can observe two key lithologic divisions of the Middle Devonian Levanna Shale Member of the Skaneateles Formation which are currently unnammed. Just above water level is a calcareous, dark grey-shale division which yields the diminutive brachiopod _Ambocoelia_ and specimens (often complete) of the trilobite _Phacops_ _rana_. A submarine discontinuity (prominent undulatory outcrop reentrant) separates this lower unit from a fissile, black shale upper division, rich in the brachiopod _Leiorhynchus_ and the aformentioned problemmatic small conical fossil _Styliolina_. This boundary is traceable as far east as Oatka Creek near Pavilion in Genesee County. The units in this section record oxygen-deficient outer shelf-to-basin conditions with the lower division recording dysoxic to minimally oxic conditions and the upper division recording lower dysoxic to near-anoxic conditions ("exaerobic" zone of Savrda and Bottjer, 1987) along the seabed.

This discontinuity is distinctive for its distinctly undulatory appearance; troughs between 0.5 and 2.0 meters (1.5-6.5 ft.) in width and between 12 and 45 cm(5 to 16 in) in depth alternate with intertrough ridges and platforms (Fig. 8A). The troughs are erosional runnels cut into division 1 deposits which are aligned in a nearly north-south direction transverse to the creek channel (Fig. 8B). Some runnels bifurcate but most remain simple and linear. Trough bottom deposits often include calcareous brachiopods, Phacops, and Styliolina debris admixed with fish teeth and dermal plates. These lags are often at channel bottoms but they can occur in axial channel sediments above channel bottoms. Some troughs appear to have been repeatedly filled with sediment and scoured out by currents; these troughs display nested erosional scour surfaces with the sharpness of scour contacts varying from clear to diffuse (Fig. 8C). Evidently some episodes of scour removed only water-rich surface mud while others cut into firm muds.

Clearly, this section records a type of sedimentary condensation where repeated sediment accumulation and scour were dominant sedimentary processes. The overall upward-change from division 1 to division 2 appears to be transgressive with the consequent development of an erosional surface; the complex channel-fills appear to correspond to the interval of maximum sediment-starvation and sedimentary condensation which overlies the transgressive erosion surface.

These erosional runnels are probably submarine furrows (sensu Flood, 1983), which are rarely reported from the stratigraphic record. Furrows are believed to form through the action of abrasive horizontal, debris-laden current vortices which scour the bottom into linear runnels within a sustained unidirectional current regime (Flood, 1983). The complex "cut-and-fill" histories of the Levanna runnels is a testament to the unidirectional character of the currents which produced them. We are currently studying these features at this locality to establish which way the currents flowed and are also examining all other similar discontinuities to determine if similar runnels are distributed along them.

		Return to vehicles. Turn right (N) onto Union Rd. and proceed toward towns of Gardenville and Cheektowaga.
56.6	0.1	Cross Buffalo Creek and enter Gardenville, New York.
57.3	0.7	Enter town of Cheektawaga.

58.6	1.3	Cross Cayuga Creek. Middle Devonian Onondaga Limestone is exposed below bridge and upstream (to east).
59.8	1.2	Junction of Union Road and Broadway (Route 130). Exit Union Road to turn right onto Broadway.
59.9	0.1	Junction of turn lane with Broadway. Turn right (E) onto Broadway.
61.2	1.3	Enter town of Depew.
62.3	1.1	Junction of Broadway (Route 130) with Rowley Road. Turn right (S) onto Rowley Road.
62.5	0.2	Cross Cayuga Creek.
62.6	0.1	Junction of Rowley Road and Borden Road. Bear right and stay on Rowley Road.
62.8	0.2	Exposure of Onondaga Limestone along Cayuga Creek (to right). Park on asphalt driveway next to first house on the right. Proceed carefully down slope to south side of creek.

STOP 5. VOLCANIC ASH LAYERS IN UPPER ONONDAGA LIMESTONE ALONG CAYUGA CREEK.

This cutbank exposes two thin metabontonite layers within the upper part of the Onondaga Limestone Formation (Middle Devonian: Eifelian) which comprise part of the "Tioga Ash" (metabentonite) complex of beds discussed in Devonian literature (see Dennison and Textoris, 1978; Conkin and Conkin, 1984; Rickard, 1984; Sparling, 1988). Dennison and Textoris (1978) and Dennison (1983) believe that the tuff originated in what is now northern Virginia and that Devonian prevailing winds carried the winds predominently westward (in the present sense) into the U.S. midwestern region.

The Cayuga Creek section exposes two metabentonite (K-rich mixed-layer, illite-smectite) beds which are about 16 inches apart with the lower one just below the water level. The higher metabentonite is probably the Onondaga Indian Nation ("O.I.N.") metabentonite of Conkin and Conkin (1984) which corresponds to the "Tioga B" metabentonite of Rickard (1984). The lower metabentonite is probably the Cheektowaga (No. 2) metabentonite of Conkin and Conkin (1984); both of these ashes are exposed in the nearby Federal Crushed Stone Corp. Quarry (see Conkin and Conkin, 1984). Although most workers place the boundary between the Moorehouse Member and the overlying Seneca Member of the Onondaga at the horizon of the O.I.N. (Tioga B) ash (see Rickard, 1975, 1984), Conkin and Conkin (1984) place it about 0.5 m higher at the position of a bone bed which is visible in the quarry and which may be present in this section. Uncertainty exists with surface correlations because outcrops in the upper Onondaga-through-Marcellus interval are nearly absent in western Genesee and Erie counties. In any case, this section and one in the nearby Federal Crushed Stone Corp. Quarry west of this stop, are the only outcrops showing any part of the Tioga complex of ashes in this region.

The upper metabentonite is a brown mudstone layer with a bronze-yellow micaceous luster which shows some evidence of bioturbation. The lower, less ash-rich unit, contains abundant fossils including rugose corals, bryozoans, pelmatozoan debris, and the trilobite Phacops. The corals, interestingly, are highly abraded and fragmental; this suggests that the lower "ash" might not be the record of a single "mega-eruption" or closely-spaced series of "eruptions" but is rather a condensed unit within Onondaga where pyroclastic debris and other terrigenous sediment became concentrated over a long period of time relative to carbonate. The occurrence of the uppermost Tioga unit ("Restricted Tioga" division of Conkin and Conkin, 1984; "Tioga A" Bed of Rickard, 1984) above the basal-Marcellus transgressive bone bed lag on uppermost Onondaga deposits, suggests a similar story of selective pyroclastic-enrichment associated with sediment-starved bottom conditions (Baird and Brett, 1986a, b).

This brief stop simply illustrates another unusual type of sediment associated with apparent sediment starvation. What is needed is detailed study of sedimentary structures and sedimentary petrography associated with the Tioga deposits to obtain answers concerning sedimentation rates and paleoenvironmental conditions associated with the formation of these enigmatic event-beds.

63.3	0.5	Board vehicles. Return to junction of Rowley and Broadway. Turn right onto Broadway and proceed east towards Lancaster.
63.8	0.5	Junction of Transit Road (U.S. Route 20) with Broadway (Route 130). Route 20 turns here to follow Broadway. Proceed straight (E) on Route 20.
64.4	0.6	Cross Cayuga Creek

64.5	0.1	Enter town of Lancaster, N.Y.
65.0	0.5	Cross Cayuga Creek. Middle Devonian black and dark grey shales of the Oatka Creek Member are intermittently exposed along creek in this area.
65.3	0.3	Junction of Lake Avenue with Route 20. Turn right (S) onto Lake Avenue.
65.5	0.2	Junction of Old Lake Avenue with Lake Avenue. Bear left (SE) onto Old Lake Avenue.
65.6	0.1	Junction of Old Lake Avenue with Pardee Avenue. Turn left onto Pardee Avenue.
	0.1	Turn right off of Pardee Avenue into a parking area at Como Park. Proceed on foot about 400 feet west to Lake Road bridge downstream from small dam on Cayuga Creek.

STOP 6. "PRECURSOR BED" AT BASE OF MIDDLE DEVONIAN STAFFORD LIMESTONE MEMBER AT COMO PARK.

Along the banks and bed of Cayuga Creek between the dam and base of the small waterfalls lip at the Lake Avenue bridge are exposures of the Stafford Limestone Member, the basal division of the Skaneateles Formation. Downstream from the water falls are intermittent exposures of the black, fissile, organic-rich Oatka Creek Member of the underlying Marcellus Formation which is mostly covered. Near the falls and bridge, the topmost few feet of the Oatka Creek Shale can be examined on the south side of the creek; these uppermost Marcellus beds are dark grey-brown in color and they yield a meager dysoxic biota consisting of the rhynchonellid brachiopod Leiorhynchus, Styliolina, and numerous flattened composite molds of an orthoconic nautiloid.

The Stafford, in Erie County, is a three-part member consisting of a basal, thin, shell-rich muddy limestone bed (Stafford "A" bed), a middle, shaley interval several feet thick which contains nodular micritic concretionary beds, and an upper cherty limestone division, termed the Stafford "B" bed, which is 0.6 - 1.3 m (2 to 4 feet) in thickness and fossiliferous (see Meyer, 1985). In central New York, the equivalent Mottville Member has a shell-rich "A" bed division, succeed by a variably-thick middle "shale" division followed by a micritic or siltstone regressive capping unit which corresponds to the Erie County, chert-rich ("B") micritic division visible by the Como Park dam (Grasso, 1986; Meyer, 1985). Despite significant thickness variations of the "middle shale" and "B" bed divisions along the Stafford-Mottville outcrop strike between Buffalo and the Chenango Valley, the "A" bed remains relatively thin, usually between 8 and 30 cm (0.2 and 1.0) foot in thickness, and it is typically a densely fossiliferous calcareous mudstone, both overlain and underlain by sparsely fossiliferous deposits. Although the "middle shale" usually grades upward into fossil-rich shaley micrites of the "B" bed, the "A" bed often rests abruptly on dysoxic to anoxic dark shales as it does at this section.

The "A" bed is a prime example of what we term a "precursor" bed, a layer which records what appears to be a sediment-starved regression event linked to, but clearly preceding, a subsequent longer-term regression which culminates in fossil-rich, regressive, limestone, siltstone, and even sandstone facies (see text). Such beds are enigmatic in terms of existing models of cyclic sedimentation, because they are followed by apparent deepening events associated with synchronous and subsequent influxes of sediment which then produce a regressive filling sequence that culminates in the "B" unit (see text discussion).

At this locality the Stafford "A" Bed yields numerous small brachiopods, including Ambocoelia umbonata var. nana, Truncalosia truncata, Devonochonetes scitulus, and small variety of Tropidoleptus. Other fossils include the large bivalve Panenka, occasional Leiorhynchus, orthoconic nautiloids often encrusted by the reptate biserial tubular organism Reptaria stolonifera, which may have "hitchiked" on the living cephalopod (see Baird et al., 1989), and wood debris.

This faunal association is rather typical of precursor beds we have seen in other units and it is nearly identical to that associated with the lower Wanakah "Darien Center" cycle at nearby localities (see Miller, FIELD TRIP SUN C, this guidebook). The diminutive brachiopod assemblage in the Stafford "A" bed appears to represent only a slight increase in bottom oxygenation relative to the underlying Oatka Creek Shale. This assemblage falls between the "Leiorhynchus" and "Ambocoelia-chonetid" biofacies of Brett et al., 1986; Vogel et al., 1986; which is indicative of non-turbid upper dysoxic to minimally oxic bottom conditions (see text).

66.0	0.4	Board vehicles and return via Pardee, Old Lake, and Lake avenues to Route 20. Turn right (E) toward Alden.
67.0	1.0	Leave town of Lancaster
67.3	0.3	Junction of Route 20 with Bowen Road. Turn right (S) onto Bowen Road.

67.7	0.4	Cross Cayuga Creek. The lower calcareous part of the Middle Devonian Levanna Shale Member is exposed along creek near bridge. One particularly prominent ledge upstream (east) from the bridge yields occasional excellent enrolled specimens of the trilobite <u>Phacops rana</u>.
70.1	2.4	Junction of Bowen Road and Clinton Road. Proceed straight (S) on Bowen Road.
70.6	0.5	Cross Buffalo Creek at Elma, N.Y. The Middle Devonian Ledyard Shale Member is exposed in the bed and banks of the creek. Pond Brook, a north-flowing tributary of Buffalo Creek just to the east of Bowen Road south of the bridge crossing has good exposures of the richly fossiliferous Wanakah Shale Member.
71.3	0.7	Junction of Bowen Road and Bullis Road, turn left (E) onto Bullis Road.
71.4	0.1	Cross the upper end of Pond Brook.
72.9	1.5	Junction of Bullis Road with Girdle Road.
73.1	0.2	Intersection with old Bullis Road (loops to south over old bridge). Turn right onto old road.
73.35	0.25	Park near old bridge over Buffalo Creek. Walk out onto bridge for brief overlook, then proceed down bank at west end of bridge and walk about 100 feet north (downstream) to limestone exposure adjacent to the new Bullis Road bridge.

STOP 7. CONDENSED CARBONATE DEPOSITS OF LOWER MOSCOW FORMATION AT BUFFALO CREEK.

In the low stepped falls interval between the old and new Bullis Road bridges one can observe a succession of thin carbonate beds representing condensed facies of the lower Moscow Formation and uppermost Ludlowville Formation (Fig. 4B). The Tichenor Limestone is visible here as at STOP 2, but it is both overlain and underlain by additional limestone layers; the underlying layer is the basal Hill's Gulch Bed of the largely-bevelled Jaycox Shale Member of the Ludlowville Formation and the overlying beds are in ascending order, the thin, condensed phases of the Deep Run, Menteth, and Kashong members (Fig. 4B). In western Erie County the Jaycox is entirely absent and the overlying complex of condensed units is absent or nearly absent due to further westward bevelling, such that the Tichenor is directly underlain by Wanakah Shale and is overlain by the Windom Member (compare Figure 4A and Figure 4B) .

Three discontinuities are present in this section; one at the base of the Hill's Gulch Bed which cuts into the Wanakah Shale; one below the Tichenor which cuts into the Hill's Gulch Bed, and one below the Windom Shale which bevels into a thin remnant of the Kashong Shale Member. The two lower discontinuities are marked by hypichnial, carbonate-casted burrow prods produced by arthropods excavating into underlying muds prior to burial of the discontinuity surface. The sub-Windom discontinuity is marked by phosphatic nodules, phosphatic steinkerns of Kashong fossils, and assorted shelly debris which have been churned and mixed by bioturbation processes during the onset of Windom deposition (Baird, 1978).

This condensed section is almost a "mirror-image" to that involving equivalent beds in central New York (Fig. 4c). Detailed mapping by Baird (1979), Baird and Brett (1981), Mayer (1989), Mayer et al. (1990) shows that each major bed in this outcrop has its time-stratigraphic counterpart in the condensed Jaycox-Kashong succession in the Owasco Lake-Ithaca region. Even the same discontinuities are represented in this latter area, though with the trend of increased bevelling and condensation in the eastward direction instead of westward, as is observed for the deposits here at Buffalo Creek. In the intervening Genesee Valley-Cayuga Lake region, each one of the componant divisions in this section balloons into thicker, mud-dominated facies, with each respectively higher division -- Jaycox, Tichenor, Deep Run, Kashong reaching maximum thickness west of that for the previous division (Brett et al., 1986). Hence, this intervening region records a westward-shifting depocenter and an aggregate maximum thickness for the Jaycox-Kashong interval exceeding 35 meters (110 ft.)!

Thus, the "mirror-image"-stacking of corresponding condensed beds in these widely-seperated regions is all the more significant given the complexity of depocenter shifts in the intervening region. It strongly suggests that eustatic sea-level changes were responsible for the succession of units, both within the depocenter areas, and on the adjoining shelf regions. Such a pattern makes all the more relevent the discussion of sequence stratigraphy applicability to the New York Devonian System (see text).

From the old Bullis Road bridge one can see the upstream bank of grey chippy Windom mudstone deposits. These rest on the sub-Windom (post-Kashong) phosphate debris-rich discontinuity which can be reached below water-level just upstream from the bridge foundation. Near the top of the Windom bank, the Bayview Coral Bed and the overlying Smoke Creek Bed can be seen. Enrolled Phacops rana can be collected from the Smoke Creek Bed further upstream in the floor of the creek. The giant blastoid Placoblastus sp. is a rare find at this locality; it occurs in the Deep Run-Menteth ledges above the Tichenor.

		Reboard vehicles. Return to new Bullis Road.
73.6	0.25	Junction of old Bullis Road with New Bullis Road. Turn right (east) onto Bullis Road and proceed toward Marilla.
75.95	2.35	Junction of Two Rod Road (Route 358) in Marilla. Turn left (north) onto Two Rod Road.
76.05	0.1	Cross Little Buffalo Creek. Middle-Upper Devonian disconformity is exposed downstream from (to northwest of) road crossing.
77.1	1.05	Intersection of Two Rod Road and Clinton Road (Route 354). Turn right (east) onto Clinton Road.
78.1	1.0	Intersection of Clinton Road with Four Rod Road. Continue straight on Clinton Road.
79.1	1.0	Intersection of Clinton Road with Three Rod Road. Continue straight on Clinton Road.

79.7 0.6 Cross Cayuga Creek. Pull off onto driveway of first house on right beyond the bridge or onto right shoulder of Clinton Road if parking space is limited.

STOP 8. MIDDLE-TO-UPPER DEVONIAN CONDENSED BASINAL DEPOSITS ALONG CAYUGA CREEK (See also Brett and Baird, 1982: NYSGA Buffalo meeting).

At this section we will first examine the Leicester Pyrite and its relationship to synjacent beds and then we will focus on the North Evans bone-conodont bed which marks the base of the Genundewa Limestone above the last falls riffle about 120 feet upstream from the bridge.

As with the lag deposit we observed below the Dunkirk at STOP 1, the Leicester Pyrite Member of the Genesee Formation is a detrital pyrite-bone-lag accumulation which directly underlies a laminated black shale unit (Geneseo Shale Member) and which occurs on a submarine discontinuity marking the base of the Geneseo. However, the Leicester is a more dramatic example of this lag type because it is made up of coarser grains and it is typically much thicker than the Dunkirk pyrite bed. Moreover, it rests on the Middle Devonian Windom Shale Member which is a grey mudstone unit usually rich in shelly fossils.

The Leicester is composed of pyrite burrow tube fragments, pyrite nodules, and pyrite fossil steinkerns derived from the Windom Member plus fish bone-and conodont debris of Windom and post-Windom age. These were exhumed under conditions of submarine anoxia or temporary dysoxia by episodic strong currents of unknown character. Reworked Windom (and Tully) lag debris was initially carbonate-dominated reflecting the overall calcitic and aragonitic character of Windom fossils. However, this carbonate debris underwent selective dissolution in a negative pH (or carbonate undersaturated) basin setting; a continuous blanket of reworked Windom carbonate debris, resembling the North Evans deposit at STOP 2, would hypothetically have been reduced to a placer lag of pyrite and bone debris only a small fraction of the original volume. Hence, the laterally discontinuous Leicester lenses represent only a remnant fraction of total Windom debris which was exhumed (Baird and Brett, 1986A). In actuality, during Geneseo time, no continuous carbonate debris blanket probably ever formed because dissolution would have removed carbonate as it was exhumed, allowing no carbonate to build up on the surface.

At this section as at other Leicester localities, some Leicester debris lenses occur within the basal Geneseo, not just at the base. This indicates that pyrite clast-transport occurred during the period of initial black mud accumulation. Morever, conodont examination by Huddle (1981) shows that the age of the youngest (Genesee-age) conodonts in the Leicester becomes progressively less as this unit is traced westward from the Seneca Valley into eastern Erie County; this westward progressive overlap of Geneseo black mud and lag debris onto the widespread Windom-Genesee disconformity surface tracks the major Taghanic Onlap Event which is now believed to mark a major eustatic transgression in the late Givetion (Johnson, 1970). This overlap cannot be followed westward indefinately; the sub-Genundewa discontinuity oversteps all lower Genesee strata in central Erie County and the Leicester is apparently cannibalized by this younger erosion surface between Cazenovia Creek and the north branch of Smoke Creek, southwest of Elma, Erie County (Brett and Baird, 1982; Baird and Brett, 1986A).

Evidence for reworking of Windom Pyrite clasts includes: 1, identical character of Windom and Leicester pyrite grains; 2, evidence of mechanical breakage of pre-formed pyrite grains; 3, abundance of pyritic Windom fossil steinkerns in Leicester; and 4, the reorientation of early diagenetic stalactitic pyrite which formed initially in voids within in-situ Windom pyrite tubes and steinkerns (see Baird and Brett, 1986a,b). Ongoing laboratory sulfur isotope studies indicate a hodgepodge pattern of variable isotopic weights between adjacent pyritic grains which further suggest exhumation and mixing of the pyrite material (see Lyons, 1990).

The mechanisms for physically exhuming and reworking pyrite in what would normally be thought of as a euxinic environment are difficult to reconstruct, given that there is no known modern example of this specific marine condition. Deep-storm waves are always a possibility as are various possible bottom current processes. Baird and Brett (1986A; Baird et al. (1988), offer an erosion model involving transgressive upslope migration of a water density-stratification boundery (pycnocline) along which storm generated internal waves are propagated. Impingment of the pycnocline with the basin slope at an given time allows for the presence of a higher energy wave-shoaling zone coincident with the slope-pycnocline intersection producing a swath of bottom erosion at a critical water depth/contour. Continue transgression would displace the wave-shoaling zone upward allowing basinal black muds and associated bone-pyrite debris to settle on the scour surface produced during earlier internal wave-shoaling. Overall westward Geneseo black mud overlap of the erosion surface may have been interrupted by brief episodes of major current flow and bottom scour events; these may explain the occurrence of Leicester lenses which are interbedded with basal Geneseo black shale deposits.

The Windom-Geneseo contact is a regional disconformity (combined sequence boundry,transgressive and down lap surface) which overlies different Windom beds at different western New York out crops. Generally, towards the north, the unconformity truncates progressive lower Windom beds. Conversely, towards the south, progresively higher Windom layers appear beneath the Leicester-Geneseo contact (Brett and Baird, 1982). At Cayuga

Creek one can observe a Windom shell-rich bed become progressivly truncated as it is traced from the falls south of the bridge northward. Near the falls, this Mediospirfer and Pseudoatrypa-rich mudstone layer, probably equivalent to the Fall Brook Coral Bed in the Genesee and Wyoming valleys, is well developed (Baird and Brett, 1983).

Above the Windom-Genesee contact there are approximately eleven feet of organic-rich black shale, dark grey shale, and concretion-bearing horizons which comprise the lower part of the Genesee Formation. This sequence of largely anoxic and minimally dysoxic basinal facies is extremely condensed; at Ithaca, New York, the equivalent stratigraphic interval is in excess of 150 m (500 feet) (deWitt and Colton, 1978; Baird and Brett, 1986A). Fossils which can be collected from the basal three feet of the Cayuga Creek Geneseo section include current-aligned Styliolina, the brachipod Devonochonetes, and scattered wood fragments.

At the highest falls, dark, organic-rich and conspicuously laminated beds of the Penn Yan Shale Member are abruptly overlain by the 20-25 cm (8 to 10 inch) thick ledge of the Genundewa Member and its associated 1-2.5 cm (0.5 to 1.0 inch) thick basal lag veneer of North Evans reworked pelmatozoan, conodont, and bone debris. Further east in Genesee County, this lag unit apparently disappears and the base of the Genundewa become conformable. Although the North Evans at this locality is a bit less conspicuous than it is at STOP 2, it still yields glauconite-coated reworked concretions and a great abundance of conodonts. As at STOP 2, the Genundewa is a dense, grainstone to packstone blanket accumulation of Styliolina shells. The grotesque Penn Yan concretionary bed below the North Evans contact is also developed at other adjacent sections; it may have formed as a geochemical response to the proximity of the North Evans erosion surface which was being scoured out from above.

END FIELD TRIP

STRATIGRAPHY, STRUCTURAL GEOLOGY, AND HYDROGEOLOGY OF THE LOCKPORT GROUP: NIAGARA FALLS AREA, NEW YORK

DOROTHY H. TEPPER
Water Resources Division
U.S. Geological Survey
Ithaca, New York 14850

WILLIAM M. GOODMAN
Department of Geological Sciences
University of Rochester
Rochester, New York 14627

MICHAEL R. GROSS
Department of Geosciences
Pennsylvania State University
University Park, PA 16802

WILLIAM M. KAPPEL
Water Resources Division
U.S. Geological Survey
Ithaca, New York 14850

RICHARD M. YAGER
Water Resources Division
U.S. Geological Survey
Ithaca, New York 14850

INTRODUCTION

The Niagara Falls area is known for its classic geology. An internationally recognized stratigraphic sequence of uppermost Ordovician and Silurian rocks is exposed along the walls of the Niagara Gorge. Summaries of the geology of this area are provided in Tesmer (1981) and Brett and Calkin (1987).

The purpose of this field trip is to provide an overview of some current research on the stratigraphy, structural geology, and hydrogeology of the Upper Silurian[1] Lockport Group, which caps the section exposed in the Niagara Gorge. There has been much recent interest in the Lockport Group, particularly in the Niagara Falls area, where ground water within it has been contaminated by hazardous waste. In 1987 the U.S. Geological Survey (USGS), in cooperation with the U.S. Environmental Protection Agency (USEPA), began a 4-year study of the hydrogeology of the Niagara region. The primary objective of this study is to define the regional ground-water flow system so that a three-dimensional ground-water flow model can be constructed that can provide a regional hydrogeologic framework for future site-specific research efforts. An important supporting objective of this study is

[1] The U.S. Geological Survey recognizes "Lower," "Middle" and "Upper" Silurian and designates the Lockport Group as Middle Silurian. However, in a New York State Geological Survey publication by Rickard (1975), the use of "Middle" Silurian was abandoned in New York and was replaced with the terms "Lower" and "Upper" Silurian to be consistent with European usage. The Lockport Group is designated as Upper Silurian by Rickard (1975) and will be referred to as such in this paper.

to identify the major water-bearing zones and to determine their regional extent, which has required detailed study of the stratigraphy and structural geology of the Lockport Group.

High-resolution stratigraphy has been used to aid in identification and correlation of horizontal regional water-bearing zones. Brett and others (written commun. 1990)[2/] have divided previously recognized stratigraphic units into new regionally extensive formations, members, beds, and informal units and have extended some units into the study area from adjacent areas where they were previously defined and(or) made changes in the placement of their lower or upper contacts. These revisions, which are the result of research conducted at the Department of Geological Sciences at the University of Rochester, in cooperation with the USGS, have been based on observations of outcrops and of cores obtained through the USGS/USEPA project. The revised nomenclature discussed herein is provisional and is formally proposed within a larger study of the Niagaran Series (C. E. Brett and others written commun. 1990) and comply with the requirements of the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature 1983). A summary of this work for the Lockport Group is presented in the "Stratigraphy" section.

Ground water in the Lockport Group flows primarily through extensive, nearly horizontal bedding planes that are connected to a limited extent by high-angle fractures. Because these fractures significantly affect ground-water flow rates and directions, it is important to determine the regional and local fracture framework. A study of fractures in the Lockport Group of western New York and southern Ontario was completed by Gross (1989); results of that work are summarized in the "Structural Geology" section.

The three stops on this field trip (Fig. 1) are described in detail in the following sections. Stratigraphy and structural geology will be discussed at the first two stops and hydrogeology will be discussed at the last stop. The first stop will be at the Frontier Stone Quarry in the Town of Gasport, where we will examine the stratigraphy of the Gasport Limestone, the lowermost formation in the Lockport Group. This quarry is located along a reentrant on the Niagara Escarpment so we will also examine changes in joint orientations along this feature. The second stop will be at the Niagara Stone Quarry in the Town of Niagara, where we will examine the Lockport Group stratigraphy from the base of the Gasport Limestone to the upper part of the Eramosa Dolomite (previously termed the "Oak Orchard Dolomite"). We will also discuss fractures and large-scale "pop-ups" on the quarry floor. The final stop (also in the Town of Niagara) will be at a site where the USGS has conducted detailed electromagnetic surveys and hydraulic tests as part of the USGS/USEPA regional hydrogeologic study. This stop will include a demonstration of electromagnetic techniques, including terrain conductivity and VLF (very low frequency), and a discussion of downhole instrumentation and the cross-hole hydraulic testing program that has been conducted at the site.

STRATIGRAPHY

by William M. Goodman and Dorothy H. Tepper

The stratigraphy of the Niagara area is well known; some of the pioneering paleontologic and stratigraphic studies in the region were conducted by the renowned geologist James Hall, beginning in 1838. These early studies resulted in the establishment of the Niagara region as the type locality for the Niagaran Series (Lower and Upper Silurian[1]) of eastern North America. The Niagara area is underlain by approximately 60 m (meters) of relatively undeformed dolomites and limestones of the Upper Silurian Lockport Group of the Niagaran Series. The rocks in the Niagara area were deposited in shallow epeiric seas in the northern Appalachian Foreland Basin. These units are widespread; their lateral equivalents outcrop to the northwest of the study area and eventually extend into the Michigan Basin. Cyclic vertical facies changes and erosional surfaces in the Lockport Group are related to relative sea-level oscillations and to emergence of the Algonquin Arch concurrent with lateral migration of the Appalachian Basin axis (Brett and others 1989).

Basis for Stratigraphic Revisions

One purpose of this field trip is to present revised stratigraphic nomenclature (Fig. 2) for the

2/ (C. E. Brett, University of Rochester; W. M. Goodman, University of Rochester; S. T. LoDuca, University of Rochester; D. H. Tepper, U.S. Geological Survey; W. L. Duke, Pennsylvania State University; and B. Y. Lin, written commun. 1990) will be cited hereafter as (C. E. Brett and others, written commun. 1990).

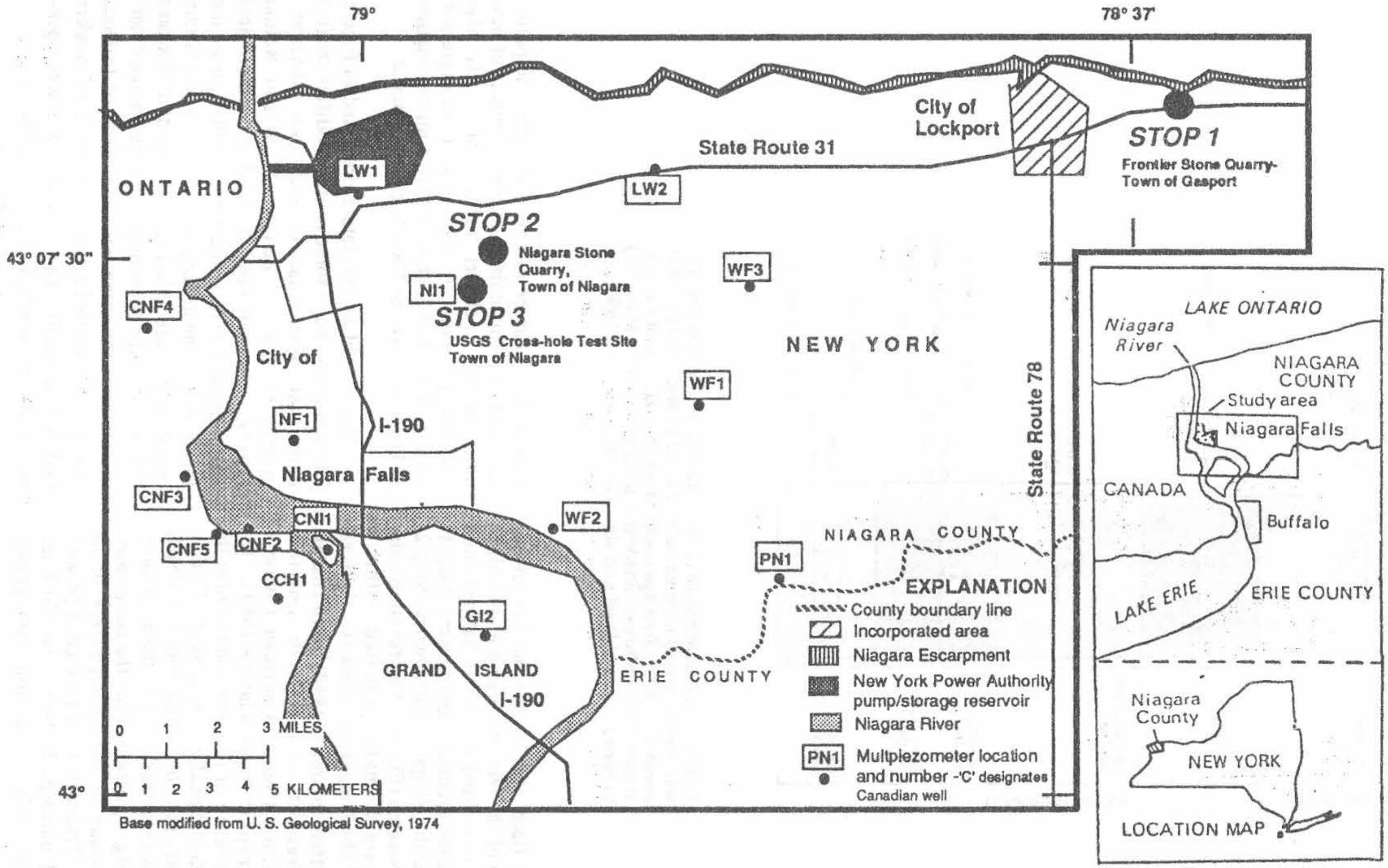


Figure 1. -- Location of field-trip stops and multipiezometer installations where core was collected.

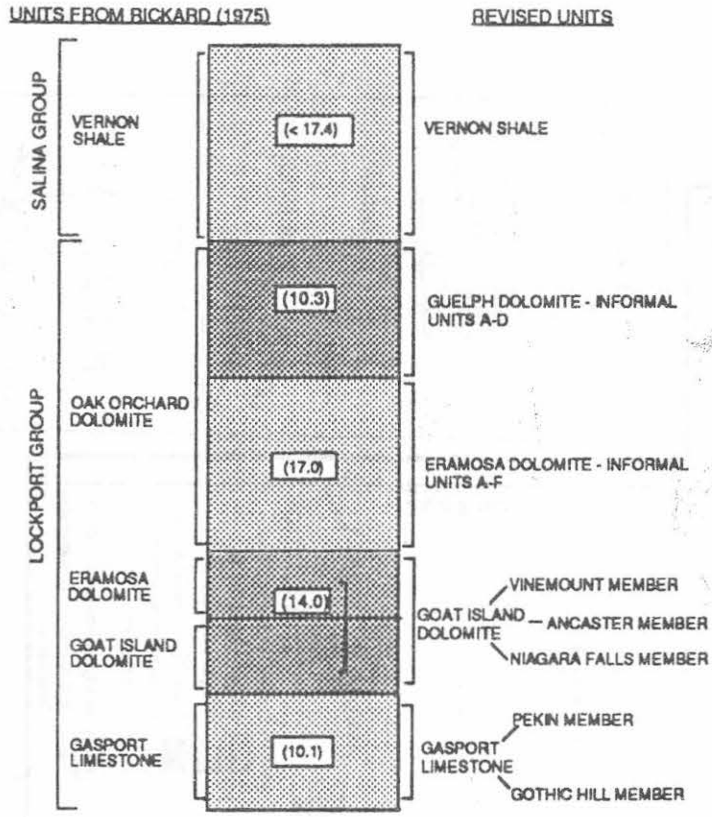


Figure 2. - Comparison of units as defined by Rickard (1975) with newly revised units as defined by Brett and others (written commun. 1990). Average thicknesses of units (in meters) are shown in parentheses. Shading of units in column is according to revised units of Brett and others (written commun. 1990).

Upper Silurian (late Wenlockian to Ludlovian) Lockport Group of the Niagara region and to highlight the details of revised and newly defined units as they appear at the Frontier Stone Quarry (Stop 1) at Gasport, and at the Niagara Stone Quarry (Stop 2) in the Town of Niagara (Fig. 1). **The revisions discussed herein are formally proposed within a larger study of the Niagaran Series (C. E. Brett and others, written commun. 1990)[2/] and comply with the requirements of the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature 1983).** The revisions are the result of research conducted at the Department of Geological Sciences at the University of Rochester, in cooperation with the USGS. This research has been facilitated by the availability of nine cores (Fig. 1) obtained for the cooperative USGS/USEPA study of the hydrogeology of the Niagara region. Traceability of rock units between these cores and outcrops in both New York and Ontario indicates that a uniform stratigraphic nomenclature can be established across the political boundary. In addition to the definition of new units in the Lockport Group, this study also resolves conflicts in the present stratigraphic nomenclature that have arisen from miscorrelation of key units between New York and Ontario.

The revised nomenclature presented in Figure 2 builds upon and to some degree modifies and(or) replaces the nomenclature based upon lithostratigraphic analyses of Zenger (1965) and Bolton (1957). Furthermore, detailed paleontological sampling by LoDuca and Brett (1990 a,b) of the formations in the lower part of the Lockport Group modifies the age relations presented in Rickard (1975). In addition to improvements in age determinations at the stage level, the work of LoDuca (1990) has provided a useful faunal/floral marker bed that has facilitated correlation of lower Lockport units between widely spaced sections.

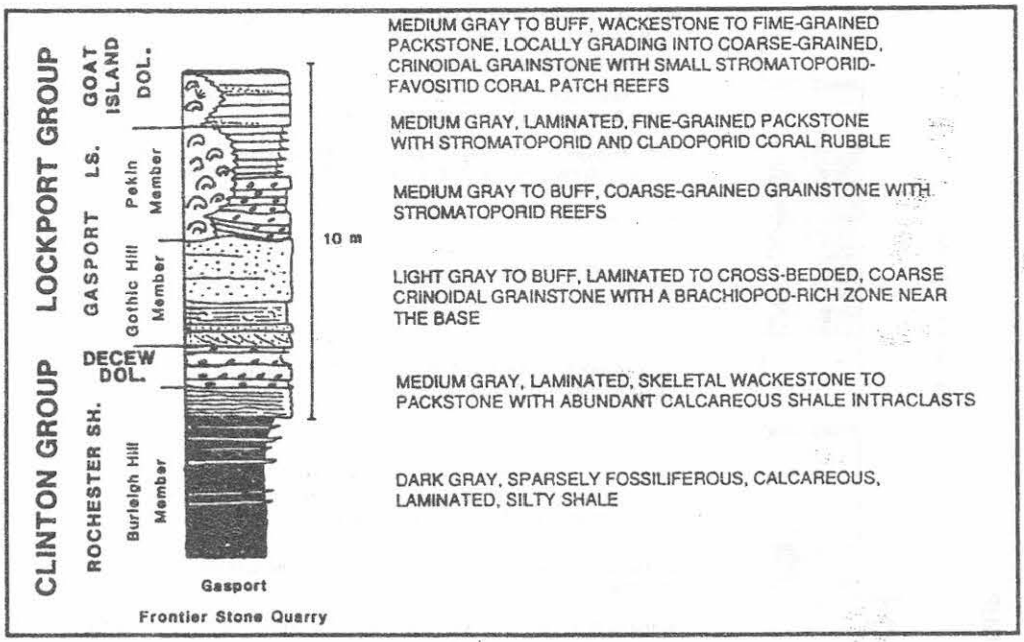


Figure 3.-- Stratigraphic section from the Frontier Stone Quarry in Gasport (Stop 1, Fig. 1).

At the two quarry stops on this trip, we will see three of the four formations that constitute the lower 37 m of the entire 50 to 60 m of medium-bedded to massive, argillaceous limestones and dolomites and thin shales that form the Lockport Group. The following discussion includes a general description of the newly revised stratigraphic units, in ascending order. The contact of the Lockport Group with the underlying Clinton Group is placed at the disconformable contact between the DeCew Dolomite and the Gasport Limestone and can be seen at the Frontier Stone Quarry at Gasport (Stop 1). Details of the stratigraphy at this quarry are shown in Figure 3. A composite stratigraphic column based on the exposures at the Niagara Stone Quarry in the Town of Niagara (Stop 2) and on core from the USGS Niagara (NI-1) test hole (Stop 3) is shown in Figure 4. The contact between the Lockport Group and the overlying Salina Group is not exposed in outcrop but can be seen in the USGS Grand Island (GI-2) core (Fig. 5). The gradational boundary of the Guelph Dolomite and the overlying Vernon Shale is arbitrarily placed by Brett and others (written commun. 1990) at the first black shale bed that is greater than 2.5 cm (centimeters) thick.

Revised Stratigraphy of the Lockport Group

As defined by Brett and others (written commun. 1990), the Upper Silurian Lockport Group of the Niagara region consists of the following four formations, in ascending order: the Gasport Limestone; the Goat Island Dolomite; the Eramosa Dolomite; and the Guelph Dolomite. Decisions on the names of the revised formation- and member-scale units have been based on the quality of the type sections, established stratigraphic boundaries, and historical precedence. Although three of the four formations retain traditional names based on New York sections, the contacts between units with Canadian type sections have been shifted considerably for consistency with Canadian usage. The revised stratigraphic units are outlined below.

Gasport Limestone. The type section for the Gasport Limestone is in the Frontier Stone Quarry at Gasport, in Niagara County (Stop 1). The Gasport Limestone as defined by Brett and others (written commun. 1990) consists of the Gothic Hill Grainstone Member (lower member) and the Pekin Member (upper member) (Fig. 4). The Gothic Hill Grainstone Member is a light pinkish gray to buff,

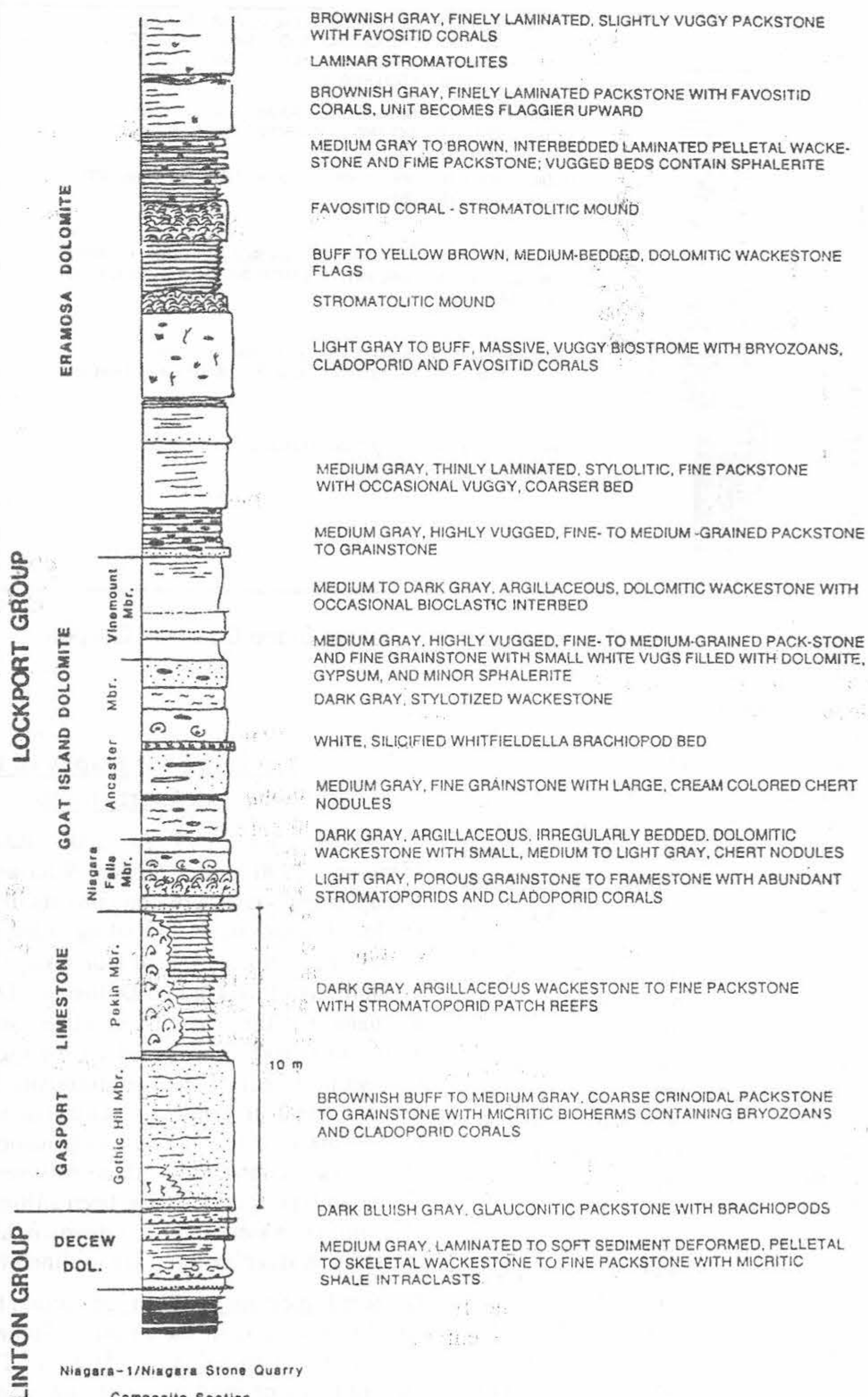


Figure 4.--Composite stratigraphic section based on exposures at the Niagara Stone Quarry in the Town of Niagara (Stop 2) and on the U.S. Geological Survey Niagara (NI-1) core (Stop 3).

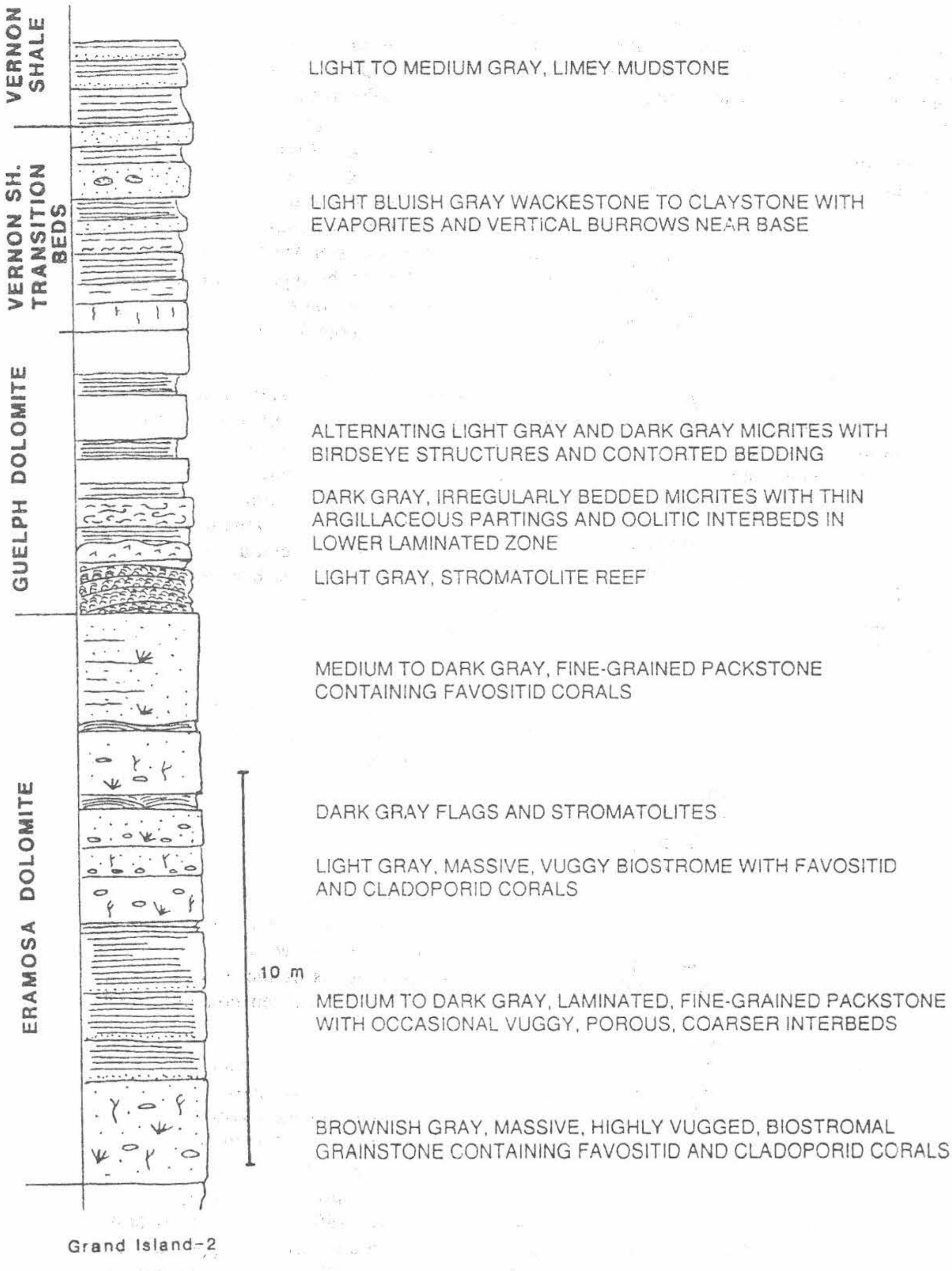


Figure 5.--Stratigraphic section from the U.S. Geological Survey Grand Island (GI-2) core (location shown on Fig. 1).

coarse crinoidal packstone to grainstone containing micritic, bryozoan-rich patch reefs. The Pekin Member is a dark olive gray, argillaceous wackestone to fine-grained packstone that commonly contains stromatoporoid, favositid, and cladoporid coral patch reefs.

The basal contact of the Gothic Hill Grainstone Member with the underlying DeCew Dolomite of the upper part of the Clinton Group is a sharp erosional unconformity in the Niagara region. The upper contact of the Gothic Hill Grainstone Member with the overlying Pekin Member is sharp but conformable over 0.3 to 0.6 m of condensed biostromal to biohermal beds. The upper contact of the Pekin Member with the overlying Niagara Falls Member of the Goat Island Dolomite ranges from gradational to sharp, depending upon local variations in the grain size of the basal beds of the Niagara Falls Member of the Goat Island Dolomite.

In the USGS cores, the thickness of the Gasport Limestone ranges from 5.9 to 11.3 m and averages 10.3 m. Both its members are highly variable in thickness in the Niagara region; the Gothic Hill Grainstone Member ranges from 0.9 to 6.4 m thick, whereas the Pekin Member ranges from near pinchout to as much as 8.2 m thick.

Goat Island Dolomite. The type section for the Goat Island Dolomite is on the north side of Goat Island, at the brink of Niagara Falls (Howell and Sanford 1947). The Goat Island Dolomite as redefined by Brett and others (written commun. 1990) has been divided into three members: the Niagara Falls Member; the Ancaster Member; and the Vinemount Member (Fig. 2).

The Niagara Falls Member is the basal member of the Goat Island Dolomite. In the Niagara region, it is similar to the Gothic Hill Grainstone Member of the Gasport Limestone in that it locally consists of very coarse, pinkish gray, crinoidal grainstones with bioherms containing stromatoporoids, and favositid and cladoporid corals. The coarse grainstones typically flank the small patch reefs and commonly grade laterally into finer grained, medium-bedded wackestones or into fine packstones over broad areas where reefs are absent. In these extensive non-reef areas, the contact between the Niagara Falls Member and the underlying Pekin Member is obscure. The thickness of this unit is variable, ranging from less than 0.9 m at the north end of Niagara Gorge to over 4.6 m at the Frontier Stone Quarry in Lockport.

The Niagara Falls Member is sharply but conformably overlain by the Ancaster Member, which consists of medium gray, medium-bedded, cherty wackestones to fine packstones. Chert nodules are commonly cream colored but may be dark gray and contain well-preserved rugose corals, brachiopods, gastropods, and, rarely, trilobites. The Ancaster Member, like the Niagara Falls Member below it, is variable in thickness and the two units appear to have a compensatory thickening/thinning relation. Thickness of the Ancaster Member ranges from a maximum of 7.6 m at the type locality to a minimum of 0.6 m at the south end of the Niagara Gorge and averages about 3.7 m.

The Ancaster Member is overlain by the Vinemount Member of the Goat Island Dolomite. The contact between the two members is relatively sharp but conformable. "Vinemount" is a Canadian term that will, in accordance with revisions in Brett and others (written commun. 1990), be used in New York consistent with Canadian usage. In the Niagara region, the Vinemount Member is a medium gray, thin- to medium-bedded, dolomitic, argillaceous wackestone to fine packstone with numerous thin, calcareous shale partings. The Vinemount Member is sparsely fossiliferous; the most common fossils are diminutive rugose corals and brachiopods. Beds are commonly bioturbated and take on a "vermicular" appearance. The thickness of the Vinemount Member in the Niagara region is relatively uniform, ranging from 3.0 to 6.1 m.

The contact between the Vinemount Member of the Goat Island Dolomite and the overlying Eramosa Dolomite is placed at the sharp contact between the thin-bedded, argillaceous carbonates and shales of the Vinemount Member and the massive dolomitic packstones of the now-abandoned (C. E. Brett and others, written commun. 1990) Oak Orchard Dolomite.

Eramosa Dolomite. Use of the name "Eramosa" has become confused in the past century. As revised by Brett and others (written commun. 1990), to be consistent with Canadian usage, the Eramosa Dolomite immediately overlies beds in western New York previously designated as Eramosa by Zenger (1965), which are regarded herein as the Vinemount Member (upper member) of the Goat Island Dolomite. As redefined by Brett and others (written commun. 1990), the Eramosa Dolomite is a tripartite massive-flaggy-massive unit that contains several distinctive marker beds.

The lower and upper members are characterized by massive-bedded to biostromal, fine grainstones containing variably abundant favositid and cladoporid corals. The basal bed of the Eramosa Dolomite is a distinctive, highly vugged 0.9- to 1.5-m-thick unit that is traceable over broad areas of western New York. The upper fine grainstone contains a medial stromatolite marker bed that is widely traceable although it is typically only 0.3 m or less thick. The middle member of the Eramosa Dolomite is flaggy and contains large, loaf-shaped stromatolitic, thrombolitic, and favositid coral mounds. The Eramosa Dolomite ranges from approximately 13.7 to 16.8 m in thickness. The contact between the Eramosa Dolomite and the overlying Guelph Dolomite is placed at the sharp boundary between the upper grainstone unit and an overlying distinctive, regionally widespread stromatolite reef horizon.

Guelph Dolomite. The name "Guelph Dolomite" has been extended into the Niagara region by Brett and others (written commun. 1990) and constitutes what has previously been termed the upper part of the Oak Orchard Dolomite. Outcrops of these strata are scarce. However, the USGS Grand Island (GI-2) core includes the entire Guelph Dolomite (Fig. 5). The basal Guelph consists of a lower 1.2- to 1.5-m-thick stromatolitic to thrombolitic reefy zone. The algal structures are overlain by argillaceous micrite with distinctive contorted bedding and oolitic layers. This enterolithic unit passes upward into thin- to medium-bedded peritidal micrites that contain vertical burrows and evaporites. The upper 4.6 m of these micrites have been labelled "Vernon transition" beds; they record the gradual change from carbonate to fine- grained clastic depositional environments in the Niagara region.

STRUCTURAL GEOLOGY

by Michael R. Gross

The sediments within the Niagara area strike approximately east-west and dip to the south at about 5.3 m/km (meters per kilometer), in a homoclinal structure. The most prominent structural feature in the Lockport Group is the Clarendon-Linden Fault (Fig. 6A), which has been interpreted as a series of north-south-trending, basement-related normal faults that intersect the Lockport Group outcrop belt in the vicinity of Clarendon (Fakundiny and others 1978). The two main systematic fracture sets within the Lockport Group are an east-northeast calcite vein set and an east-northeast joint set. Rose diagrams of fracture orientations for vertical veins and joints are shown in Figure 6B; these fractures were mapped in 17 quarries and one roadcut along the 250-km-long outcrop belt of the Lockport Group in western New York (Fig. 6A) and southern Ontario. Most of the fractures in these quarries are vertical.

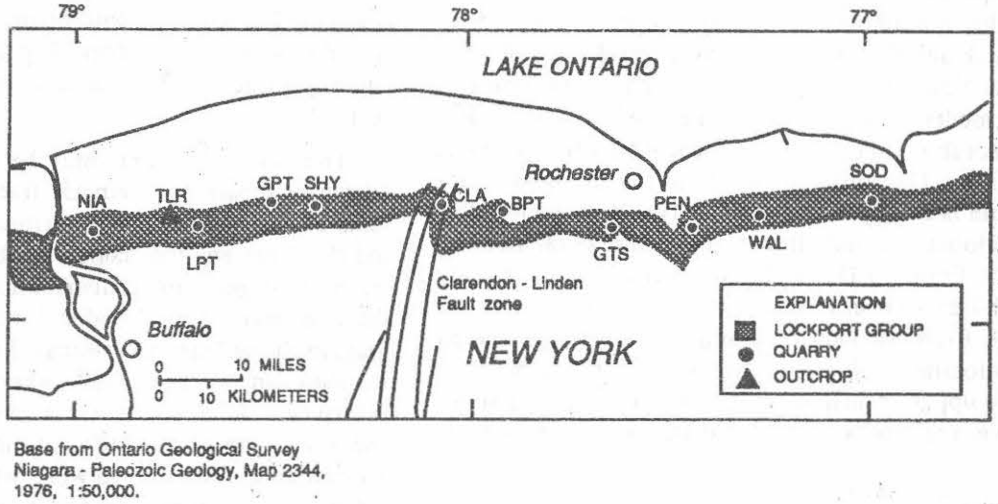
The veins and joints may be distinguished as distinctly separate systematic fracture sets on the basis of their geographic distribution, orientation, and the depth and mechanism of their propagation. In terms of geographic distribution, the veins are absent between the Clarendon-Linden Fault and the Niagara Stone Quarry, whereas the joints are most common in those areas where the Niagara Escarpment is a prominent topographic feature. The veins display a significant clockwise rotation in orientation from east to west, whereas the joints and topographic reentrants are consistently oriented across the outcrop belt. The veins also differ from the joints in their depth and mechanism of propagation; the veins appear to have propagated under considerable overburden, possibly as a result of high fluid pressures driven by a topographic high to the east, whereas the joints appear to have propagated later and closer to the surface, possibly during uplift and erosion.

Asymmetric Reentrants in the Niagara Escarpment: A Case for Neotectonic Joints

The northeastern United States is currently in a state of compression, with the maximum horizontal stress oriented approximately 060° (Sbar and Sykes 1973; Zoback and Zoback 1980). An east-northeast regional fracture set is present in many localities throughout the Northeast that correlates with this orientation (Engelder 1982). According to Hancock and Engelder (1989), these neotectonic joint systems are the most recent to form within a region subject to uplift and erosion, and they generally form within the upper 0.5 km (kilometer) of the crust in response to an effective tensile stress developed during unloading. These late-formed or neotectonic joints in some terrains are likely to reflect the orientation of the neotectonic or contemporary tectonic stress field.

The general trend of the Niagara Escarpment is east-west, although it actually consists of a series of angular reentrants that form a zig-zag pattern (Fig. 7). The distinct linear aspect of the Niagara Escarp-

A.



B.

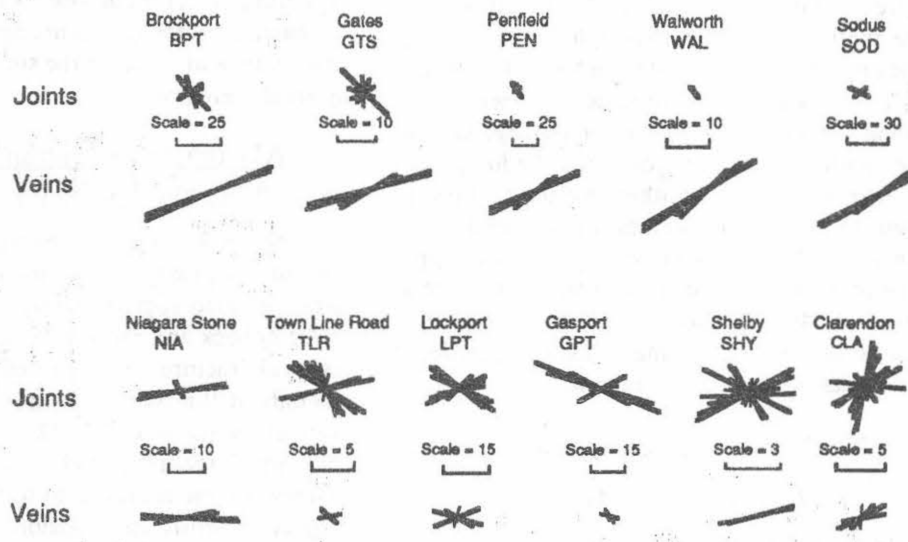


Figure 6. -- A. Locations of quarries and outcrops referenced in 6B. (Modified from Gross 1989). B. Rose diagrams of fracture data from 11 sites between Sodus and Niagara Falls, New York. The data are divided into two categories: joints and veins. Rose diagrams are in 5° intervals, and the scales differ from site to site. (Modified from Gross 1989).

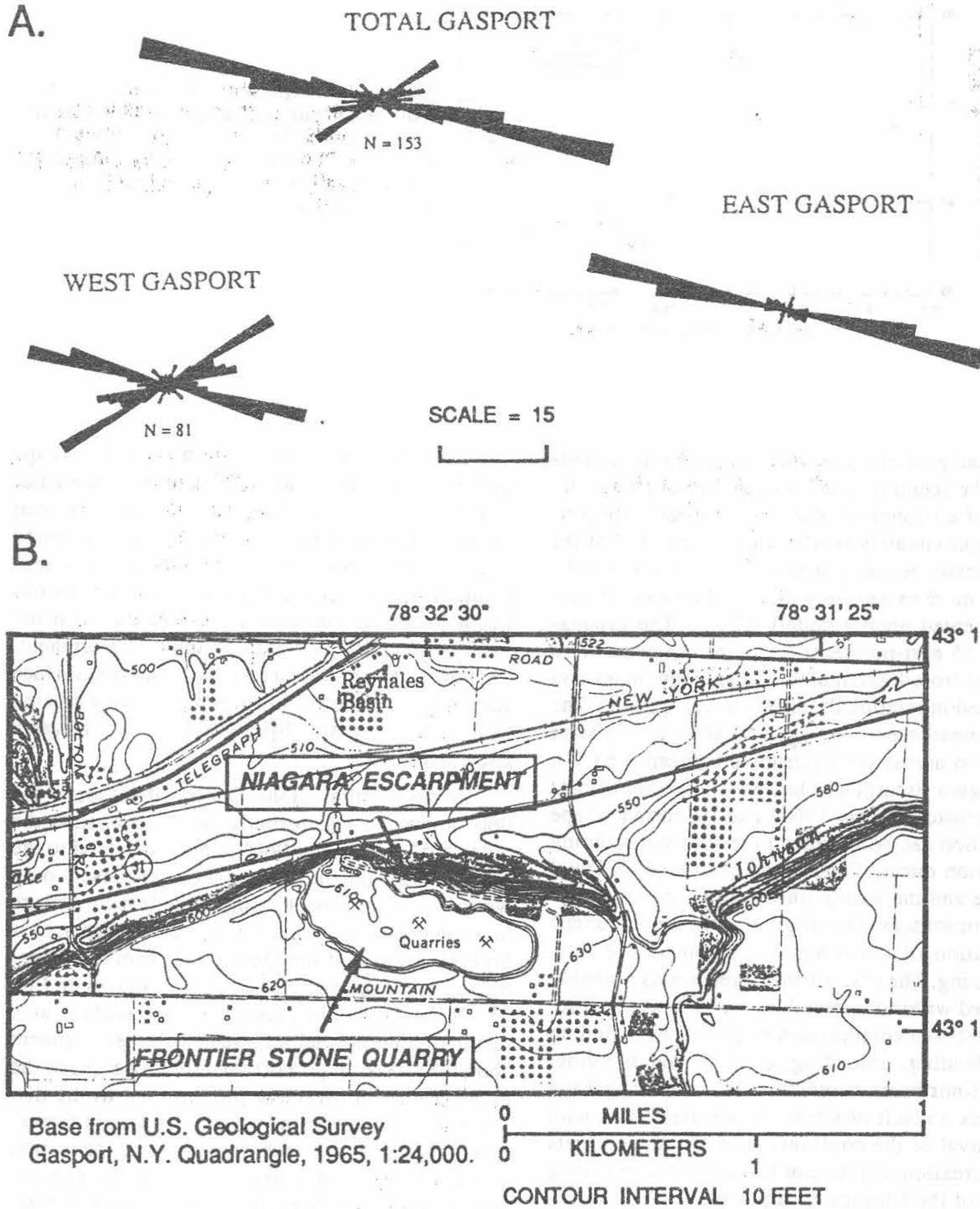


Figure 7. -- A. Rose diagrams of fractures from the eastern and western sectors of the Frontier Stone Quarry in Gasport (stop 1). (Modified from Gross 1989). B. The quarry is located at a major change in joint orientation along the Niagara Escarpment.

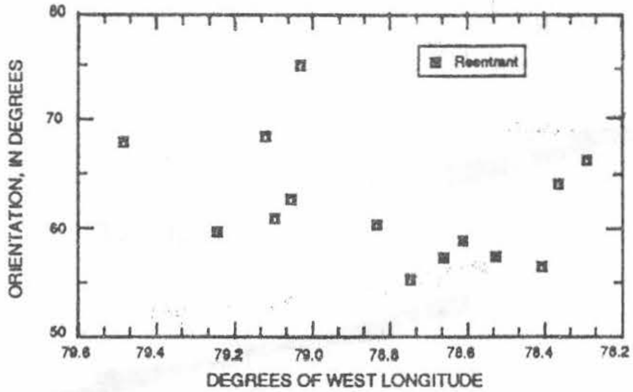


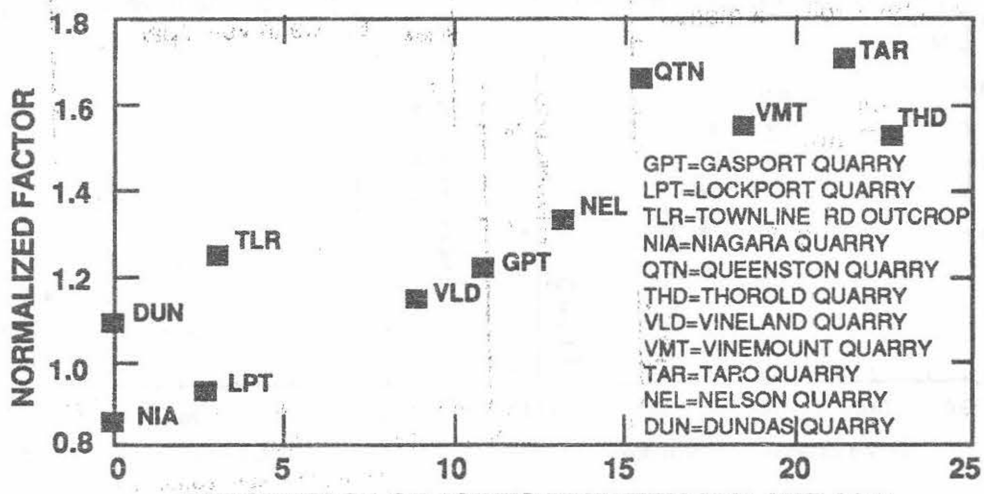
Figure 8. - - Orientation of east-northeast reentrant segments in relation to longitude. The reentrants show a consistent orientation across the outcrop belt (mean = 61.7°). (Modified from Gross 1989).

ment clearly reflects the fracture pattern in the bedrock. The reentrants are asymmetric and typically consist of a strongly linear east-northeast face (oriented approximately 060°), which is parallel to the contemporary tectonic stress field, and a less well-defined, more extensively dissected west-northwest face (oriented approximately 105°). The orientations of 15 east-northeast reentrant segments were measured from topographic and geologic maps and are plotted in relation to longitude in Figure 8. The east-northeast reentrant segments show a consistent orientation across the outcrop belt (mean = 61.7°). The Niagara Escarpment has undergone significant glacial erosion (Straw 1968) and its current shape and position can be assumed to result from scouring and erosion during Pleistocene glaciation. Glacial evidence and the strong linearity of the east-northeast reentrant segments imply that the current shape and position of the Niagara Escarpment are relatively young. The east-northeast reentrants probably developed within the past 3 million years and, if so, may relate to a combination of processes including glacial loading, unloading, erosion, and scouring. The east-northeast neotectonic features may have formed as a result of stress release associated with the removal of the confining load of the ice sheets and the creation of a state of low normal stress along the face of the Niagara Escarpment.

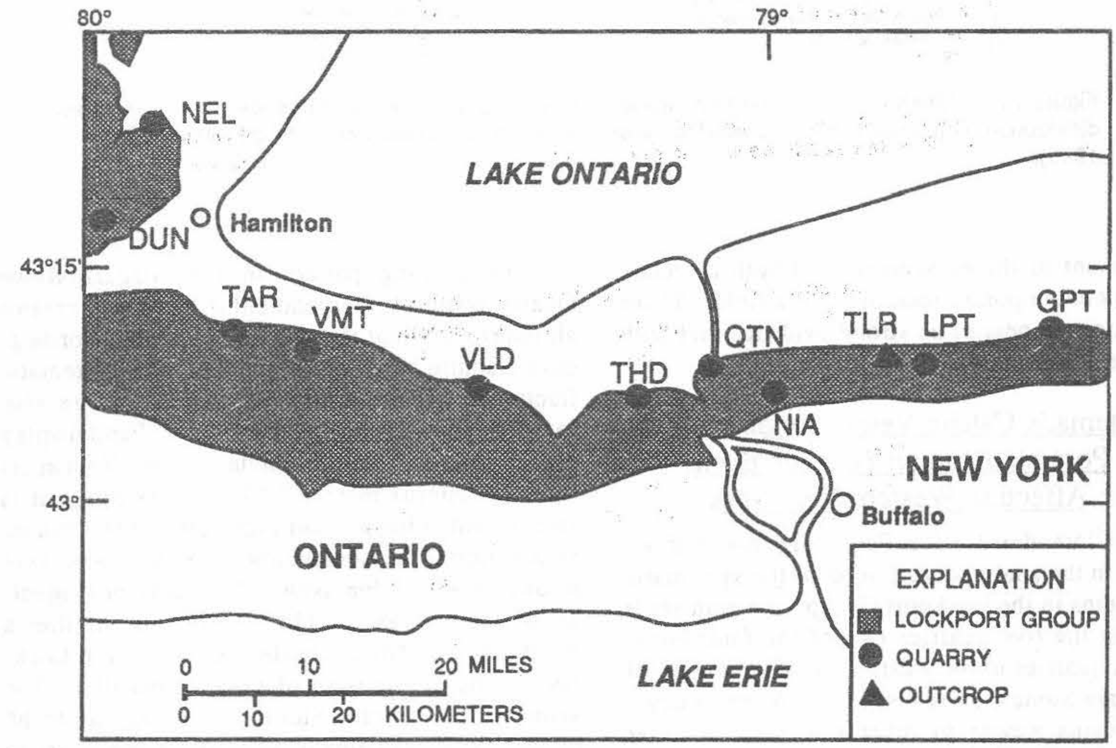
The Frontier Stone Quarry in Gasport (Stop 1) is located at a major change in joint orientation along the escarpment and thus permits study of the effect of topography on fracturing; that is, the relation between escarpment orientation and bedrock fractures. The escarpment is oriented approximately 105° adjacent to the eastern sector of the quarry and about 060° adjacent to the western sector (Fig. 7). The quarry is approximately 1.5 km across. The rose diagram of the eastern sector shows the dominant fracture set at 105°, parallel to the adjacent escarpment wall. The 105° joint set is present in the western sector along with a joint set oriented 060°, which is parallel to the adjacent 060° segment. The 060° set is absent in the eastern sector. Local joint sets correlate in orientation with the local orientation of the Niagara Escarpment. Therefore, these fractures form near the surface and may be related to low normal stresses created by the topographic differential of the Niagara Escarpment.

In the Frontier Stone Quarry at Gasport, the 060° joints are late forming and abut against the older 105° joints. The 060° joints are related to the Niagara Escarpment because they are present only where it is a prominent topographic feature and are most abundant in quarries where the escarpment is highest, such as at the Queenston, Thorold, Vinemount, and Taro quarries (locations shown on Fig. 9). If these fractures form near the surface as a result of low normal stresses, their development should be directly proportional to the height of the escarpment and inversely proportional to the distance from the escarpment. A "normalized factor" was developed for each locality that takes into account height and proximity to the escarpment, and this factor has been plotted in Figure 9 in relation to the percentage of joints within a 10-degree interval of the neotectonic orientation of 060°. Although it is difficult to prove quantitatively that these fractures are neotectonic, qualitatively they are consistently late forming and appear to correlate with the prominence of the Niagara Escarpment. The distinction, if any, between joints and the east-northeast reentrant segments is not clear because both are linear features related to the

A.



B.



Base from Ontario Geological Survey
Niagara - Paleozoic Geology,
Map 2344, 1976, 1:50,000

Figure 9.-- A. Normalized factor for selected quarries and outcrops in relation to the percentage of joints in the 10° interval surrounding the neotectonic orientation of 060°. (Modified from Gross 1989). B. Locations of quarries and outcrops referenced in 8A. (Modified from Gross 1989).

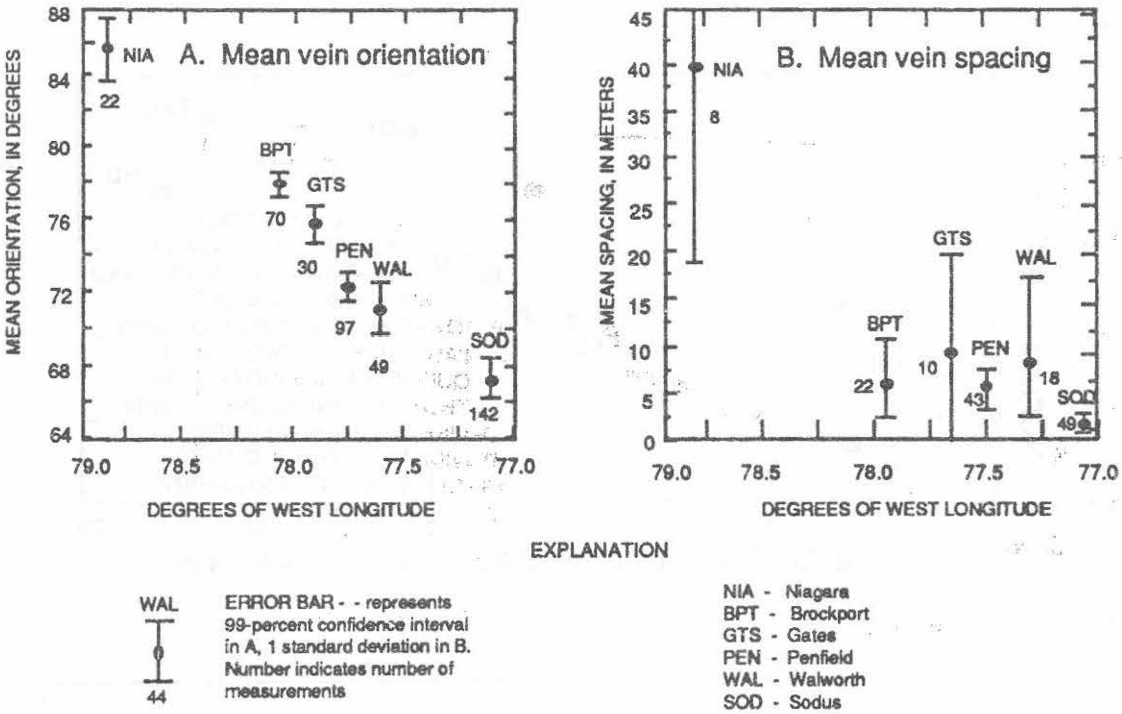


Figure 10. - - Mean vein orientation and mean vein spacing in relation to longitude. The von Mises distribution (Cheeney 1983) was used to determine mean vein orientation. (Modified from Gross 1989).

development of the escarpment and both correlate with the contemporary tectonic stress field. These observations appear to be strong evidence that both are neotectonic.

Systematic Calcite Veins: Evidence for a Possible Acadian Tectonic Event Affecting Western New York

The Clarendon-Linden Fault (Fig. 6A) plays a key role in the spatial distribution of the systematic calcite veins in the Lockport Group: the vein set is present in the five quarries east of the fault but is absent in quarries to the west, with the exception of the Niagara Stone Quarry (Fig. 6B). A few nonsystematic veins appear in other quarries. Another characteristic of the systematic vein set is that the mean vein orientation rotates clockwise from 067° in the east to 086° in the west (Fig. 10A). The veins are extremely consistent in orientation at each outcrop, as indicated by a 99% confidence interval of less than 2°. The veins do not appear to propagate into the DeCew Dolomite (immediately below the base of the Lockport Group), and vein spacing generally increases from east to west (Fig. 10B).

The blasting pattern in the Niagara Stone Quarry results in perpendicular faces and creates an impression that the bedrock consists of orthogonal fracture sets; however, only one systematic fracture set is present--the east-west calcite vein set. The veins are oriented 086° ± 1.7° and display a mean spacing of approximately 37 m. Vein apertures are approximately 2.23 cm. Although it is rare to find suitable fluid inclusions in the calcite vein material, homogenization temperatures were measured in five inclusions. The mean homogenization temperature of 115° C (Celsius) implies a minimum of 3 km of overburden above the Lockport Group at the time of vein propagation. The veins in the Niagara Stone Quarry appear to be related to the systematic vein set in the quarries east of the Clarendon-Linden Fault. Their orientation is east-northeast (consistent with the clockwise rotation) and they are remarkably consistent in orientation within the quarry. The veins are planar, vertically and horizontally extensive, and consist of the same fill material.

A model of the geologic conditions that would result in the propagation of the systematic vein set

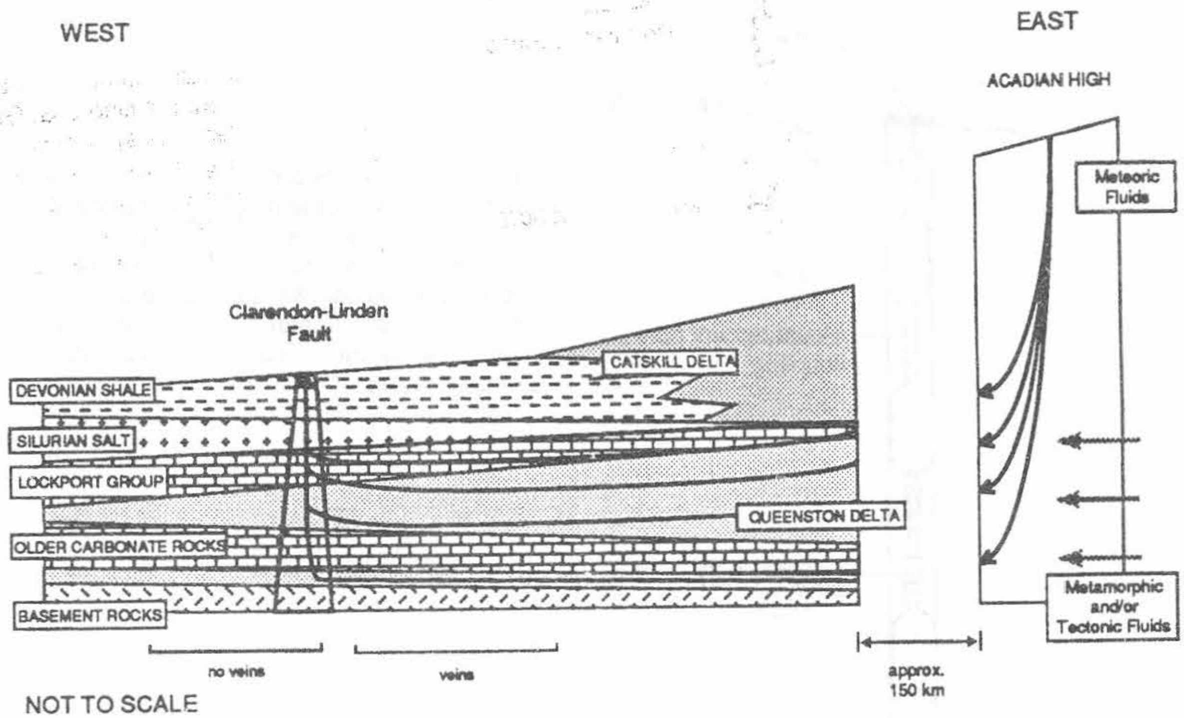


Figure 11. -- Conceptual pattern of fluid circulation during the late Paleozoic that may have resulted in propagation of the systematic veins. (Modified from Gross 1989).

in the Lockport Group must take into account the following observations: (1) the presence of veins east of the Clarendon-Linden Fault and their absence west of it; (2) the clockwise swing in vein orientation from east to west; (3) homogenization temperatures of vein fill which imply a circulation of fluids in excess of 115° C at the time of fracture propagation; and (4) a general increase in vein spacing from east to west. All of the above individual pieces of evidence point toward a highland to the east that provided the topographic head required to circulate fluid westward and to propagate veins in the Lockport Group.

Stress trajectories at the time of crack propagation can be inferred from the strike of a regional joint or vein set. S_H (maximum horizontal stress) trendlines for cross-fold joints in forelands may converge toward a major depocenter that emerges from the core of a fold-thrust belt. The eastward convergence of S_H trendlines drawn parallel to the strike of the veins is incompatible with S_H inferred from Alleghenian structures of the Appalachian Plateau and with other post-Paleozoic structures. However, east-west trendlines are compatible with tectonic features dated as Acadian.

A possible scenario of fluid circulation during the Acadian Orogeny is shown in Figure 11. As a result of the topographic high created during the Acadian Orogeny, fluid flow was downward into the Lockport Group from the east, and the fluids were then trapped beneath the overlying impermeable Salina Group salt units, which pinch out near Syracuse. The salt prevents fluids from circulating upward; therefore, high fluid pressures built up in the Lockport Group that resulted in vein propagation. As the fluids continued to circulate westward, they intersected the Clarendon-Linden Fault zone and were drained to the surface, which lowered fluid pressures immediately to the west of the fault. At Clarendon, the systematic veins disappear and remain absent until their reappearance at Niagara Falls. Vein spacing generally increases westward; the closest spacing is in Sodus and the widest is in Niagara Falls. This westward increase in vein spacing may reflect a westward decrease in fluid pressure.

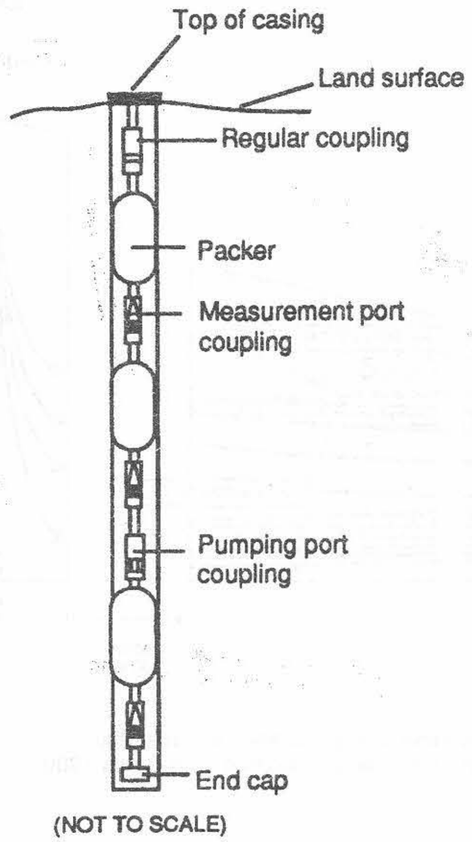


Figure 12. -- Multipiezometer system that was used to instrument the nine U.S. Geological Survey test holes. The system consists of down-hole casing, packers, regular couplings, and specialized couplings designed to function as measurement ports or pumping ports. The packers are arranged to isolate zones of interest.

The dominant fracture peak at the Clarendon Quarry (Fig. 6B) is oriented north-south, which corresponds to the north-south-trending Clarendon-Linden Fault zone. The overlying sedimentary cover may have been draped over the upthrown basement blocks, resulting in extension and the observed fracture distribution. Such pervasive fracturing would have provided a high permeability conduit through which fluids from the east could have drained.

HYDROGEOLOGY

by Dorothy H. Tepper, Richard M. Yager, and William M. Kappel

Although the Lockport Group is the principal aquifer in the Niagara area, it is not heavily pumped because the Niagara River provides the major public-water supply. Well yields in areas not affected by induced infiltration from the Niagara River typically range from 0.0006 m^3/s (cubic meters per second) to 0.0062 m^3/s but yields as high as 0.0601 m^3/s have been recorded. Some industrial wells near the Niagara River produce more than 0.1263 m^3/s as a result of induced infiltration from the river (Johnston 1964). Ground water in the Niagara Falls region generally flows southwestward from recharge areas near the escarpment toward the Niagara River, the primary discharge zone. However, the direction of flow has been altered by manmade structures such as the Falls Street Sewage Tunnel and the buried conduits system that transports water from the Niagara River to the reservoir at the Robert Moses Power Plant. A detailed study of the effects of the power plant on ground-water flow in the upper part of the Lockport Group is presented in Miller and Kappel (1987).

As part of the cooperative USGS/USEPA study of the hydrogeology of the Niagara Falls area, nine test holes (Fig. 1) were drilled to provide information on stratigraphy, fracture distribution, hydraulic head and other hydraulic characteristics, and ground-water geochemistry. Four of the holes were drilled to the Queenston Shale (top of the Ordovician System) and five were drilled to the Neagha Shale (Clinton Group of the Silurian System). A triple-tube wireline-coring technique was used to obtain continuous core from each test hole.

Multipiezometer Casing and Pressure Profiles

A multipiezometer system designed by Westbay Instruments Ltd.[3/] was used to instrument each of the nine test holes. This system, which will be discussed in detail at Stop 3, consists of downhole casing, packers, regular couplings, and specialized couplings designed to function as measurement ports or as pumping ports (Fig. 12). The packers on the casing string are used to isolate zones of interest. Each isolated zone contains a measurement port and may contain a pumping port. Specialized downhole tools are lowered into the central access pipe to activate these ports. Hydraulic head measurements and samples for water-quality analyses can be obtained through the measurement ports. The pumping ports are used to perform hydraulic tests.

At each site, an initial set of pressure measurements was taken soon after the packers were inflated to check for packer-seal integrity and to determine the initial pressure conditions in each isolated zone. Pressure distribution with depth was then measured bimonthly until equilibrium conditions were reached.

Profiles of dynamic pressure in relation to depth at the USGS Wheatfield (WF-1) test hole (Fig. 1) are shown in Figure 13A. "Zero" dynamic pressure represents hydrostatic pressure with depth. Any zone that plots to the left of the zero line is underpressured with respect to hydrostatic conditions, and any zone that plots to the right of the zero line is overpressured. As can be seen from the plot for November 1987, most zones below an altitude of 130 m were initially overpressured; this is attributed to the low permeability of the rock and to "packer squeeze" effects, which created temporarily high initial pressure readings. Recent downhole pressure conditions, which were stable from October 1988 to October 1989, are indicated on the October 1989 plot.

The relation between pressure and depth shown for the USGS Wheatfield (WF-1) test hole in Figure 13 is typical for the Niagara region. It shows an overpressured zone extending from the Rochester Shale down into the Power Glen Shale. An underpressured zone occurs below this, near the contact between the Queenston Shale and the Whirlpool Sandstone. The formation pressure profile in Figure 13B presents the same data as pore pressures in relation to depth. This plot shows that a constant pressure of about 1,520 kilopascals extends throughout the overpressured zone. This condition is typical in gas-bearing formations where the gas density is so small that the weight of the gas does not increase the pressure in the lower part of the reservoir. In contrast, the density of water is much greater, so that pressure generally increases linearly with depth in the Lockport Group (Fig. 13B). The presence of gas in the zones of overpressuring has been confirmed during drilling and by later sampling in these intervals.

An underpressured zone near the contact between the Queenston Shale and the Whirlpool Sandstone is observed in seven of the nine test holes instrumented with multipiezometer casing in the Niagara Falls area. Four of these seven test holes are within 3 km of the Niagara Gorge. Underpressuring is commonly encountered in partially depleted gas-bearing formations where the resaturation of pore spaces is restricted by low formation permeability and reduction of relative permeability due to the presence of two fluid phases. The gas-bearing formations in the Niagara Falls area probably have been depleted by production in both Canada and the United States and, in addition, gas may be leaking from formations that outcrop along the Niagara Gorge and the escarpment.

Use of Electromagnetic Techniques to Map High-Angle Fractures at the USGS Niagara (NI-1) Site

Fractures in the Lockport Group in the Niagara area are difficult to map because outcrops are limited to the Niagara Stone Quarry, the Niagara River Gorge (much of which is inaccessible), and a few areas along the Niagara Escarpment. In addition, glacial overburden, which averages 3.1 to 18.3 m thick, covers the bedrock surface and limits the use of aerial photographs and(or) satellite imagery for identification of fractures. Little information on high-angle fracture distribution can be obtained from cores from vertical test holes. However, ground-water flow data derived from aquifer tests and well-yield data, structure-contour maps of var-

3/ Use of brand and(or) firm names in this report is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey, the University of Rochester, or the Pennsylvania State University.

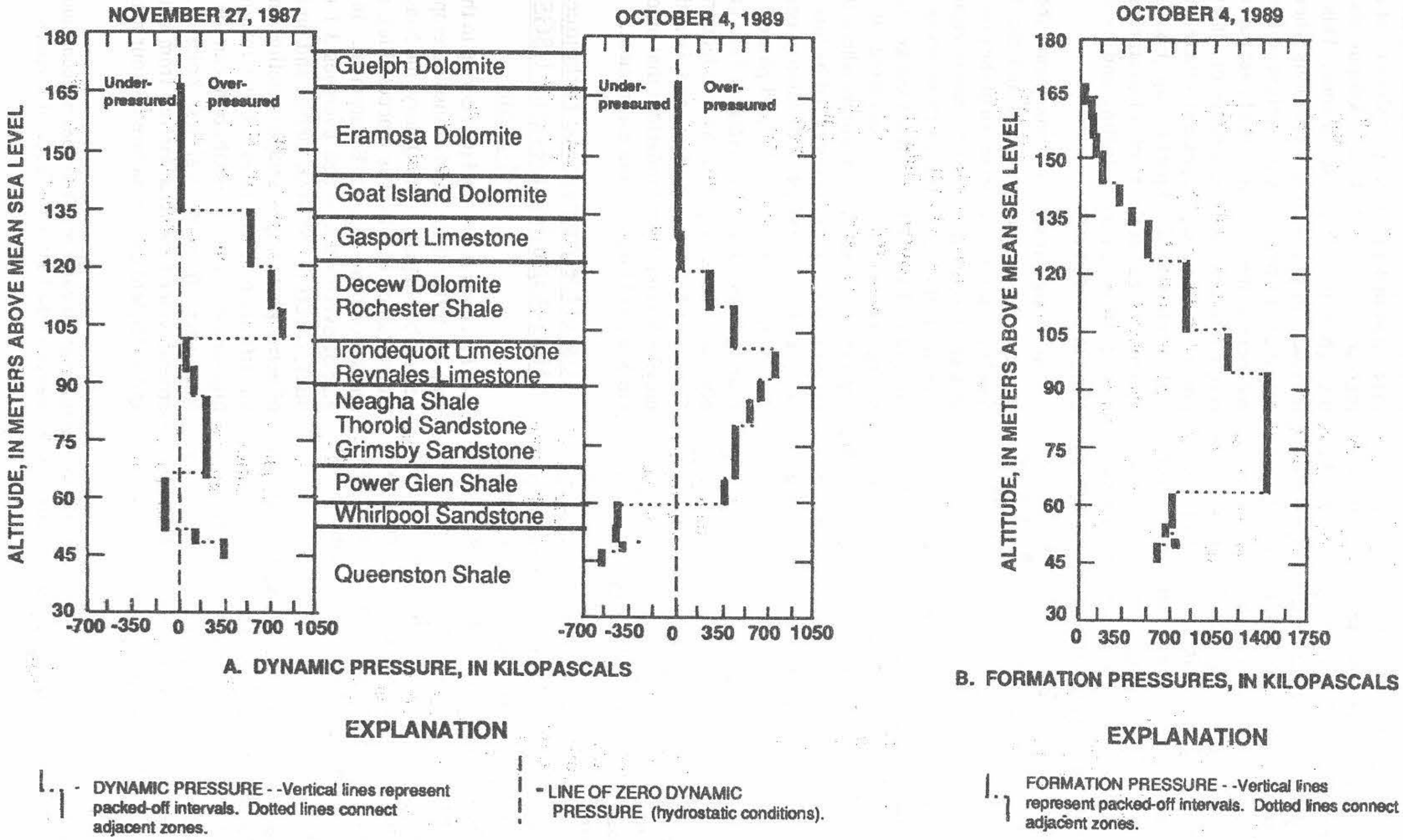


Figure 13. - - Profiles of dynamic pressure and formation pressure in relation to depth at the U. S. Geological Survey Wheatfield (WF-1) test hole. (Location shown in Fig. 1).

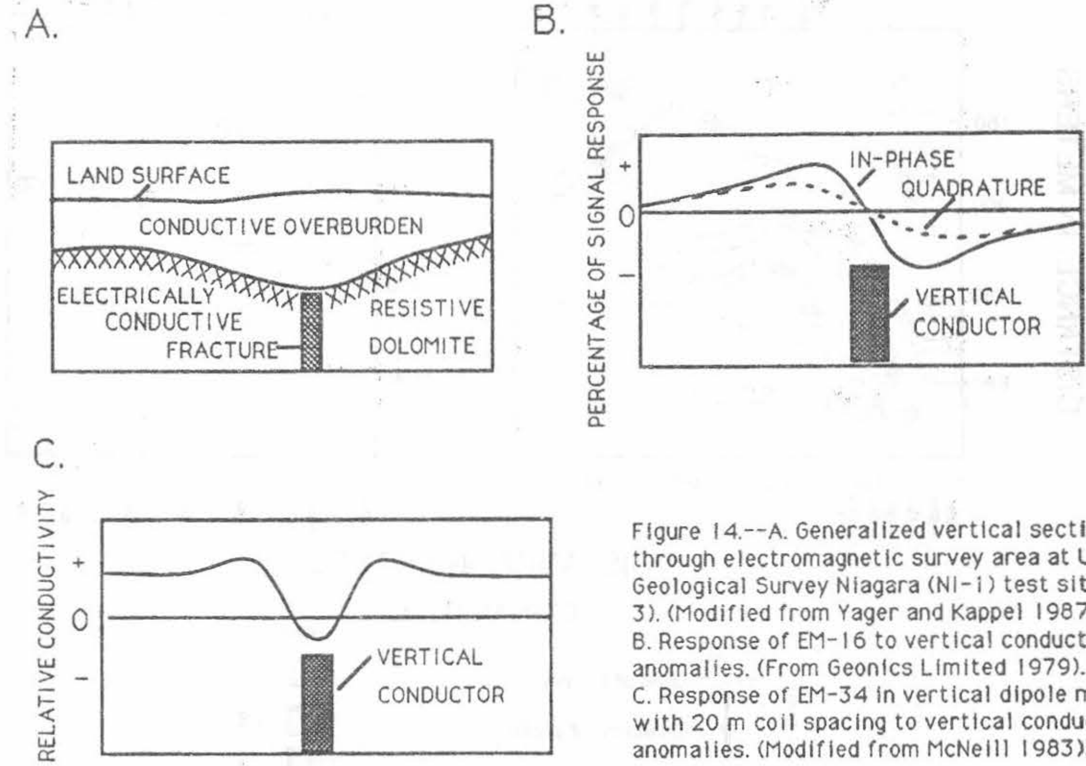


Figure 14.--A. Generalized vertical section through electromagnetic survey area at U.S. Geological Survey Niagara (NI-1) test site (stop 3). (Modified from Yager and Kappel 1987). B. Response of EM-16 to vertical conductive anomalies. (From Geonics Limited 1979). C. Response of EM-34 in vertical dipole mode with 20 m coil spacing to vertical conductive anomalies. (Modified from McNeill 1983).

ious stratigraphic horizons, and electromagnetic techniques can be used to delineate fracture trends and zones of intense fracturing.

The USGS Niagara (NI-1) site (Fig. 1) is unusual among the nine USGS test hole sites because the horizontal water-bearing zones appear to be well connected by high-angle fractures. As a result of these connections and a strong downward hydraulic gradient, freshwater (total dissolved solids < 3,000 mg/L (milligrams per liter)) is found deeper (34 m) at this site than at the others studied. Evidence for high-angle fractures in the Lockport Group at this site was first obtained through electromagnetic surveys using VLF (EM-16) and terrain conductivity (EM-34) equipment manufactured by Geonics Ltd. Reconnaissance surveys were performed rapidly with the EM-16, and areas of interest were then mapped in greater detail with the EM-34, which provides finer resolution. A detailed discussion of these techniques and the results of these surveys are presented in Yager and Kappel (1987).

Mapping electrically conductive high-angle fractures in the resistive dolomite (Fig. 14A) with these instruments is similar to mapping buried ore deposits or pipelines. The instruments will record a distinctive response (or anomaly) when crossing over a buried conductor (Fig. 14B,C). The extent and orientation of the vertical fractures can be mapped if measurements are taken along several parallel profiles.

Johnston (1964) hypothesized that a band of high transmissivity extends across the Lockport Group and is probably caused by an abundance of vertical joints or enlargement of the horizontal bedding planes. Additional evidence for the presence of this high-transmissivity zone is presented in Yager and Kappel (1987). The NI-1 site was chosen for study with electromagnetic techniques because it lies along this trend. The frequency of conductive anomalies was much higher at the NI-1 site than at another site 4.8 km east of this trend (Fig. 15).

Cross-Hole Hydraulic Testing Program

Cross-hole hydraulic tests were conducted at the NI-1 site to confirm the fracture trends identi-

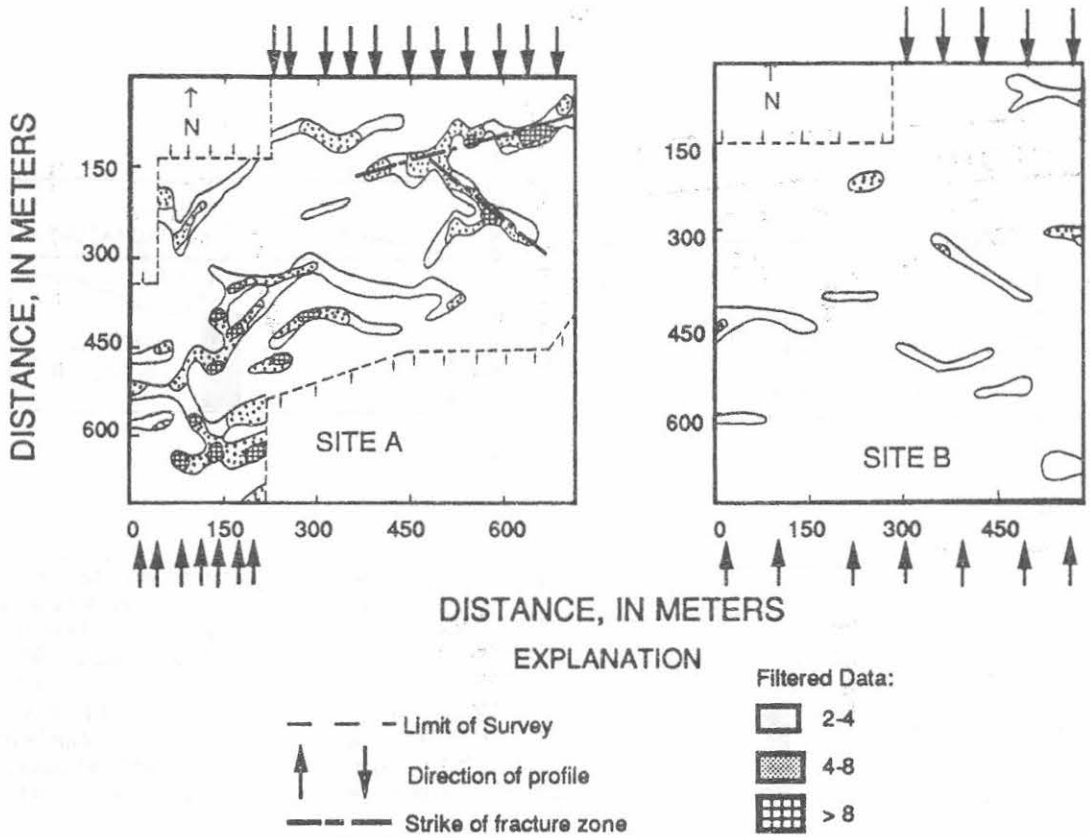


Figure 15.--Electromagnetic anomalies interpreted as generalized fracture zones at site A (NI-1 site) (Stop 3, Fig. 1) and at site B, 4.8 km east of site A. Interpreted from in-phase data measured by EM-16, filtered by method of Fraser (1969). (Modified from Yager and Kappel 1987).

fied with the electromagnetic equipment and to measure the vertical conductance between the horizontal fracture zones (Fig. 16A). Four additional test holes were sited to complete a 5-spot pattern (Fig. 16B) along or away from fractures mapped by the electromagnetic techniques. These test holes were cored and multipiezometer casing was then installed. Core obtained from the center hole contained a high-angle fracture, and information provided by an acoustic televiewer log revealed the strike of this fracture to correspond with a trend identified from the electromagnetic surveys. After coring was completed, the center hole was reamed to a 15.2-cm diameter and was left open to serve as a pumped hole. Drawdowns induced by pumping were measured in the pumped fracture zone and in two or three adjacent zones that were hydraulically isolated from each other by a removable packer string. "Mini-packer" strings allowed simultaneous measurement of four fracture zones in each test hole. As many as 20 isolated intervals were monitored in each test. Drawdowns were measured with vibrating-wire transducers manufactured by Geokon Inc, and were then recorded by a CR10 data logger built by Campbell Scientific Instruments.

Results of the cross-hole tests confirm the presence of vertical connections between horizontal fracture zones; however, these connections do not appear to extend to the weathered bedrock surface. The distribution of drawdown within some of the horizontal fracture zones is isotropic, as in a homogeneous porous medium. In other zones, the drawdown distributions display a high degree of heterogeneity, indicating pathways of high transmissivity within the fracture plane, perhaps as a result of increased dissolution at the intersections of horizontal and high-angle fractures. The directions of highest transmissivity correspond to orientations of high-angle fractures interpreted from the electromagnetic surveys.

ACKNOWLEDGMENTS

The authors would like to thank the owners of the Frontier Stone Quarry in Gasport and the Niagara Stone Quarry in Niagara for providing access to their respective quarries.

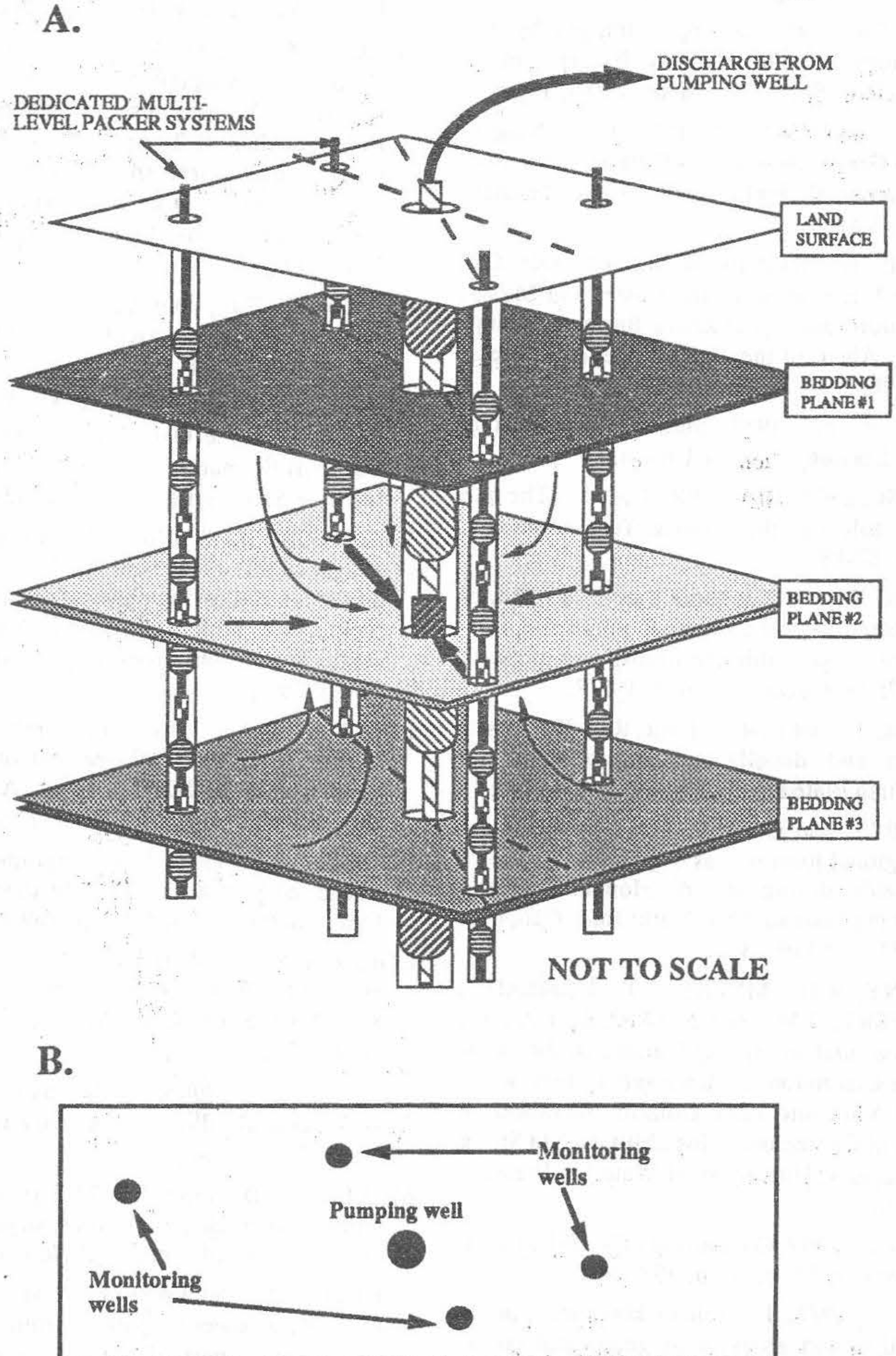


Figure 16. -- A. Isolation of bedding planes for cross-hole hydraulic testing with the multipiezometer system in each of the four monitoring wells. The pumping well is shown in the center. B. Plan view of the U. S. Geological Survey Niagara (NI-1) cross-hole hydraulic test site (Stop 3, Fig. 1).

REFERENCES CITED

BOLTON, T.E., 1957, Silurian stratigraphy and paleontology of the Niagara Escarpment in Ontario: Geol. Surv. Can. Memoir 289, 145 p.

BRETT,C.E., and CALKIN, P.E., 1987, Niagara Falls and Gorge, New York - Ontario: Geol. Soc. Am. Centennial Field Guide - Northeastern Sect., p. 97-105.

BRETT, C.E., GOODMAN, W.M., and LODUCA, S.T., 1989, Evolution of the Lower and Middle Silurian northern Appalachain Basin: Washington, D.C., Abstr. of the 28th Int. Geol. Congress, v. 1, p. 198.

CHEENEY, R. F. , 1983, Statistical methods in geology: London, Allen & Unwin, 169 p.

DAVIS, D.M., and ENGELDER, T., 1985, The role of salt in fold-and-thrust belts: Tectonophysics, v. 119, p. 67-88.

ENGELDER, T., 1982, Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America?: Tectonics, v. 1, p. 161-177.

ENGELDER, T., and ENGELDER, R., 1977, Fossil distortion and decollement tectonics of the Appalachian Plateau: Geology, v. 5, p. 457-460.

ENGELDER,T., and GEISER, P.A., 1980, On the use of regional joint sets as trajectories of paleostress fields during the development of the Appalachian Plateau, New York: Jour. Geophys. Res., v. 85, p. 6319-6341.

FAKUNDINY, R.H., MYERS, J.T., POMEROY, P.W., PFERD, J.W., and NOWACK, T.A., Jr., 1978, Structural instability features in the vicinity of the Clarendon-Linden Fault system, western New York and Lake Ontario: Advances in Analysis of Geotechnical Instabilities, SM Study no. 13, paper 4, University of Waterloo Press, p. 121-178.

FRASER, D.C., 1969, Contouring of VLF-EM data: Geophysics, v. 34, no. 6, p. 958 - 967.

GEISER, P.A., 1988, The role of kinematics in the construction and analysis of geological cross-sections in deformed terranes, in Mitra, G. and Wojtal, S., eds., Geometries and mechanisms of thrusting, with special reference to the Appalachians: Geol. Soc. Am. Special Paper 222, p. 47-76.

GEONICS LIMITED, 1979, Operating manual for EM-16 VLF-EM: Mississauga, Ontario, 78 p.

GROSS, M.R., 1989, Fractures in the Lockport Dolomite of western New York and southern Ontario: 1. Veins: constraints on late Paleozoic fluid circulation; 2. Asymmetric reentrants in the Niagara Escarpment: a case for neotectonic joints in North America [unpublished M.S. thesis]: University Park, Pennsylvania State Univ., 99 p.

HANCOCK, P.L., and ENGELDER, T., 1989, Neotectonic joints: Geol. Soc. Am. Bull., v. 101, p. 1197-1208.

HOWELL, B.F., and SANFORD, J.T., 1947, Trilobites from the Oak Orchard Member of the Lockport Formation of New York: Wagner Free Institute Science Bulletin 22, p. 33-39.

JOHNSTON, R.H., 1964, Ground water in the Niagara Falls area, New York, with emphasis on the water-bearing characteristics of the bedrock: New York State Conservation Department, Water Resources Commission Bulletin GW-53, 93 p.

LIBERTY, B.A., 1981, Structural geology, in Tesmer, I.H., ed., Colossal cataract--the geologic history of Niagara Falls: Albany, State University of New York Press, p. 57-62.

LODUCA, S.T., 1990, Medusaegraptus Mirabilis Ruedemann as a noncalcified dasyclad algae: Jour. Paleon., v. 64, no. 3, p. 469-474.

LODUCA, S.T., and BRETT, C.E., 1990a, Placement of the Wenlockian/Ludlovian boundary in New York State: Geol. Soc. Am. Abstr. with Prog., v. 22, no. 2, p. 31.

__________, 1990b, Placement of the Wenlockian/Ludlovian boundary in New York State: Lethaia, in press.

MCNEILL, J.D., 1983: EM 34-3 survey interpretation techniques: Mississauga, Ontario, Geonics Limited, Technical Note TN-8, 16 p.

MILLER, T.S., and KAPPEL, W.M., 1987, Effect of Niagara Power Project on ground-water flow in the upper part of the Lockport Dolomite, Niagara Falls area, New York: U.S. Geological Survey Water-Resources Investigations Report 86-4130, 31 p.

NORTH AMERICAN COMMISSION ON STRATIGRAPHIC NOMENCLATURE, 1983, North American Stratigraphic Code: Am. Assoc. Pet. Geol. Bull., v. 67, no. 5, p. 841-875.

RICKARD, L.V., 1975, Correlation of the Silurian and Devonian rocks in New York State: New York State Museum of Science Service Map and Chart Series no. 24, 16 p., 4 pl.

SBAR, M.L., and SYKES, L.R., 1973, Contemporary compressive stress and seismicity in eastern North America: an example of intra-plate tectonics: Geol. Soc. Am. Bull., v. 80, p. 1231-1264.

STRAW, A., 1968, Late Pleistocene glacial erosion along the Niagara Escarpment of southern Ontario: Geol. Soc. Amer. Bull., v. 79, p. 889 - 910.

TESMER, I.H., (ed.), 1981, Colossal cataract: the geologic history of Niagara Falls: Albany, State University of New York Press, 219 p.

YAGER, R.M., and KAPPEL, W.M., 1987, Detection and characterization of fractures and their relation to ground-water movement in the Lockport Dolomite, Niagara County, New York, in Khanbilvardi, R.M., and Fillos,J., eds., Pollution, risk assessment, and remediation in groundwater systems: Washington, D.C., Scientific Publications Company, p. 149-195.

ZENGER, D.H., 1965, Stratigraphy of the Lockport Formation (Middle Silurian) in New York State: New York State Museum and Science Service Bulletin no. 404, 210 p.

ZOBACK, M.L., and ZOBACK,M.D., 1980, State of stress in the conterminous United States: Jour. Geophys. Res., v. 85, p. 6113-6156.

ROAD LOG FOR LOCKPORT GROUP FIELD TRIP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
		From Fredonia State College, proceed north on I-90 toward Buffalo. Follow I-90 east from Buffalo and get off at Exit 49 for Depew, Rt. 78
0.0		At traffic light at toll booth, turn left (north) onto Rt. 78 and follow into Lockport. Enter City of Lockport at 14.1 miles.
15.3	15.3	At intersection of Rt. 78, Rt. 93, and Rt. 31, turn right (east) onto Rt. 31. At the time this road log was compiled (March 1990), there was construction (including detours) on Route 31 in Lockport. Follow Rt. 31 through Lockport and continue eastward toward Gasport.
18.1	2.8	At junction of Rt. 77, veer left and **continue to follow Rt. 31 east.**
20.7	2.6	Enter Town of Gasport. Stay on Route 31 east, which is called Rochester Street in town.
23.4	2.7	Pass under conveyor belt that crosses Rt. 31 and turn left immediately after the conveyor belt into the Frontier Stone parking lot. Wait in driveway for further instructions. **THIS IS STOP #1.**
		Return to vehicles and follow Rt. 31 west back into Lockport.
29.8	6.4	Enter City of Lockport.
31.6	1.8	At intersection of Rt. 31 and Rt. 78, turn left (south) onto Rt. 78. **THIS IS THE LUNCH STOP.** There are a number of restaurants and fast-food chains on this road. Reconvene at the specified time at the parking lot of the Niagara County Office Building at 59 Park Avenue. To get there, return to the intersection of Rt. 78 and Rt. 31 and turn left (west) onto Route 31. Go 0.1 miles on Route 31, and turn right onto Hawley St. just after passing a small park. Go 1 block on Hawley St. to the stop sign, then turn right onto Park Avenue and immediately turn left into the parking lot of the Niagara County Office Building.
31.7	0.1	Turn left out of the parking lot of the Niagara County Office Building. Turn right at the yield sign onto Route 31 west (around 300 feet from the parking lot).
33.3	1.6	Route 93 goes off to the right but stay on Route 31, which is now called Saunders Settlement Road.
35.1	1.8	Enter Town of Cambria.
41.4	6.3	Pass entrance to Niagara County Community College on the right.

41.8	0.4	Enter Town of Lewiston.
43.5	1.7	Pass Niagara-Wheatfield Middle School and senior high school on the left.
44.2	0.7	Turn left onto Walmore Road.
45.8	1.6	At stop sign, turn right onto Lockport Road.
46.3	0.5	Pass entrance to Niagara Falls Air Force Base on left.
47.3	1.0	At fork in road, continue straight (right side of fork), staying on Lockport Road.
48.1	0.8	Turn right onto Miller Road; after 0.1 miles, turn right onto Quarry Road. Follow Quarry Road through the gate, then turn left and park at Niagara Stone Division parking parking lot. **THIS IS STOP #2.**
		Return to vehicles and proceed to end of Quarry Road.
49.0	0.9	At end of Quarry Road, turn left onto Miller Road.
49.1	0.1	Turn left at stop sign onto Lockport Road.
49.8	0.7	Turn left at intersection of Lockport Road and Packard Road.
50.3	0.5	Turn right (first right) onto Tuscarora Road.
50.6	0.3	Turn right onto dirt road at fence line and follow lead vehicle to **STOP #3.** *Beware of broken glass and pipes sticking up in the road!*

END OF FIELD TRIP

PALEOECOLOGY AND SEDIMENTOLOGY OF MIDDLE DEVONIAN SHORT-TERM SEALEVEL FLUCTUATIONS RECORDED WITHIN THE WANAKAH SHALE MEMBER OF WESTERN NEW YORK

KEITH B. MILLER
Department of Earth Science
Clemson University
Clemson, South Carolina 29634

INTRODUCTION

The fine-grained sediments of the Middle Devonian Hamilton Group, deposited within the northern tongue of the storm-dominated Appalachian Basin, have proven to be an excellent testing ground for many paleoecologic and stratigraphic concepts as demonstrated by many past NYSGA guidebook articles. The extensive stream and lake shore outcrop exposures, relatively undisturbed "layer cake" stratigraphy, and well-preserved benthic faunas of the Hamilton Group in western and central New York are ideal for detailed sedimentologic, paleontologic, and taphonomic studies. Such studies have begun to uncover the wealth of information contained within the superficially homogeneous mudrocks and thin limestones of the Hamilton Group.

Fossil taphonomy is an important, if not vital, source of data for the interpretation of depositional environments, particularly within fine-grained rocks lacking obvious sedimentary structures. The relative effects of the post-mortem processes of reorientation, disarticulation, fragmentation, corrasion and encrustation upon various skeletal materials yields considerable information on rates of burial, exposure time, and degree of physical reworking (Brett & Baird, 1986a). Comparison of these taphonomic features from bed to bed or outcrop to outcrop enables subtle environmental gradients to be discerned (Norris, 1986).

As attention has been drawn to the details of individual shell beds, their storm-event origins have increasingly been recognized in shallow shelf settings. Using individual storm event-beds as building blocks, a temporal hierarchy of physical processes and biological responses can be reconstructed, from the scale of single beds to the ordering of facies sequences within an entire depositional basin (Aigner, 1984, 1985). Internally, shell beds record the complex short-term interactions of episodic storm-generated physical disturbance, substrate consistency, and benthic community composition (Miller, *et al.*, 1988). Onshore-offshore gradients in the relative frequency, sedimentologic character and fossil taphonomy of storm beds enables relative water depths to be determined (Brett *et al.*, 1986a). In addition, widely traceable storm event-beds or packages of event beds possessing distinctive taphonomic and faunal signatures can be used as isochronous markers for very high-resolution correlation. At a yet greater temporal scale, cyclic patterns in the taphonomy and faunal composition of shell beds reflect sea-level fluctuations at several temporal scales (Savarese *et al.*, 1986; Brett and Baird, 1986b; Miller, in press). These cycles afford a potentially valuable means for correlation since the positions of maximum regression and transgression provide basinwide isochronous markers. The character of these cycles also provides a very important clue to the dynamics of deposition within the northern Appalachian Basin during the Middle Devonian.

STRATIGRAPHIC AND PALEOGEOGRAPHIC SETTING

The general basin paleogeography and tectonic setting of the northern Appalachian Basin during the deposition of the Middle Devonian (Givetian) Hamilton Group is now fairly well known (Fig. 1). The Appalachian Basin was characterized by a western carbonate margin and an eastern siliciclastic margin for most of the Paleozoic (Cotter, 1983; Read, 1980; Walker *et al.*, 1983; Brett and Baird, 1985). During the Givetian the western New York area was occupied by a muddy carbonate ramp with a very gentle paleoslope (<< 1°) dipping south to southeastward toward the basin center. The basin axis trended approximately northeast to southwest and was centered in the Seneca and Cayuga Lake area, with a somewhat steeper clastic slope to the east (Baird, 1981; Brett *et al.*, 1986b). A silty and sandy platform then extended eastward to a prograding "deltaic" shoreline in the vicinity of present day Albany (Dennison, 1985). This clastic progradation represents the first influx of sediment from the initial pulses of the Acadian Orogeny.

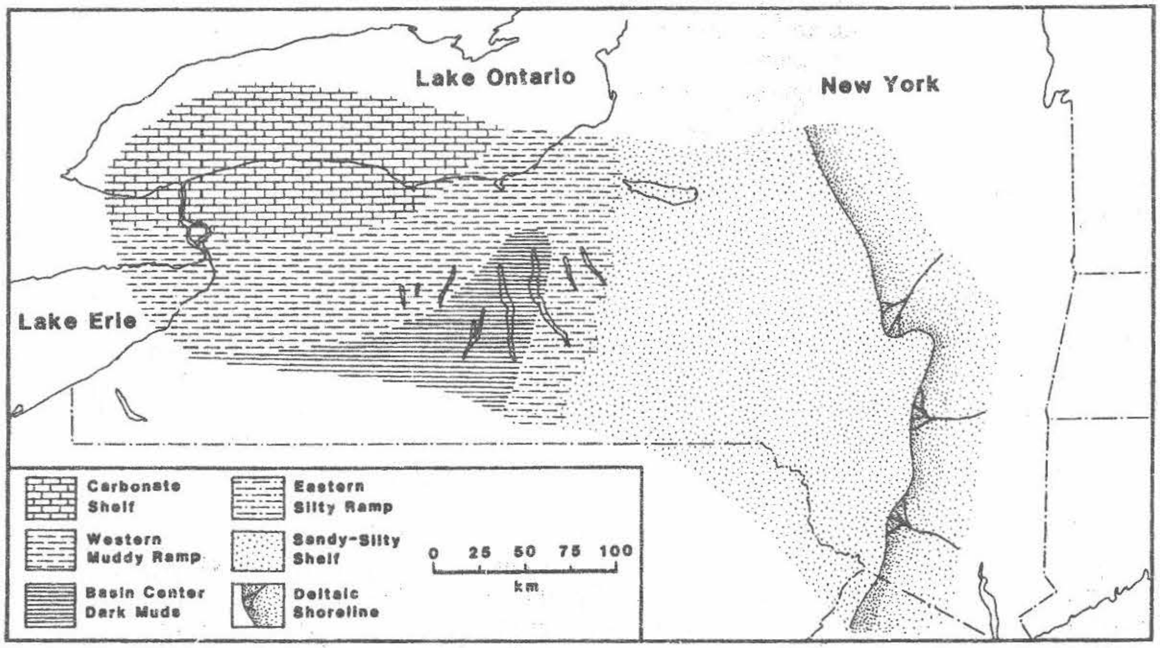


Figure 1. Paleogeographic reconstruction of the northern Appalachian Basin during deposition of the Wanakah shales. Note the nearly E-W depositional strike of the muddy carbonate ramp in western New York State.

The generally fine-grained sediments of the Hamilton Group have been subdivided at the formational level on the basis of very widespread carbonate units such as the Stafford, Centerfield, and Tichenor Limestones (see Fig. 2). Member and submember boundaries are likewise marked by widely persistent thin shelly carbonate beds or diastemic horizons which contain taphonomic and sedimentologic evidence of significant stratigraphic condensation. These diagnostic condensed beds, therefore, delineate clearly defined, bounded units of regional extent within which the detailed correlation of numerous shell-rich horizons can be attempted. One such bounded interval, and the focus of the following discussion, is the lower Wanakah Member of the Ludlowville Formation in western New York, recently named the Darien Center Submember (Brett *et al.*, 1986b).

DESCRIPTION OF DARIEN CENTER SUBMEMBER

Lithologic and Faunal Description of Internal Units

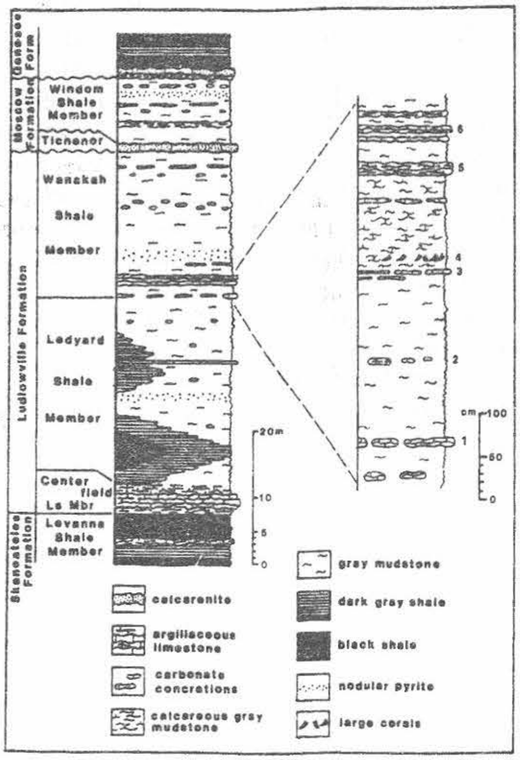
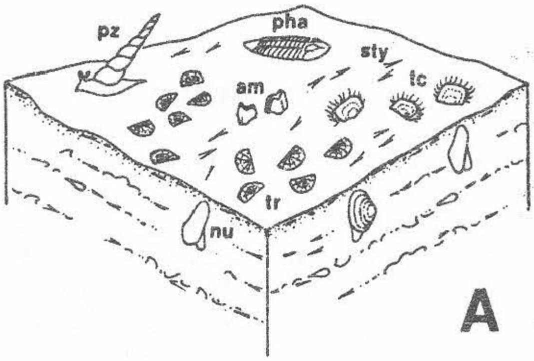


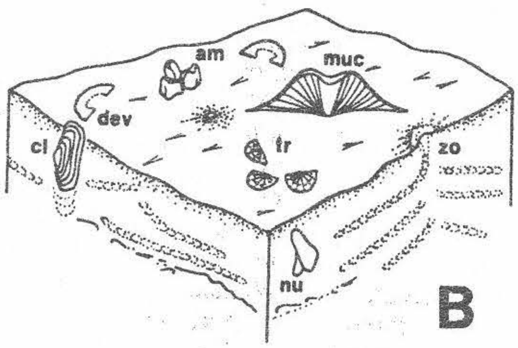
Figure 2. General stratigraphic column for the Ludlowville and Moscow Formations of the Hamilton Group in Erie County, western New York. The Darien Center Submember of the Wanakah Shale Member, which is the focus of this paper, is shown in detail to the right. Beds widely traceable throughout western and central New York designated by numbers as follows: 1) Mt. Vernon Bed marking base of Wanakah Member, 2) Girdle Road Bed, 3) Lakeview (*Nautilus*) Bed, 4) Darien Coral Bed, 5) Murder Creek (trilobite) Bed, and 6) Bidwell (trilobite) Bed. Descriptions of beds can be found in Kloc (1983) and Miller (in press). Figure modified from Brett *et al.*,1986b.

The Wanakah Member of the Ludlowville Formation in western and central New York State is ideal for studying cyclic and event processes. It is a thin sequence of gray calcareous shales and thin argillaceous limestones and concretionary layers. Fossils are concentrated in centimeter-thick layers separated by poorly fossiliferous to barren shales. Of the numerous persistent shell layers and fossil-rich calcareous beds, several have long been recognized as widely traceable stratigraphic markers with characteristic faunas (Grabau, 1899; Cooper, 1930). Cooper (p.225) defined the base of the Wanakah by one such bed, the "*Strophalosia* Bed" (renamed the Mt. Vernon Bed by Kloc, 1983), which can be recognized from Lake Erie as far as the Owasco Valley (Brett *et al.*, 1986b). The sedimentary package discussed herein extends upward from this prominent key bed to include the widespread lower Wanakah "Trilobite Beds" of Grabau (1899, p.40-41). Two of these beds, the Murder Creek and Bidwell Beds (Kloc, 1983), are traceable from Lake Erie at least to the Seneca Valley, a distance of over 150 km. The well-defined, thin (3-5 meter), stratigraphic interval bounded by the Mt. Vernon and Bidwell Beds has been named the Darien Center Submember (Brett *et al.*, 1986b; Miller, in press) (see Fig. 2). The designated type section for the Darien Center Submember is the stream bank exposure along Elevenmile Creek within Darien Lakes State Park near the town of Darien Center.

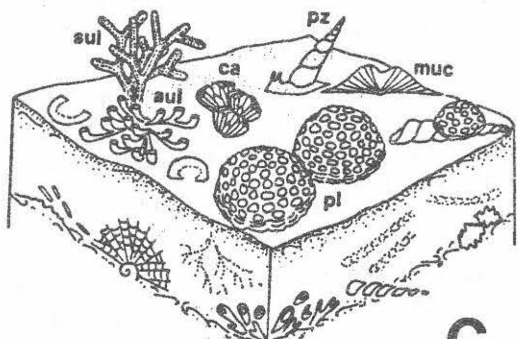
The **Mt. Vernon Bed** is a thin (~10-15 cm thick) argillaceous limestone typically containing the small productid brachiopod *Truncalosia truncata*, previously referred to as *Strophalosia truncata*. In addition to *Truncalosia*, the Mt. Vernon bed is characterized by an abundance of



A

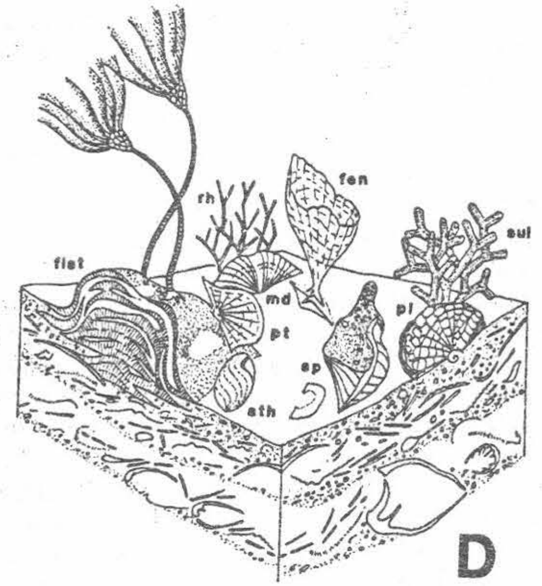


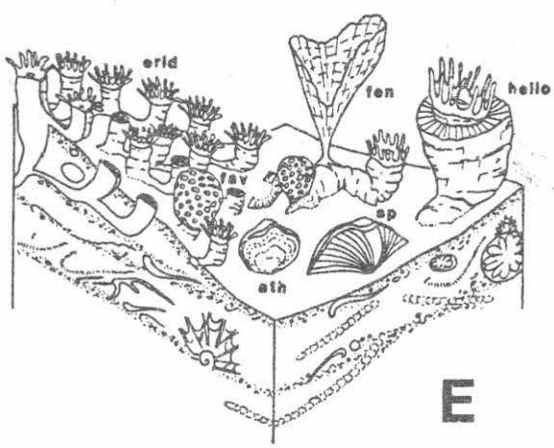
B



C

Figure 3. Schematic block diagrams illustrating characteristic benthic faunal assemblages and event bed appearance of key-bed-bounded intervals. A) Diminutive Brachiopod Assemblage of the Mt. Vernon Bed. Closely spaced shelly pavements are typical of this bed and the shales below. B) Very low diversity Chonetid-nuculid Assemblage characteristic of the interval between the Stolle Road and Girdle Road Beds. *Zoophycos* spreiten are abundant and shell pavements are rare. C) *Camarotoechia* Assemblage between the Girdle Road and Lakeview Beds characterized by abundant *Pleurodictyum* corals. Pavements of chonetids and winnowed crinoidal layers are colonized by moderately diverse epifaunal assemblage. Fossil abreviations are as follows: (am) *Ambocoelia*, (tc) *Truncalosia*, (tr) *Tropidoleptus*, (dev) *Devonochonetes*, (muc) *Mucrospirifer*, (ca) *Camarotoechia*, (sty) *Styliolina*, (pz) *Palaeozygopleura*, (nu) nuculid, (cl) *Cypricardella*, (aul) *Aulocystis*, (pl) *Pleurodictyum*, (sul) *Sulcoretepora*, (pha) *Phacops*, (zo) *Zoophycos*..





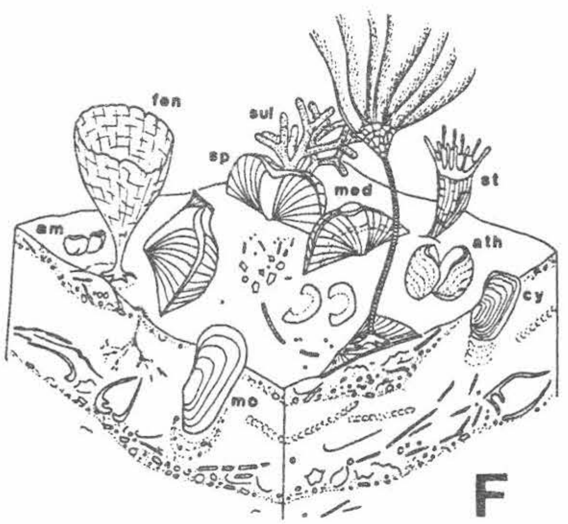
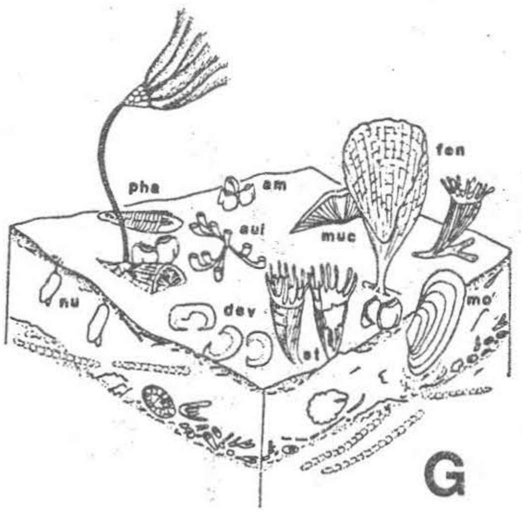
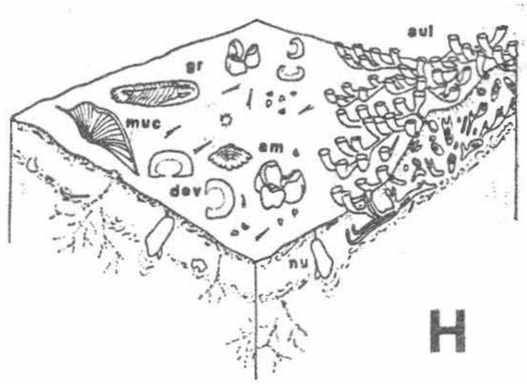


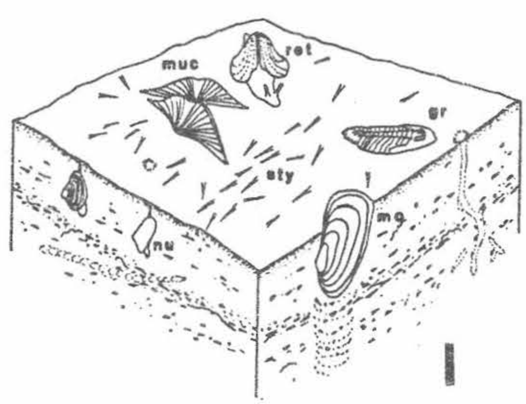
Figure 3 (cont'd). D) Diverse Brachiopod Assemblage characteristic of the Lakeview Bed and the central portion of the Darien Center Submember. Note dominance and variety of bryozoans, and the large spiriferid brachiopods. Epifaunal pectinid and pterioid bivalves are particularly common. Event beds are amalgamated into complex beds recording multiple winnowing and colonization events. E) Heliophyllum Assemblage of the Darien Coral Bed. Note *Eridophyllum* colony with multiple internal trapped mud layers, and geniculated *Heliophyllum* coral developed in response to toppling and regrowth. Small favositid coral colonies grow on dead coralites. F) Diverse Brachiopod Assemblage of the Fargo Bed with life-position clusters of large spiriferid brachiopods. Event beds show sharp, scoured bases with finely comminuted shell debris and crinoid ossicles, and smothered tops with unfragmented, articulated, and often life-position fossils. Fossil abbreviations are as follows: (am) *Ambocoelia*, (dev) *Devonochonetes*, (md) *Mediospirifer*, (ath) *Athyris*, (sp) *Spinocyrtia*, (cy) *Cypricardinia*, (mo) *Modiomorpha*, (pt) *Pterinopectin*, (pl) *Pleurodictyum*, (st) *Stereolasma*, (helio) *Heliophyllum*, (erid) *Eridophyllum*, (fav) *Favosites*, (sul) *Sulcoretepora*, (rh) *Rhombipora*, (fen) fenestrate bryozoan, (fist) fistuliporoid bryozoan.



G



H



I

Figure 3 (cont-d). G) *Athyris* Assemblage characteristic of the interval around the Murder Creek Bed. Moderately diverse assemblage of small, usually clustered, brachiopods and auloporid corals with associated cryptostome bryozoans and crinoids. *Phacops* and *Greenops* trilobites are common and frequently occur as enrolled specimens. Note thin, locally discontinuous, fossil horizons with *in situ* articulated fossils. H) *Ambocoelia*-chonetid Assemblage with small ubiquitous brachiopods and mats of auloporid corals. Thin, closely spaced bedding plane concentrations of largely disarticulated skeletal material with fine *Chondrites* burrows. I) Chonetid-nuculid Assemblage of the Bidwell Bed with low-diversity fauna and high concentration of pelagic styliolinids. Fossil abreviations are as follows: (am) *Ambocoelia*, (dev) *Devonochonetes*, (muc) *Mucrospirifer*, (sty) *Styliolina*, (ret) *Retispira*, (nu) nuculid, (mo) *Modiomorpha*, (aul) *Aulocystis*, (st) *Stereolasma*, (fen) fenestrate bryozoan, (gr) *Greenops*, (pha) *Phacops*.

pelagic styliolinids together with orthoconic nautiloid cephalopods, the gastropods *Palaeozygopleura hamiltoniae* and *Retispira leda*, and the unusually small brachiopods *Ambocoelia* cf. *nana*, *Devonochonetes scitulus*, and minute individuals of *Tropidoleptus carinatus* (see Fig. 3A). This faunal assemblage is equivalent to the "Diminutive Brachiopod" assemblage defined by Brett and others (1986, in press). This fauna probably reflects dysaerobic conditions at the sea floor.

Overlying the Mt. Vernon Bed is a peculiar undulatory concretionary horizon which is particularly well displayed along Buffalo Creek, and will be discussed in detail later. The large ellipsoidal concretions of this bed mark a wavy surface which rises over irregularly-spaced lenticular bodies of nearly barren blue-gray shale, but between them becomes welded to the top of the Mt. Vernon Bed. This bed will be hereafter referred to as the **Stolle Road Bed** for its bedding plane exposure on the floor of Buffalo Creek near Stolle Road. Where it is welded onto the Mt. Vernon Bed, the Stolle Road Bed is characterized by dense accumulations of tiny *Ambocoelia* cf. *nana* brachiopods and styliolinids, together with orthoconic nautiloids, palaeozygopleurid gastropods, diminutive individuals of *Devonochonetes* and *Tropidoleptus*, and *Mucrospirifer mucronatus*. An unusual feature of these *Ambocoelia* -rich portions of the bed is the occurrence of 10 cm-wide branching concretionary bodies up to a meter in length (Miller,1986). Resembling large horizontal burrow structures, these concretions have typically high angles of branching which range up to 90°. Their randomness of orientation, which excludes an interpretation as diagenetically enhanced gutter casts, and their relatively uniform width would also seem to suggest a biological rather than physical origin.

As the Stolle Road Bed is traced up the sides of the barren shale lenses, it thins and progressively loses many of its faunal components and becomes increasingly pyrite rich. The tiny ambocoeliid brachiopods are the first to drop out as the shell bed rises from the Mt. Vernon Bed. The diminutive *Tropidoleptus* brachiopods also rapidly decline in number, and *Mucrospirifer* and *Devonochonetes* become the dominant brachiopods within a thin styliolinid hash. Near the top of the lenses, this shelly styliolinid-rich horizon is replaced by an indistinct interval of pyrite nodules and pyritic burrow tubes. The ellipsoidal carbonate concretions which outline the barren shale lenses at the outcrop tend to occur several centimeters below the undulatory shelly and pyritic layer. These concretions appear to be nearly barren of fossils except in the few places where the shelly layer passes through their tops.

Above the undulatory Stolle Road Bed, shell beds are essentially flat-lying. The first traceable shelly horizon is a widespread bed containing a moderately diverse fauna. Recognizable throughout western New York, this shell bed is here named the **Girdle Road Bed** for its exposure along the banks of Buffalo Creek next to the Girdle Road bridge. It marks a pronounced sedimentologic and faunal change from fossil-poor gray shales below to fossiliferous gray calcareous shales above. *Zoophycos* spreiten are particularly common within the fossil-poor shales. Laterally restricted fossil pavements are present, however, and within these pavements, and scattered widely throughout the shale, is a very low diversity faunal assemblage including *Mucrospirifer*, *Devonochonetes*, *Styliolina*, and the trilobite *Phacops rana* (see Fig. 3B). This assemblage is equivalent to the "Chonetid-nuculid" Assemblage defined by Brett and others (1986, in press).

The Girdle Road Bed marks the first occurrence of the discoid tabulate coral *Pleurodictyum americanum* above the Mt. Vernon Bed. Commonly occurring within this distinctive bed are the corals *Stereolasma* and *Aulocystis*, the brachiopods *Athyris*, *Spinocyrtia*, *Protoleptostrophia*, and *Megastrophia*, the trilobites *Phacops* and *Greenops*, as well as crinoids and ramose cryptostome bryozoans. The interval above this bed is equivalent to the

"*Pleurodictyum* beds" recognized by early workers (Grabau 1899, p.58; Cooper,1930, p.225), and has a slightly less diverse fauna than the Girdle Road Bed. In addition to *Pleurodictyum,* the most common faunal elements of this interval are auloporid and stereolasmatid corals, the brachiopods *Devonochonetes*, *Ambocoelia*, and *Camarotoechia*, and nuculid bivalves (see Fig. 3C). The secondarily occupied shells of *Palaeozygopleura* were the preferred attachment sites for *Pleurodictyum* (Brett and Cottrell, 1982), and this gastropod is invariably present, often in abundance. This assemblage is here designated the "*Camarotoechia* Assemblage," and is equivalent to part of the "Mucrospirifer-chonetid" Assemblage of Brett and others (1986, in press). Shell beds within this interval range from thin shelly pavements to beds with scoured bases and basal winnowed layers of crinoid ossicles and shell hash. Within many beds, small individuals of *Devonochonetes* form a basal layer upon which a more diverse fauna was developed. Auloporid corals commonly occcur in thickets or mounds built on shell pavements.

The next prominent concretionary horizon above the Girdle Road Bed overprints a complex amalgamated (multi-event) shell bed at the base of a very fossiliferous interval of gray calcareous shale. I have suggested (Miller, in press) that this bed, named the "*Nautilus* Bed" by Grabau (1899), be renamed the **Lakeview Bed** for its excellent exposure along the Lake Erie shore near the town of Lakeview. The fauna of this bed and the overlying shell beds is highly diverse. Brachiopods include *Mucrospirifer*, *Mediospirifer*, *Spinocyrtia*, *Athyris*, *Rhipidomella*, *Devonochonetes* and *Megastrophia*. The large *Spinocyrtia* brachiopods typically are heavily encrusted with auloporid corals and trepostome bryozoans. The semi-infaunal bivalves *Modiomorpha*, *Cypricardella*, and *Cypricardinia* are common elements, along with the pectins *Pterinopectin* and *Pseudoviculopectin*, and the pterioid *Actinopteria*. To the trilobites *Phacops* and *Greenops* is added the large phacopid, *Dipleura*. A very rich and diverse bryozoan assemblage is present, including fistuliporoids, branching and massive trepostomes, fenestrates, and cryptostomes. Hemispherical fistuliporoid mounds up to 20 cm or more in diameter are particularly characteristic of the shales overlying the Lakeview Bed. Crinoid ossicles are ubiquitous and are typically the primary component of the basal winnowed lags of the shell beds. The *Pleurodictyum* corals are replaced in abundance by *Stereolasma*, and auloporids remain a major component of the fauna. This assemblage, illustrated in Fig. 3D, is equivalent to the "*Pseudoatrypa* (Diverse Brachiopod)" Assemblage of Brett and others (1986, in press).

Within this highly fossiliferous interval, individual fossil layers typically consist of finely comminuted shell debris and disarticulated, often abraded, crinoid ossicles at the base, overlain by diverse epifaunal assemblages with abundant bryozoans. Sharp scoured bases, which commonly have elongate fossil-hash-filled prods similar to gutter casts, cut into underlying shales and may locally intersect older fossiliferous layers. Fossils on the upper bed surfaces are commonly *in situ* and often in life position. Complete articulated crinoids occur within immediately overlying shales, and may be found still anchored to the shell bed below. The significance of this preservational contrast between the bases and tops of shell beds is discussed in detail elswhere (Parsons *et al.*, 1986; Miller *et al.*, 1988).

At several localities in western New York a coral bed is present just above the Lakeview Bed. It varies from a bed of scattered large solitary corals to a nearly continuous 15 cm-thick biostrome of solitary and branching colonial forms. Kloc (1983) has named this thin coral-rich unit the **Darien Coral Bed** for exposures near the town of Darien where the biostromal bed is best developed. The large solitary rugosan *Heliophyllum halli* predominates in the beds of scattered corals, while the branching colonial rugosan *Eridophyllum* dominates the biostromal beds. Other elements of the coral fauna include the solitary rugosans *Cystiphylloides* and

Stereolasma, the branching tabulate *Trachypora*, as well as *Pleurodictyum* and several species of *Favosites*. Associated brachiopods include *Tropidoleptus*, *Mucrospirifer*, *Mediospirifer*, *Spinocyrtia*, *Athyris*, *Devonochonetes*, and *Megastrophia*. Crinoids and bryozoans, particularly fenestrates, continue to be important elements of the fauna. This bed, illustrated in Fig. 3E, represents the "*Pentamerella-Heliophyllum*" Assemblage of Brett and others (1986, in press). Solitary corals typically show geniculated and rejuvenated coralla, and broad conical forms are often found in inverted, calyx-down positions. Branching coral colonies show multiple internal levels of burial, corrosion, and rejuvenation. Corroded coral fragments are common, especially in the upper portion of the coral bed.

The diverse fauna characteristic of the Lakeview Bed continues upward above the Darien Coral Bed to a thin persistent calcareous bed called the **Fargo Bed** by Kloc (1983). The Fargo Bed and the shales immediately above and below are characterized by the presence of large articulated brachiopods, particularly *Athyris*, *Mediospirifer*, and *Spinocyrtia*, which often occur in small clusters of life-position individuals (see Fig. 3F). Within this interval, fossil layers with sharp scoured bases are separated by 4-6 cm of fossil-poor shale containing bivalves such as *Modiomorpha* and *Cypricardinia*. The fauna of the shell beds above the Darien Coral Bed differs little from that of the Lakeview Bed.

Above the Fargo Bed is a widely traceable prominent diagenetic carbonate bed containing abundant trilobites, which are often preserved in enrolled positions. This lowest of Grabau's (1899) "Trilobite Beds" has since been called the **Murder Creek Bed** (Kloc, 1983). Discontinuous thin fossil layers with a moderately diverse epifaunal assemblage are separated by fossil-poor shales a few centimeters thick. The highly abundant tiny spiriferid *Ambocoelia*, and the strophomenids *Pholidostrophia*, *Douvillina*, and *Devonochonetes* are the characteristic brachiopods of this interval. Though absent within the Murder Creek Bed, abundant *Athyris* occurs in a widely traceable horizon a few centimeters above. The trilobites *Phacops* and *Greenops*, and the corals *Aulocystis* and *Stereolasma* are all abundant components of the fossil horizons. Modiomorpha and nuculid bivalves occur as common elements of the low-diversity fauna between the shelly layers. This assemblage (see Fig. 3G) is essentially equivalent to the "*Athyris*" Assemblage of Brett and others (1986, in press).

Significant faunal and taphonomic changes occur above the *Athyris*-rich horizon mentioned above. Here, closely spaced horizons or pavements of finely fragmented brachiopod debris, ostracods, and styliolinids are characteristic. Within these horizons are patches of disarticulated *Greenops* trilobites, and disarticulated valves of *Mucrospirifer*, *Devonochonetes*, and *Ambocoelia*. The auloporid coral *Aulocystis* is common, and occurs as small *in situ* patches, or at some localities as irregular mats up to 3 cm thick. A few small *Stereolasma* are associated with these mats, some of which still have apices attached to the auloporids. Iluustrated in Fig. 3H, this assemblage is the "*Ambocoelia*-chonetid" Assemblage of Brett and others (1986, in press).

Lastly, the top of the Darien Center Submember is marked by a very prominent carbonate bed which has been named the **Bidwell Bed** (Kloc, 1983). This highly condensed bed contains thin graded styliolinid layers with disarticulated *Greenops* trilobites and *Mucrospirifer* and *Devonochonetes* brachiopods. The nuculid bivalve *Paleoneilo*, semi-infaunal bivalves *Modiomorpha* and *Modiella*, and the archeogastropod *Retispira* are characteristic of this bed. Burrowing is extensive, with *Zoophycus* spreiten, vertical burrows up to 1.5 cm in diameter, and fine *Chondrites* burrows. The fauna of this bed (see Fig. 3I) would fall within the "Chonetid- nuculid" Assemblage.

The key shell beds described above, and the intervals which they bound, are remarkable in their persistence over western New York for lateral distances exceeding 100 km. The vertical sequence of distinctive key beds and the vertical faunal and taphonomic patterns have proven to be so consistent that they can be used in a predictive fashion from outcrop to outcrop, thus greatly improving confidence in the accuracy of field correlation. This lateral uniformity is especially significant in light of the very thin stratigraphic intervals involved, with individual key-bed-bounded units only decimeters thick. The correlation of these internal units of the Darien Center Submember among ten measured and sampled sections is shown in Fig. 4, and illustrates the amazing "layer cake" stratigraphic pattern present.

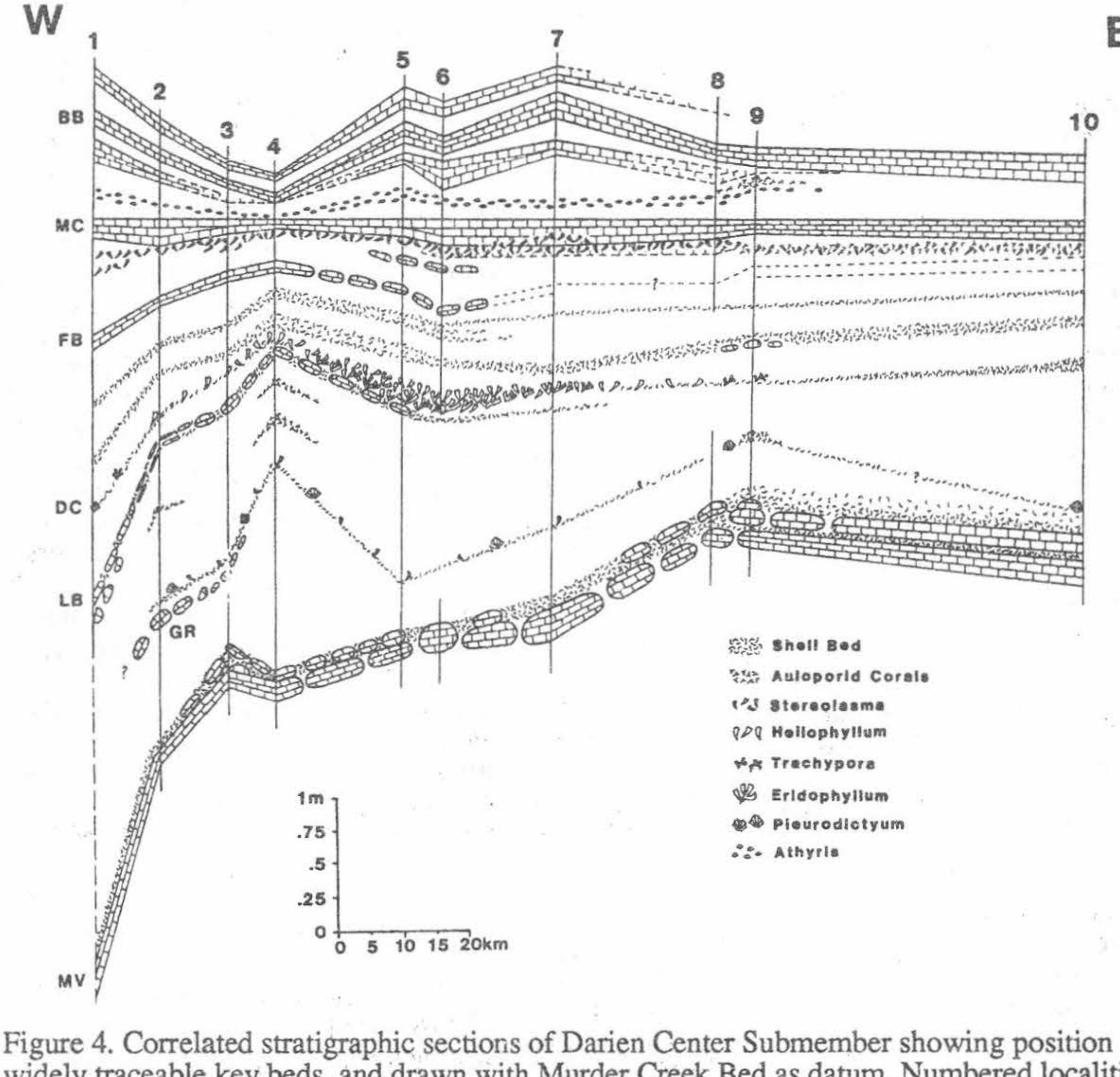


Figure 4. Correlated stratigraphic sections of Darien Center Submember showing position of widely traceable key beds, and drawn with Murder Creek Bed as datum. Numbered localities are as follows: 1) Lake Erie Shore, 2) Rush Creek, 3) Cazenovia Creek, 4) Buffalo Creek, 5) Elevenmile Creek, 6) Murder Creek, 7) Francis Road railroad cut, 8) Salt (Bidwell) Creek, 9) Wheeler Gully, and 10) Hopewell Gully.

The numerous closely spaced key beds function as very high-resolution correlation lines across western New York. Not only do they provide important marker beds within the thin, laterally uniform western facies, but they have also been subsequently traced across the substantial thickness and lithofacies transitions of the basin axis in the Seneca and Cayuga Lake valleys (see Fig. 1). Individual key beds can be traced from the western calcareous shales into the silty mudstones of the basin axis, and even into the hummocky cross-stratified siltstone and fine sandstone on the eastern side of the basin axis (Brett *et al.*, 1986b). Such facies independence suggests that these beds, each composed of several superimposed and amalgamated event horizons, may represent time lines and thus have important chronostratigraphic utility.

Several lines of argument can be made for the isochroneity of the key beds described above (a detailed discussion of these arguments can be found in Miller, in press). 1) When traced across the basin they can be seen to clearly cross-cut facies. The close parallel of key beds and facies units within western New York is an artifact of the near coincidence of the trend of the ourcrop belt and depositional strike. However, across the basin axis to the east, they show substantial internal lithologic, taphonomic, and faunal change (Miller, 1988). The genesis of these beds was therefore independent of specific depositional environments. 2) Most key beds are overprinted by very early diagenetic limestones and concretionary layers which may record chemical responses to specific sedimentation events. The rapid burial of long-term shelly accumulations, and the living communities colonizing them, by storm-redeposited mud could catalyze widespread concretion formation (Berner, 1968). 3) The commonly scoured basal surfaces of individual shell beds, and the internal winnowed surfaces within complex shell beds, represent the great majority of time, while the shales between record geological instants of storm mud deposition (see discussion in Parsons et al., 1986; Miller, in press). The basal portions of shell beds typically consist of corroded, abraded, and highly fragmented shell material, which may include a resistant residue of durable skeletal elements like crinoid ossicles. This contrasts with the excellent preservation observed on shell bed tops. The articulated, multi-element skeletons of crinoids and trilobites, and life position brachiopods and corals, are persuasive evidence of rapid burial. The small-scale stratigraphic discontinuities represented by the winnowed and eroded surfaces associated with these shell beds are not different in principle from unconformities used as isochrons in sequence stratigraphy.

The ability to trace very thin marker beds over hundreds of kilometers is not unique to the northern Appalachian Basin. The time-parallel nature of such beds has also been previously demonstrated by other workers for other stratigraphic settings. For example, Hattin (1985) has successfully correlated closely spaced, thin chalky limestone beds within calcareous shales of the Upper Cretaceous over large areas of the western interior of the United States. A wide range of event bed types are potentially available as the basis for very high-resolution correlation, and Kauffman (1988) has synthesized these into a system of chronostratigraphy called "High-Resolution Event Stratigraphy" (HIRES).

DESCRIPTION AND INTERPRETATION OF CYCLIC PATTERNS

Proximality Spectrum of Storm-Event Beds

The vertical trends in shell bed taphonomy exhibited by the Darien Center Submember, and by other stratigraphic intervals within the Hamilton Group, can be used to reconstruct a bathymetric gradient of storm-generated event beds. As seen from the descriptions above, shell beds typically have a basal winnowed, or non-depositional, shell lag colonized by an *in situ* well-preserved epifaunal assemblage and subsequently buried by a rapidly deposited layer of poorly fossiliferous mud. The Darien Center Submember displays a nearly symmetric cyclic pattern in the taphonomy and internal structure of these storm-event beds. Within the central portion of the submember, between the Lakeview and Fargo Beds, amalgamated or multi-event shell beds containing several superimposed fossil layers are typical. These beds were probably deposited well within storm wave-base. As the degree of basal scour and erosion decreases away from this central portion of the interval, shell beds lose their amalgamated character and become thinner until reduced to shell pavements only a single shell layer thick. The shales with shell pavements lacking any evidence of basal scour are presumed to have been deposited below the effective base of even the most severe storms. Mud burial layers overlying the shell beds are up to 10 centimeters thick near the center of this cyclic sequence, and decrease to one centimeter or less over the shelly pavements near the boundaries of the cycle.

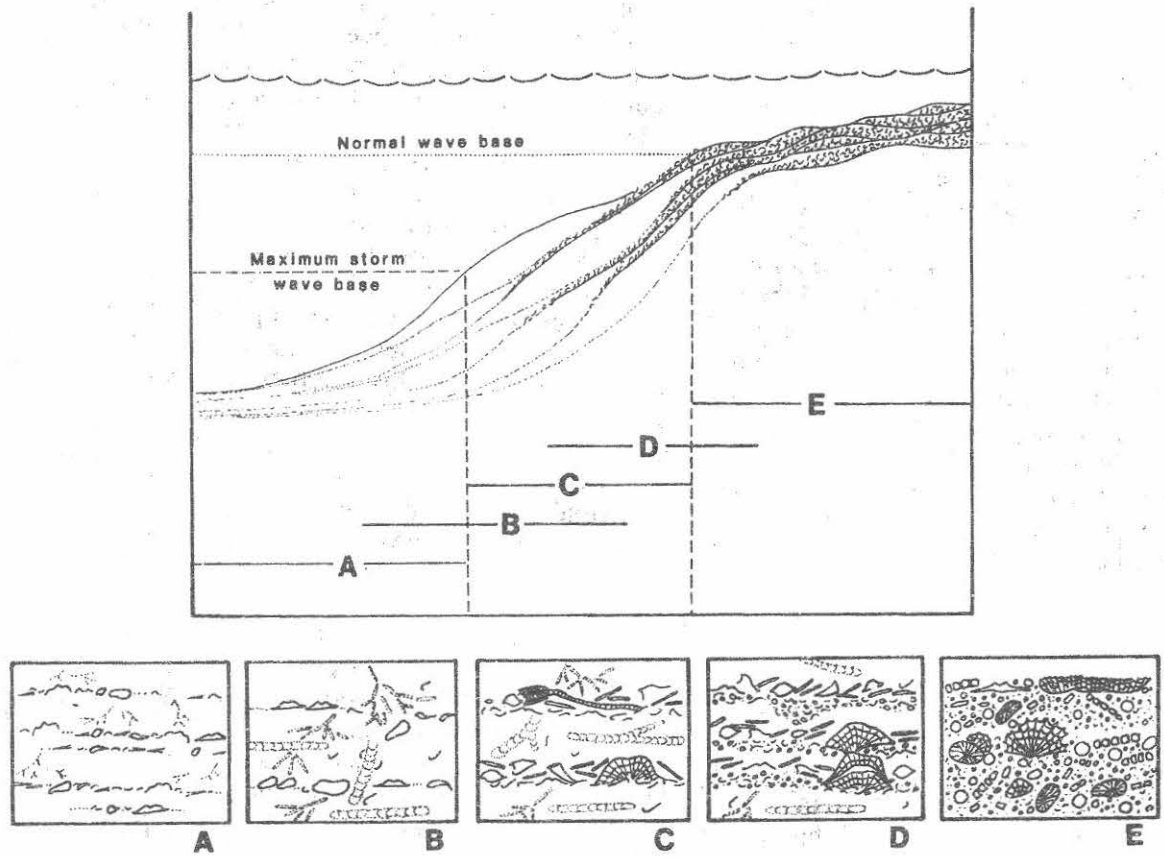


Figure 5. Diagram of storm depositional model showing successive event deposits produced by storms of varying intensity. Lettered bars show expected depth ranges of different shell bed types: A) non-depositional surfaces smothered by thin distal mud layers, B) colonized soft-bottom surfaces buried by upslope winnowed muds, C) colonized winnowed pavements buried by thick layers of redeposited muds, D) rewinnowed and amalgamated shell beds smothered by thick mud layers, and E) winnowed crinoidal grainstones with a few subtle internal burial horizons. Reprinted from Miller and others (1988) with permission of S.E.P.M.

A very simplified model of storm deposition proposed by Miller and others (1988) can explain these observed trends. During any given storm, winnowing would occur above storm wave-base, with mud redeposited in a basinward-thinning wedge below storm wave-base. The absolute depth at which winnowing and mud deposition occur would vary with storm intensity. Averaged over many storms, mud layers would tend to be thickest somewhat below the reach of the "average" storm and taper into deeper water. Conversely, the frequency and intensity of storm winnowing would tend to increase into shallower water environments with the increasing occurrence of shell bed amalgamation and multiple reworking. No mud layers would be expected above fairweather wave-base where wave agitation would be an essentially continuous process. A proximality spectrum of shell beds is therefore generated which can be used to estimate depth relative to fairweather and storm wave-base (see Fig. 5). Though similar to other models of storm-bed sedimentation (Aigner and Reineck, 1982; Aigner, 1985; Driese, 1988), this model focuses on the more distal spectrum of storm-related deposits common within fine-grained offshore marine sediments. The more distal range of events beds preserved in the Darien Center Submember is also ideal for the excellent *in situ* preservation of a highly diverse, benthic fauna.

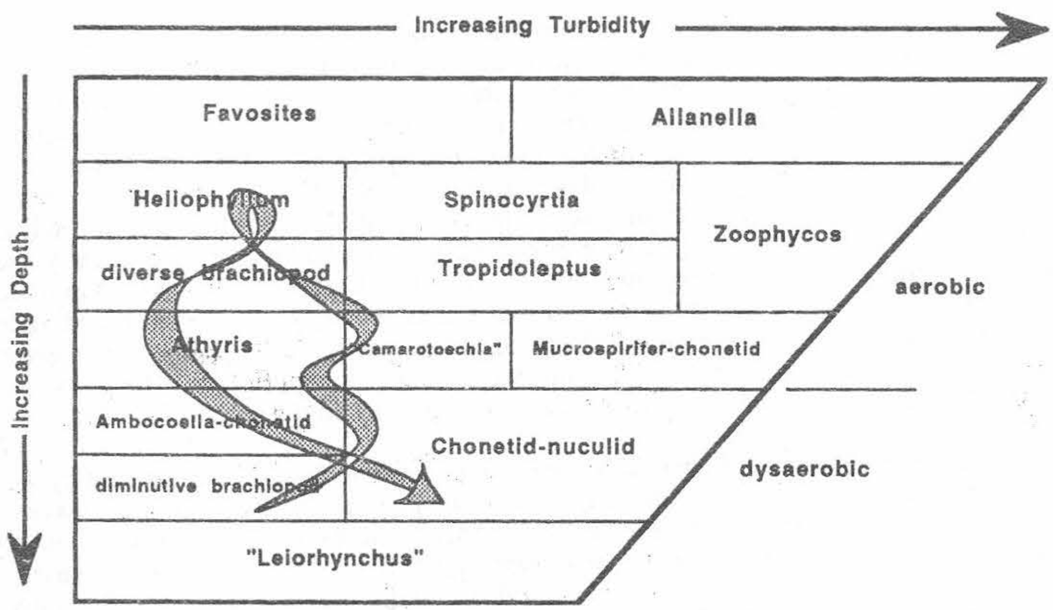


Figure 6. Paleoecologic model for Hamilton benthic assemblages showing relationships to inferred gradients of depth and turbidity and/or sedimentation rate (modified from Brett *et al.*, 1986). Arrowed path shows vertical sequence of assemblages for Elevenmile Creek locality.

Depth Gradients of Fossil Assemblages

The symmetrical cyclic pattern in storm bed taphonomy within the Darien Center Submember is matched quite closely by the vertical sequence of associated fossil assemblages. These assemblages are interpreted as representing bathymetric facies belts which migrated laterally in response to a regressive-transgressive cycle. The relative ordering of these assemblages is part of the consistent pattern of recurrent and laterally intergrading assemblages of the Hamilton Group which have been recognized by many workers, and related to gradients of depth and turbidity (see references in Brett *et al.*, 1986b). The depth and turbidity limits for the recurrent faunal assemblages of the Hamilton Group are summarized in Fig. 6. Recent work on the density and diversity of microborings within Hamilton assemblages (Vogel *et al.*, 1987) has supported these proposed depth and turbidity relationships. Further, the presence of algal microborings in all facies indicates a range of depths entirely within the photic zone.

The close parallel of taphonomic and faunal trends raises interesting questions regarding the extent and nature of the interaction between the dynamics of the physical environment and the benthic fauna. The depth ranges of benthic taxa may have been controlled to a large degree by the frequency and intensity of storm activity (Miller *et al.*, 1988). As different regimes of physical disturbance migrated up and down slope with changing sea level, they would have been tracked by their resident disturbance-adapted benthic faunas, resulting in the observed vertical sequence of fossil assemblages.

Reconstruction of Depth and Eustatic Sea-Level Curves

The depth relationships of the benthic assemblages of the Hamilton can be used to reconstruct depth curves. Since the composition of the benthic fauna appears to have been highly sensitive to depth, detailed centimeter by centimeter sampling of stratigraphic intervals can provide the basis for depth curves of very fine temporal, and bathymetric, resolution. The diagram of Fig. 6 can be used to visually display the changing depths (and turbidities) recorded by the vertical sequence of benthic assemblages at a given locality. A "path" can be drawn on this diagram tracing out the sequence of assemblages encountered from the base of the Darien Center Submember to its top. An example of one such path, shown on Fig. 6, clearly shows the regressive-transgressive nature of the interval. Note that the path is a closed loop indicating a return to origin conditions, and the completion of a sedimentary cycle. From this diagram, it also is apparent that the regressive half of the cycle was deposited under somewhat more turbid conditions than the transgressive half. This latter conclusion is supported by the more condensed character and higher fossil density of the upper portion of the cycle.

When the presence/absence faunal data is examined in more detail, an additional cyclic pattern can be recognized which is superimposed on the general regressive/transgressive cycle (Miller, in press). This subcyclicity is reflected in sharp changes in faunal diversity closely associated with the widely traceable key beds (see Figs. 11, 12, 13 in road log). Sudden diversity increases are associated with the Girdle Road and Lakeview Beds, and diversity decreases are associated with the Murder Creek and Bidwell Beds. Highest faunal diversities are recorded within amalgamated shell beds between the Darien Coral Bed and the Fargo Bed. These diversity fluctuations, together with the symmetrical pattern of depth-controlled faunal assemblages, have been used to reconstruct a depth curve for the Darien Center Submember (Fig. 7). This depth curve is used to infer an eustatic sea-level curve with two superimposed cycles of different periodicites. The apparent association of key beds with the subcycle boundaries is significant in providing another line of argument for the isochroneity of these widely traceable beds.

At least five subcycles are recognized within the thin Darien Center interval, each bounded by key beds. Though not shown on Fig. 7, a sixth subcycle is probably recorded by the interval between the Mt. Vernon Bed and the undulatory Stolle Road Bed. The prominent fossil horizons occurring at the boundaries of these cycles appear to be condensed beds and probably record periods of reduced sediment accumulation associated with increased sediment bypass and/or reduced sediment supply. The apparent asymmetry of the subcycles is very interesting and significant. Those subcycles within the regressive phase of the larger cycle have condensed regressive portions, and those within the transgressive phase have condensed transgressive portions. The superimposition of a hierarchy of eustatic cycles would be expected to generate just such accentuated regressions and transgressions. As a result, deepening upward subcycles are produced during regression, and shallowing upward subcycles are produced during transgression.

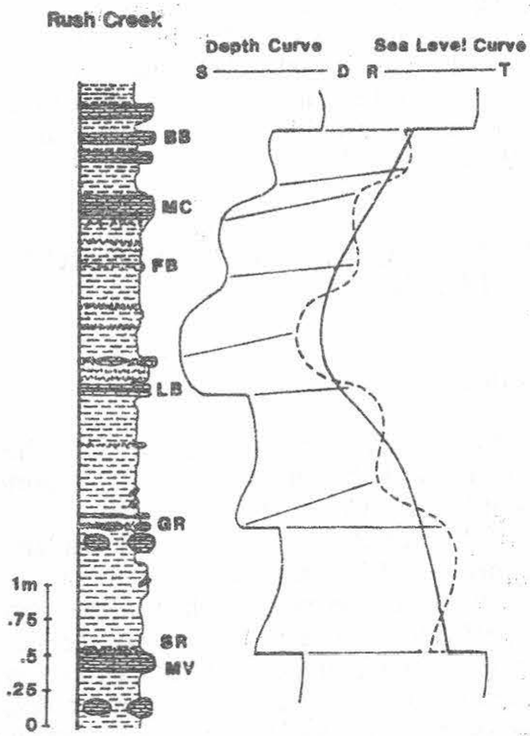


Figure 7. Stratigraphic column from Rush Creek with interpreted depth and sea-level curves for Darien Center Submember based on taphonomic and faunal patterns. Solid sea-level curve represents 5th order regressive-transgressive cycle and dashed line indicates superimposed 6th order subcycles. The Stolle Road Bed (SR), which immediately overlies the Mt. Vernon Bed (MV) at the base of the interval, represents an additional complete subcycle not shown on the curve (see Miller,1988, ch.3). The Girdle Road Bed (GR) is a subtle, but widespread, key bed within the regressive half-cycle.

The subcycles recognized within the calcareous shales and thin limestones of western New York have been correlated across the basin axis of the northern Appalachian Basin into the siltstones and hummocky cross-stratified sandstones of the eastern clastic ramp (Miller, 1988). The regressive subcycles are progressively lost on the eastern margin of the basin axis, and the transgressive subcycles become prominent coarsening-upward clastic cycles (Brett *et al.* 1986b). Specifically, the subcycles bounded by the Lakeview and Fargo Beds and by the Fargo and Murder Creek Beds can each be correlated with 5 to 7 meter-thick coarsening-upward cycles to the east. Likewise, the interval between the Murder Creek and Bidwell Beds appears to correlate with a thinner 1 meter coarsening-upward cycle. The traceability of these cycles across significant thickness and facies changes argues for a basinwide eustatic causal mechanism.

The absolute duration of the cycles recorded by faunal and taphonomic data are very difficult to estimate (Miller, in press). This is in part due to the absence of high-resolution biostratigraphic markers and datable horizons such as bentonites. The time scale for deposition of the entire Darien Center Submember is much smaller than that resolvable with traditional biostratigraphy, with the complete regressive-transgressive cycle representing only a fraction of single conodont and ammonoid zones. The most severe difficulty, however, is that surfaces rather than sediment record the vast majority of geologic time (Ager, 1981; Dott, 1983). If the marker beds were found to represent more time than the cyclic intervals which they bound, the durations of those cycles would be seriously overestimated. Nonetheless, crude order of magnitude approximations are possible by subdividing the Hamilton Group using a recognized hierarchy of sedimentary cycles. The Hamilton Group as a whole includes the uppermost Eifelian and much of the Givetian

(Klapper, 1981), a duration of about 6 million years. It has been divided into three transgressive-regressive cycles by Johnson and others (1985), which they believe are correlative over Euramerica. The Darien Center cycle, in turn is one of 14 or 15 yet smaller cycles within the Hamilton Group (Brett and Baird, 1986b). Simple division yields time estimates for these cycles of 300 to 400 thousand years, a value roughly comparable to midcontinent cyclothems, or to 5th order T-R cycles as defined by Busch and Rollins (1984) and Busch and West (1987). The six subcycles likely present within the Darien Center Submember would therefore represent perhaps 50 thousand years, a period at the short end of the time range estimated for punctuated aggradational cycles (Goodwin and Anderson, 1985) and 6th order T-R cycles (Busch and West, 1987). It must be emphasized that these time estimates are only a general first approximation, and merely provide an upper limit to the duration of these cyclic patterns (see Algeo and Wilkinson, 1988).

FORMATION OF AN UNDULATORY DISCONTINUITY SURFACE DURING REGRESSIVE PHASE OF CYCLE

Description of Undulatory Surface

The undulatory nature of the Stolle Road Bed has already been briefly discussed. The irregular nature of the concretionary horizon associated with this shell bed was first reported by McCollum (1981) for exposures along the banks of Buffalo Creek. However, this concretionary horizon was misindentified as the Mt. Vernon Bed, and interpreted as a late diagenetic artifact. As a result, McCollum abandoned the Mt. Vernon Bed as the basal boundary of the Wanakah Member in preference to the first appearance of the tabulate coral *Pleurodictyum*, the definition utilized by Cooper (1930) for localities to the east of the Genesee Valley. More recent detailed study, however, has recognized that a traceable shelly and pyritic horizon immediately overlies the undulating concretion layer, and that the Mt. Vernon Bed itself remains nearly horizontal and marks the base of lenticular bodies of nearly barren blue-gray shale. The Mt. Vernon Bed can be traced as a persistent shell bed even though the diagenetic concretionary limestone with which it is usually associated becomes discontinuous or is completely lost below the barren shale lenses. Its diagnostic fauna makes it an easily recognized marker bed, and it therefore should be retained as the basal boundary of the Wanakah Shale Member according to Cooper's (1930) original definition.

The repeated splaying and coalescence of the Mt. Vernon and Stolle Road Beds can be seen in the sketch of the stream bank exposure shown in Fig. 8. The appearance is one of a series of irregularly spaced swells on an otherwise flat surface. The swells differ substantially in amplitude, varying from a few tens of centimeters to nearly a meter and a half in height, and similarly range from 10 to over 60 meters in width. Also recognizable from Fig. 8 is the slight but consistent asymmetry of the swells, with the western margins noticeably steeper than the eastern margins. At Pond Brook, a nearby tributary to Buffalo Creek, some indication of the orientation of these structures can be obtained. At that locality, a N12,E trend was measured for a single swell exposed on both banks of the stream. This NNE-SSW orientation is significant, bcause it represents a nearly onshore-offshore direction based on the reconstructed basin paleogeography (see Fig. 1).

The detailed relationships between concretionary horizons and bedding surfaces for two swells are shown in Fig. 9. This diagram illustrates the important characteristics of the undulatory horizon and of the entire regressive half of the Darien Center cycle. Three distinct

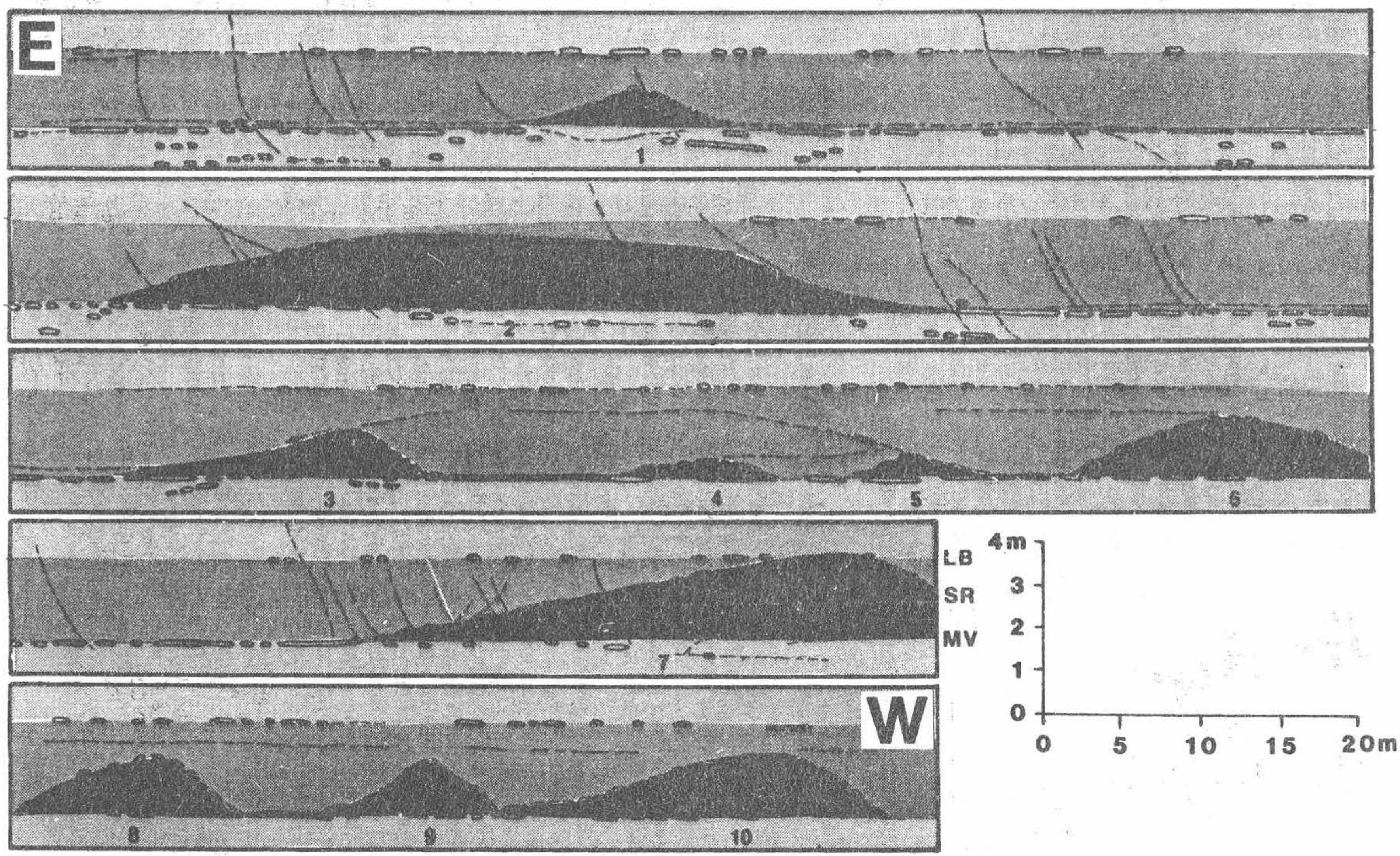


Figure 8. Sketch based on photographs of nearly 500-meter-long section of Buffalo Creek cut bank showing the outcrop appearance of the swells (highlighted with dark shading) and their variable spacing and amplitude. Individual swells are numbered. Vertical scale has three-fold exaggeration, and the scale shown is approximate. Labelled beds are: MV) Mt. Vernon Bed, SR) Stolle Road Bed, and LB) Lakeview Bed.

intervals can be distinguished: 1) the nearly barren shales below the Stolle Road Bed, 2) the fossil-poor shales filling the broad "channels" between swells, and 3) the closely-spaced fossil horizons overlying the Girdle Road Bed.

Below the Stolle Road Bed, the lenses of hard blue-gray shale tend to stand out somewhat on the stream bank exposures. As mentioned earlier, the ellipsoidal carbonate concretions which conveniently outline the swells in outcrop tend to occur several centimeters below the undulating shelly and pyritic layer of the Stolle Road Bed. Within larger swells, ellipsoidal concretions, as well as thin discontinuous carbonate layers, appear to define other internal undulatory horizons which may either be concordant or discordant with the contours of the swells. These indistinctly defined horizons are only infrequently associated with discernable fossil layers. Between the Stolle Road and Girdle Road Beds are poorly fossiliferous gray shales with abundant *Zoophycus* spreiten. These shales pinch out against the swells, and rare laterally restricted fossil pavements and pyritic horizons within this interval can be seen to truncate against the sides of the swells and may slope away from them (see Fig. 9B). The Girdle Road Bed, which contains a moderately diverse fauna, is amalgamated to the tops of the larger swells, and may even truncate them somewhat (see Fig. 9A). The fossiliferous gray calcareous shales above the Girdle Road Bed contain closely-spaced shell beds that appear relatively unaffected by the underlying irregularities.

Depositional Dynamics

Because of the fine-grained nature of the sediments within which the undulatory Stolle Road Bed surface was formed, it is highly unlikely that the swells represent bedforms built by bedload transport. However, two sedimentary features reported from modern marine environments provide possible analogs: 1) longitudinal ripples, and 2) sedimentary furrows. The formation of both of these structures is believed to be associated with secondary helical flow cells within bottom-flowing currents.

Small silty-clay longitudinal ripples have been described from the ocean floor at depths of almost 5000 meters. These 2-10 meter long bedforms are symmetrical in cross-section, up to 15 cm high and 75 cm wide and are spaced about 1-2 meters apart over an essentially flat surface (Flood, 1981a). Though these ripples are smaller than the Wanakah mud swells by more than an order of magnitude, their morphology is surprisingly similar. Also, a field of discontinuous ripples developed on a planar surface would yield a cross-sectional appearance very like that of the swells, with apparent uneven spacing and variable amplitudes. The depositional dynamics of these longitudinal ripples therefore needs to be considered. McCave and others (1984) argue that the ripples were formed by rapid deposition from a concentrated suspension under waning current conditions. They and Flood (1981a) suggest that short-lived, high velocity bottom currents eroded 1-2 cm of fine sediment over the ripple field, and then redeposited it in longitudinal ripples in response to secondary helical flow.

Large linear erosive features, probably similar in origin to gutter casts, have been described from both shallow-water (Flood, 1981b) and abyssal environments (Hollister *et al.*, 1974; Flood and Hollister, 1980). These sedimentary furrows appear to provide the best actualistic model for the undulatory surface. Furrows are characterized by steep sides and flat floors, and range in size from gutter-like "minifurrows" (Flood, 1981b) to large furrows up to 150 meters wide and 20 meters deep (Hollister *et al.*, 1974). Of particular interest is the presence of shell lags on the furrow floors. The shallow-water estuarine furrows described by Flood (1981b) have layers of articulated cockle shells up to 7 cm thick. These are strikingly similar to the

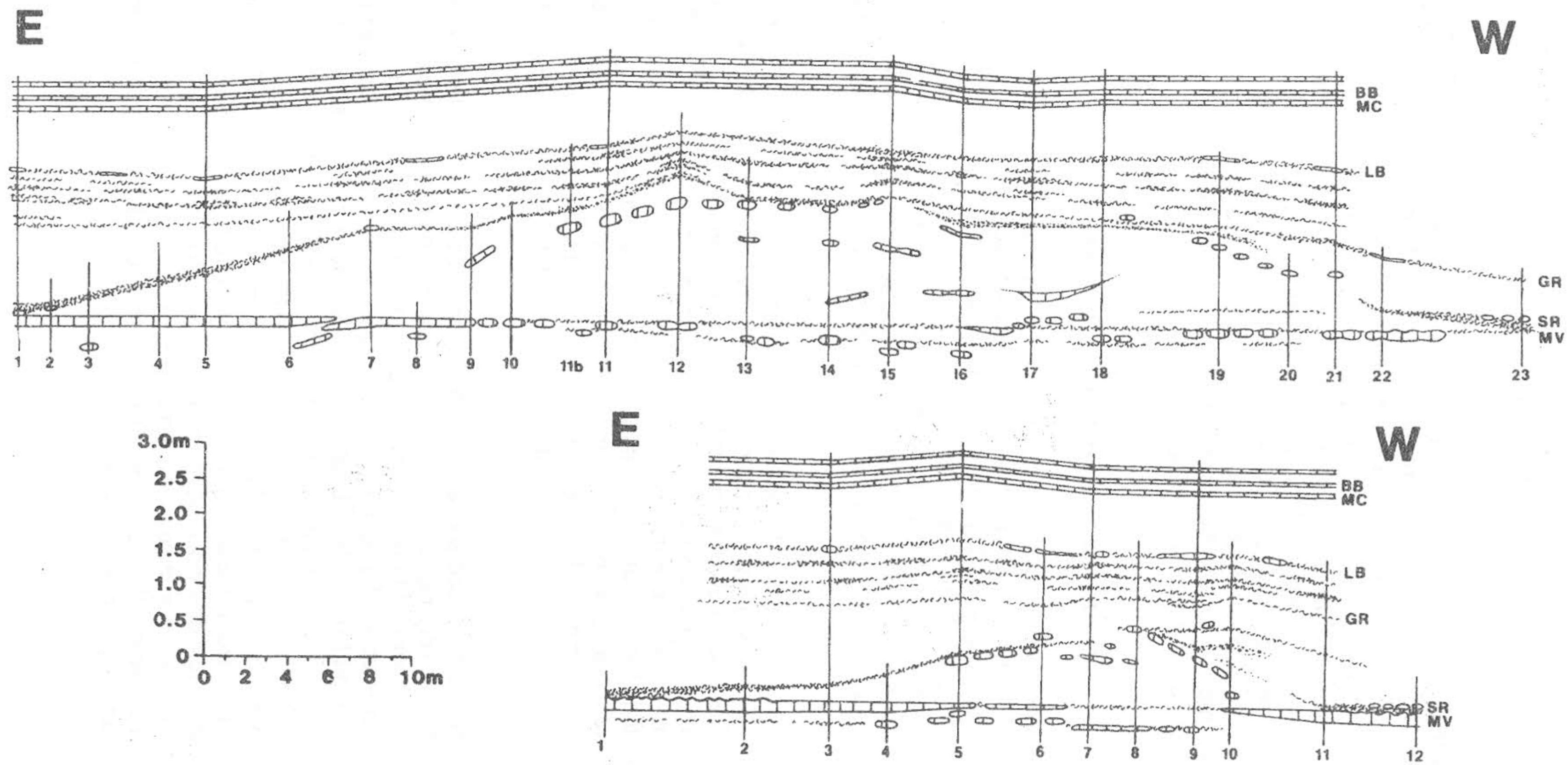


Figure 9. Detailed drawings of two swells based on measurements from closely spaced vertical transects: (A) corresponds to swell 2 of Figure 4, and (B) corresponds to swell 3. Numbers mark measured vertical transects. Important marker beds are indicated by letters as follows: MV) Mt. Vernon Bed; SR) Stolle Road Bed; GR) Girdle Road Bed; LB) Lakeview Bed; MC) Murder Creek Bed; and BB) Bidwell Bed, which marks top of Darien Center Submember. Note the three distinct stratigraphic intervals discussed in text: 1) between the undulating Stolle Road Bed and the Mt. Vernon Bed, 2) between the Stolle Road and Girdle Road Beds, and 3) between the Girdle Road and Lakeview Beds. Although shown blank, the interval between the Lakeview and Murder Creek Beds is also highly fossiliferous and characterized by closely spaced shell beds.

Ambocoelia-rich fossil layers of the Stolle Road Bed found on the "channel floors" between the swells.

As in the case of longitudinal ripples, the origin of furrows is believed to be associated with secondary helical flow cells. Flood (1981b) proposes that "minifurrows" are initiated by the scouring action of rows of shells aligned by converging helical cells. According to this interpretation, these minifurrows are subsequently deepened and widened during "erosional events" by the abrasive action of the coarse shell material concentrated on the furrow floors. Study of deep-sea furrows by Flood and Hollister (1980) has shown that once formed, furrows may persist for periods exceeding 10,000 years. Furthermore, they also appear to be fairly dynamic with active deposition on furrow walls. This observation is significant in indicating that furrows may be formed by combinations of erosive and depositional processes under at least moderately aggradational conditions.

Proposed Depositional History

From the preceding discussions of the inferred sea level fluctuations within the Darien Center Submember, and of the physical processes which may have contributed to the formation of the undulatory surface, a plausible depositional history can be constructed. The principle phases of this history are 1) the formation of the mud swells on a previously planar surface, 2) the filling of "channels" between the swells in response to sea-level rise, 3) the development of a shelly lag during subsequent sea-level fall, and 4) the accumulation of winnowed/non-depositional shell beds and rapidly deposited mud layers on a nearly planar storm-influenced sea floor.

Based on the models of longitudinal ripples and sedimentary furrows, two scenarios seem possible for producing the unusual irregular geometry of the nearly barren mudrocks overlying the Mt. Vernon Bed. The first involves a primarily erosive mechanism following initial rapid mud deposition. During accentuated shallowing resulting from the regression of the first subcycle, episodic storm-driven currents began resuspending and entraining sediment. Erosion was concentrated along linear current-parallel paths by helical secondary flow. Beginning as gutter-like scours, furrows deepened and widened with time, at least in part due to the abrasive action of shell material concentrated on their floors. Bottom current flow became more and more channelized, and the furrows widened until the present geometry was attained.

According to the second proposed scenario, the formation of the mud swells proceeded synchronously with the relatively rapid aggradation of sediment over the Mt. Vernon Bed. In this case, episodic storm-generated bottom currents, laden with suspended sediment winnowed from nearer shore by wave action, deposited sediment in parallel longitudinal features in response to secondary helical flow cells. Elongate mud ridges were built up progressively as mud continued to be supplied to the shelf area below storm wave-base. As in the first scenario, the developing bottom topography increasingly controlled subsequent current flow patterns. Sediment bypass and winnowing in the areas between ridges, especially during subsequent shallowing, resulted in the accumulation of a styliolinid and *Ambocoelia*-rich shell hash layer. Both scenarios just described have end results which could be quite similar. The preesent data probably points to some combination of these processes. The irregular winnowed surfaces within the swells would seem to indicate a complex dynamic history involving both erosive and depositional processes.

Events following the formation of a field of elongate onshore-offshore-oriented mud ridges or furrows can be more confidently reconstructed. With the change to deepening conditions associated with a second transgressive/regressive subcycle, the offshore-directed bottom currents no longer affected the area, and sediment began accumulating between the ridges. The very low fossil abundances and near absence of shelly horizons within the shales filling the inter-swell "channels" point to a relatively rapid filling of the bottom topography. By the conclusion of this depositional phase, the sea floor was again nearly planar.

Another accelerated shallowing, due to the regressive kick of the second subcycle, resulted in widespread sediment bypass and the formation of the nearly planar Girdle Road Bed. Subsequent slight deepening of the third subcycle was accompanied by renewed deposition under conditions of episodic storm influence near the limit of storm wave-base. Nearly uniform, rapidly deposited, storm mud layers alternating with simple shell beds and shell pavements now accumulated over a flat, very gently sloping bottom. The regressive kick of the third transgressive/regressive subcycle resulted in the generation of a multiply rewinnowed complex shell bed (ie. the Lakeview Bed) which is overlain by the coral-rich horizon marking the time of maximum shallowing of the larger Darien Center Submember cycle. After a period of low sea level stand, a series of three transgressive pulses, punctuated by relatively rapid regressions or stillstands, returned conditions to those that existed during the formation of the Mt. Vernon Bed. These transgressive pulses reflect the three subsequent subcycles, and were associated with relatively low sedimentation rates resulting in a more condensed and fossil-rich interval.

ACKNOWLEDGMENTS

This paper is based on work conducted under the direction of C. E. Brett with the support of a grant from the National Science Foundation (no. EAR-8313103). I express my appreciation to R. R. West and A. W. Archer whose critical reviews helped clarify my thoughts and expression. My wife R. D. Miller provided assistance in the field, and with word processing.

REFERENCES CITED

AGER, D.V., 1981, The Nature of the Stratigraphical Record, 2nd edition, John Wiley & Sons, New York, 122 p.

AIGNER, T., 1984, Dynamic stratigraphy of epicontinental carbonates, Upper Muschelkalk (M. Triassic), South-German Basin: Neues Jb. Geol. Paläont. Abh., v.169, p.127-159.

_____, 1985, Storm depositional systems: dynamic stratigraphy in modern and ancient shallow-marine sequences: Lecture Notes in Earth Sciences 3, Springer-Verlag, Berlin, 174p.

_____ and Reineck,H.-E., 1982, Proximality trends in modern storm sands from the Helgoland Bight (North Sea) and their implications for basin analysis: Senckenbergiana Maritima, v.14, p.183-215.

ALGEO, T.J. and WILKINSON, B.H., 1988, Periodicity of mesoscale Phanerozoic sedimentary cycles and the role of Milankovitch orbital modulation: Jour. Geol., v.96, p.313-322.

BAIRD, G.C., 1981, Submarine erosion on a gentle paleoslope: a study of two discontinuities in the New York Devonian: Lethaia, v.14, p.105-122.

BERNER, R.A., 1968. Calcium carbonate concretions formed by the decomposition of organic matter: Science, v.159, p.195-197.

BRETT, C.E. and BAIRD G.C., 1985, Carbonate-shale cycles in the Middle Devonian of New York: An evaluation of models for the origin of limestones in terrigenous shelf sequences: Geology, v.13, p.324-327.

_____ and _____, 1986a, Comparative taphonomy: a key to paleoenvironmental interpretation based on fossil preservation: Palaios, v.1, p.207-227.

_____ and _____, 1986b, Symmetrical and upward shallowing cycles in the Middle Devonian of New York State and their implications for the punctuated aggradational cycle hypothesis: Paleooceanography, v.1, p.431-445.

_____ and COTTRELL, J.F., 1982, Substrate specificity in the Devonian tabulate coral *Pleurodictyum* : Lethaia, v.15, p.247-262.

_____, SPEYER, S.E. and BAIRD, G.C., 1986a, Storm-generated sedimentary units: tempestite proximality and event stratification in the Middle Devonian Hamilton Group of New York, *in* Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part 1: New York State Mus. Bull., v. 457, p.129-156.

_____, BAIRD, G.C. and MILLER, K.B., 1986b, Sedimentary cycles and lateral facies gradients across a Middle Devonian shelf-to-basin ramp, Ludlowville Formation, Cayuga Basin, *in* Cisne, J.L., ed., New York State Geol. Assoc. 58th Annual Mtg. Field Trip Guidebook, Cornell University, p.81-127.

_____, MILLER, K. B. and BAIRD, G. C., in press, A temporal hierarchy of paleoecologic processes within a Middle Devonian epeiric sea, *in* Miller, W., III, ed., Paleocommunity Temporal Dynamics: The Long-Term Development of Multispecies Assemblies: Paleontological Society Special Publication.

BUSCH, R.M. and ROLLINS, H.B., 1984, Correlation of Carboniferous strata using a hierarchy of transgressive-regressive units: Geology, v.12, p.471-474.

_____ and WEST, R.R., 1987, Hierarchical genetic stratigraphy: A framework for paleoceanography: Paleoceanography, v.2, p.141-164.

COOPER, G.A., 1930, Stratigraphy of the Hamilton Group of New York: Amer. Jour. Sci., v.19, p.116-134, 214-236.

COTTER, E., 1983, Silurian depositional history, *in* Nickelsen, R.P. and Cotter, E., eds., Silurian Depositional History and Alleghanian Deformation in the Pennsylvania Valley and Ridge: 48th Annual Field Conference of Pennsylvania Geologists, Lewisburg, p.3-28.

DENNISON, J.M., 1985, Catskill delta shallow marine strata, *in* Woodrow, D.L. and Sevon, W.D., eds., The Catskill Delta: Geol. Soc. Amer. Spec. Publ. 201, p.91-106.

DOTT, R.H., 1983, 1982 S.E.P.M. Presidential address: Episodic sedimentation -- How normal is average? How rare is rare? Does it matter? : Jour. Sed. Petrol., v.53, p.5-23.

DRIESE, S.G., 1988, Depositional history and facies architecture of a Silurian foreland basin, eastern Tennessee, *in* Driese, S.G. and Walker, D., eds., Depositional History of Paleozoic Sequences, Southern Appalachians: Univ. of Tenn., Dept. of Geol. Sci., Studies in Geology, no.19, p.62-96.

FLOOD, R. D., 1981a, Longitudinal triangular ripples in the Blake-Bahama Basin: Marine Geology, v.39, p.M13-M20.

_____, 1981b, Distribution, morphology, and origin of sedimentary furrows in cohesive sediments, Southampton Water: Sedimentology, v.28, p.511-529.

_____ and HOLLISTER, C. D., 1980, Submersible studies of deep-sea furrows and traverse ripples in cohesive sediments: Marine Geology, v.36, p.M1-M9.

GOODWIN, P.W. and ANDERSON, E.J., 1985, Punctuated aggradational cycles: general hypothesis of episodic stratigraphic accumulation: Jour. Geol. v.93, p.515-533.

_____, _____, GOODMAN, W.M. and SAROKA, L.J., 1986, Punctuated aggradational cycles: implications for stratigraphic analysis: Paleoceanography, v.1, p.417-429.

GRABAU, A.W., 1899, The palaeontology of Eighteen Mile Creek and the lake shore sections of Erie County, New York: Buffalo Soc. Nat. Sci. Bull., v.6, nos. 2,3,4.

HATTIN, D.E., 1985, Distribution and significance of widespread, time-parallel pelagic limestone beds in the Greenhorn Limestone (Upper Cretaceous) of the central Great Plains and southern Rocky Mountains, in Pratt, L.M., Kauffman, E.G., and Zelt, F.B., eds., Fine-grained Deposits and Biofacies of the Cretaceous Western Interior Seaway: Evidence of Cyclic Sedimentary Processes: Soc. Econ. Paleon. Mineral., 2nd Annual Midyear Mtg., Field Trip Guidebook No. 4, Golden, Colo., p.28-37.

HOLLISTER, C. D., FLOOD, R. D., JOHNSON, D. A., LONSDALE P. and SOUTHARD, J. B., 1974, Abyssal furrows and hyperbolic echo traces on the Bahama Outer Ridge: Geology, v.2, p.395-400.

JOHNSON, J.G., KLAPPER, G. and SANDBERG, C.A., 1985, Devonian eustatic fluctuations in Euramerica: Geol. Soc. Amer. Bull., v.96, p.567-587.

KAUFFMAN, E.G., 1988, Concepts and methods of high-resolution event stratigraphy: Ann. Rev. Earth Planet. Sci., v.16, p.605-654.

KLAPPER, G., 1981, Review of New York Devonian conodont biostratigraphy, *in* Oliver, W.A. and Klapper, G., eds., Devonian Biostratigraphy of New York, Part 1: International Union of Geol. Sci. Subcomm. on Devonian Stratigraphy, pp.57-66.

KLOC, G.J., 1983, Stratigraphic distribution of ammonoids from the Middle Devonian Ludlowville Formation in New York [unpub. MA thesis]: State Univ. at Buffalo, 73p.

McCAVE, I.N., HOLLISTER, C.D., DeMASTER, D.J., NITTROUER, C.A., SILVA, A.J. and YINGST, J.Y., 1984, Analysis of a longitudinal ripple from the Nova Scotian continental rise: Marine Geology, v.58, p.275-286.

McCOLLUM, L.B., 1981, Distribution of marine faunal assemblages in a Middle Devonian stratified basin, lower Ludlowville Formation, New York [unpub. doctoral dissertation]: State University of New York at Binghamton, 143 pp.

MILLER, K.B., 1986, Depositional environments and sequences, "*Pleurodictyum* Zone", Ludlowville Formation of western New York, *in* Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part 1: New York State Mus. Bull. v.457, p.57-77.

_____, 1988, A temporal hierarchy of paleoecologic and depositional processes across a Middle Devonian epeiric sea [doctoral dissertation]: Univ. of Rochester, 249p.

_____, in press, High-resolution correlation within a storm-dominated muddy epeiric sea: taphofacies of the Middle Devonian Wanakah Member, *in* Landing, E. and Brett, C.E., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part 2: New York State Mus. Bull.

_____, PARSONS, K.M. and BRETT, C.E., 1988, The paleoecologic significance of storm-generated disturbance within a Middle Devonian muddy epeiric sea: Palaios v.3, p.35-52.

NORRIS, R.D., 1986, Taphonomic gradients in shelf fossil assemblages: Pliocene Purisma Formation, California: Palaios, v.1, p.256-270.

PARSONS, K.M., BRETT, C.E. and MILLER, K.B., 1988, Taphonomy and depositional dynamics of Devonian shell-rich mudstones: Palaeogeog. Palaeoclim. Palaeoecol., v.63, p.109-139.

READ, J.F., 1980, Carbonate ramp-to-basin transitions and foreland basin evolution, Middle Ordovician, Virginia Appalachians: Amer. Assoc. of Petrol. Geol. Bull., v.64, p.1575-1612.

SAVARESE, M., GRAY, L.M. and BRETT, C.E., 1986, Faunal and lithologic cyclicity in the Centerfield Member (Middle Devonian: Hamilton Group) of western New York: a reinterpretation of depositional history, *in* Brett, C.E., ed., Dynamic Stratigraphy and Depositional Environments of the Hamilton Group (Middle Devonian) in New York State, Part 1: New York State Mus. Bull. v.457, p.32-56.

VOGEL, K., GOLUBIC, S. and BRETT, C.E., 1987, Endolith associations and their relation to facies distribution in the Middle Devonian of New York State: Lethaia, v.20, p.263-290.

WALKER, K.R., SHAMMUGAN, G. and RUPPEL, S.C., 1983, A model for carbonate to terrigenous clastic sequences: Geol. Soc. Amer. Bull., v.94, p.700-712.

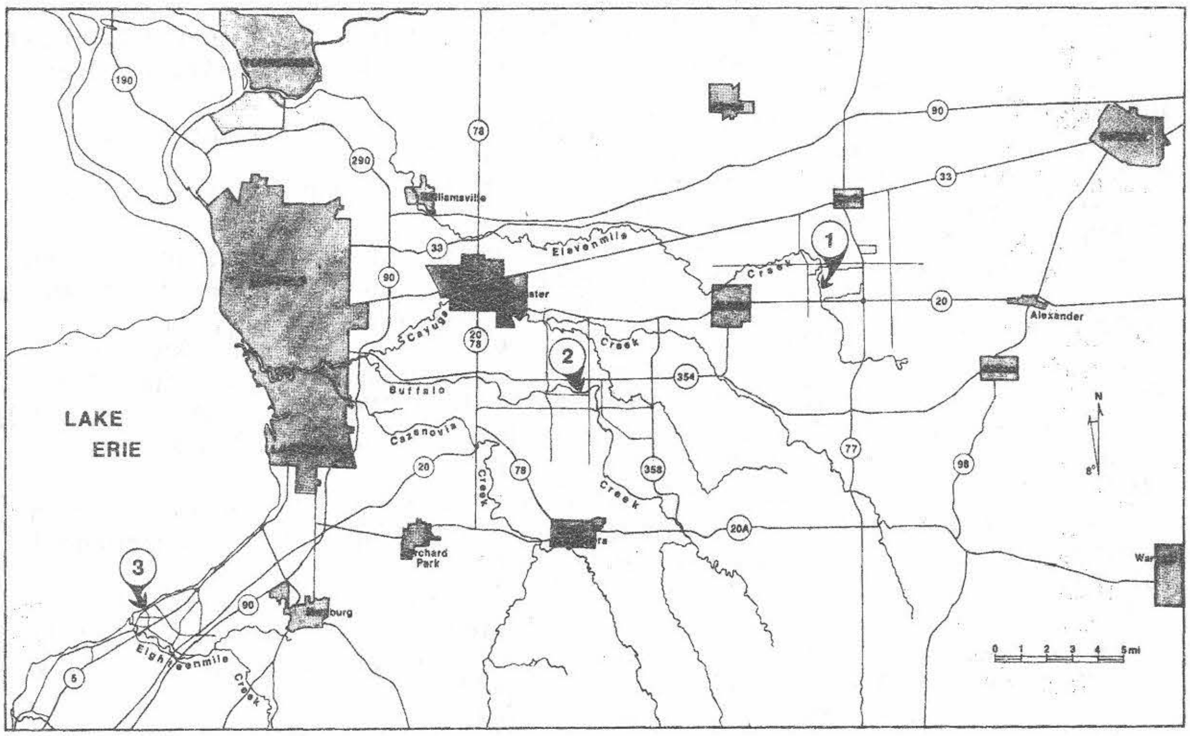


Figure 10. Locality map for the three field trip stops.

ROAD LOG FOR THE PALEOECOLOGY AND SEDIMENTOLOGY OF THE DARIEN CENTER SUBMEMBER

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Road log begins at exit 53 on the New York State Thruway (I 90) around Buffalo. Turn east on Clinton Street (Rt. 354).
2.0	2.0	Cross Union Road (Rt. 277) at Gardenville. Continue on Clinton Street paralleling Buffalo Creek.

4.9	2.9	Cross Transit Road (Rts. 20/78)
12.0	7.1	Cross Two Rod Road (Rt. 358).
14.5	2.5	Cross Cayuga Creek.
14.7	0.2	Bear left (north) onto Exchange Street (Rt. 239) toward the town of Alden.
17.7	3.0	At Alden turn right (east) onto Broadway (Rt. 20).
20.9	3.2	Turn left (north) on Harlow Road. Darien Lakes State Park will be on your right after the turn.
21.4	0.5	Turn right into Park entrance. After passing the gate house drive to left around small recreational lake and then bear right and continue to end of camping area and park vehicles. A nature trail leads to Elevenmile Creek just down the bank.

STOP 1 - ELEVENMILE CREEK AT DARIEN LAKES STATE PARK

The streambank exposure along Elevenmile Creek within the Darien Lakes State Park is the type locality of the Darien Center Submember. The entire 4-meter stratigraphic interval is well exposed here, and the major key beds can be easily recognized at the outcrop. We will use this stop to become familiar with the distinctive character of each of the various key beds discussed in the text. This locality provides an excellent opportunity to closely examine the vertical sequence of shell bed types and fossil assemblages characteristic of the Darien Center Submember.

Figure 11 displays the presence/absence macrofaunal data for this locality. This data reveals a nearly symmetric pattern, interpreted as reflecting a regressive-transgressive eustatic sealevel cycle. The higher faunal diversities of the transgressive half of the cycle probably reflect lower turbidity conditions, and are associated with more closely spaced and more condensed shell beds.

Near the center of the cyclic sequence is a prominent biostromal coral bed. Named the Darien Coral Bed, it attains its best development at this locality, and is dominated by the branching tabulate coral *Eridophyllum*. At Buffalo Creek (Stop 2) to the west, this same coral bed is represented only by widely scattered *Heliophylum* and *Cystiphylloides* corals. Along the outcrop belt to both the east and west of Elevenmile Creek, the Darien Coral Bed thins and becomes a horizon of scattered rugose corals before ceasing to be a recognizable marker bed (see Fig. 4 in the text).

NOTE: NO COLLECTING IS PERMITTED ON THE PROPERTY OF THE DARIEN LAKES STATE PARK.

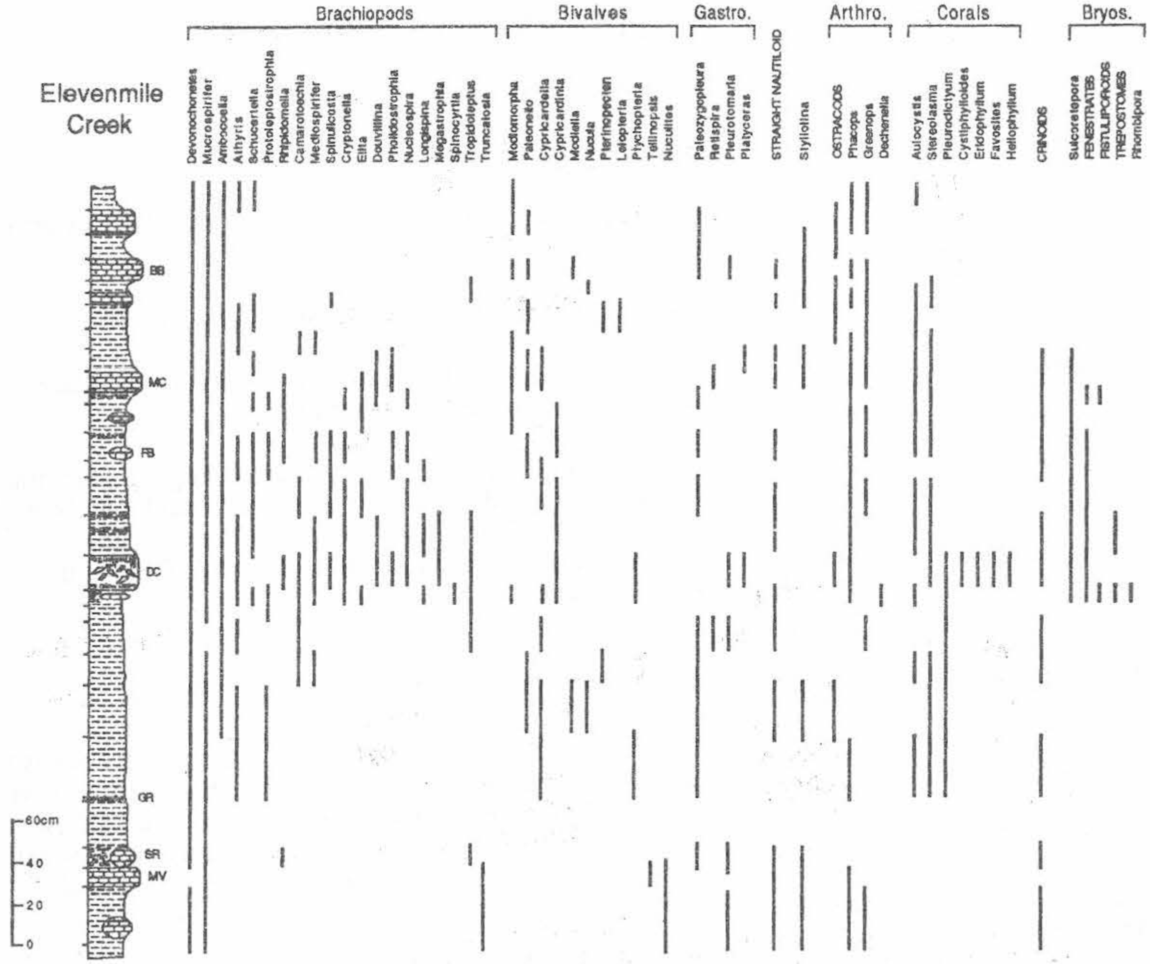


Figure 11. Stratigraphic column of Elevenmile Creek section with bed-by-bed presence/absence macrofaunal data. Note the general faunal symmetry of the interval reflecting a regressive-transgressive cycle, as well as the sharp faunal changes at the Girdle Road and Lakeview Beds. Important marker beds are indicated by letters as follows: MV) Mt. Vernon Bed, SR) Stolle Road Bed, LB) Lakeview Bed, MC) Murder Creek Bed, and BB) Bidwell Bed.

21.4	0.0	Turn left (south) on Harlow Road upon leaving Darien Lakes State Park.
21.9	0.5	Turn right (west) on Broadway (Rt. 20) toward Alden
25.1	3.2	At Alden turn left (south) onto Exchange Street (Rt. 239)
28.1	3.0	Turn right (west) onto Clinton Street (Rt. 354).
30.8	2.7	Cross Two Rod Road.

33.3	2.5	Turn left (south) onto Girdle Road.
33.6	0.3	Girdle Road Bridge over Buffalo Creek. Park along right side of road just before bridge. We will walk to south side of Buffalo Creek, walk down bank to the creek below and proceed downstream.

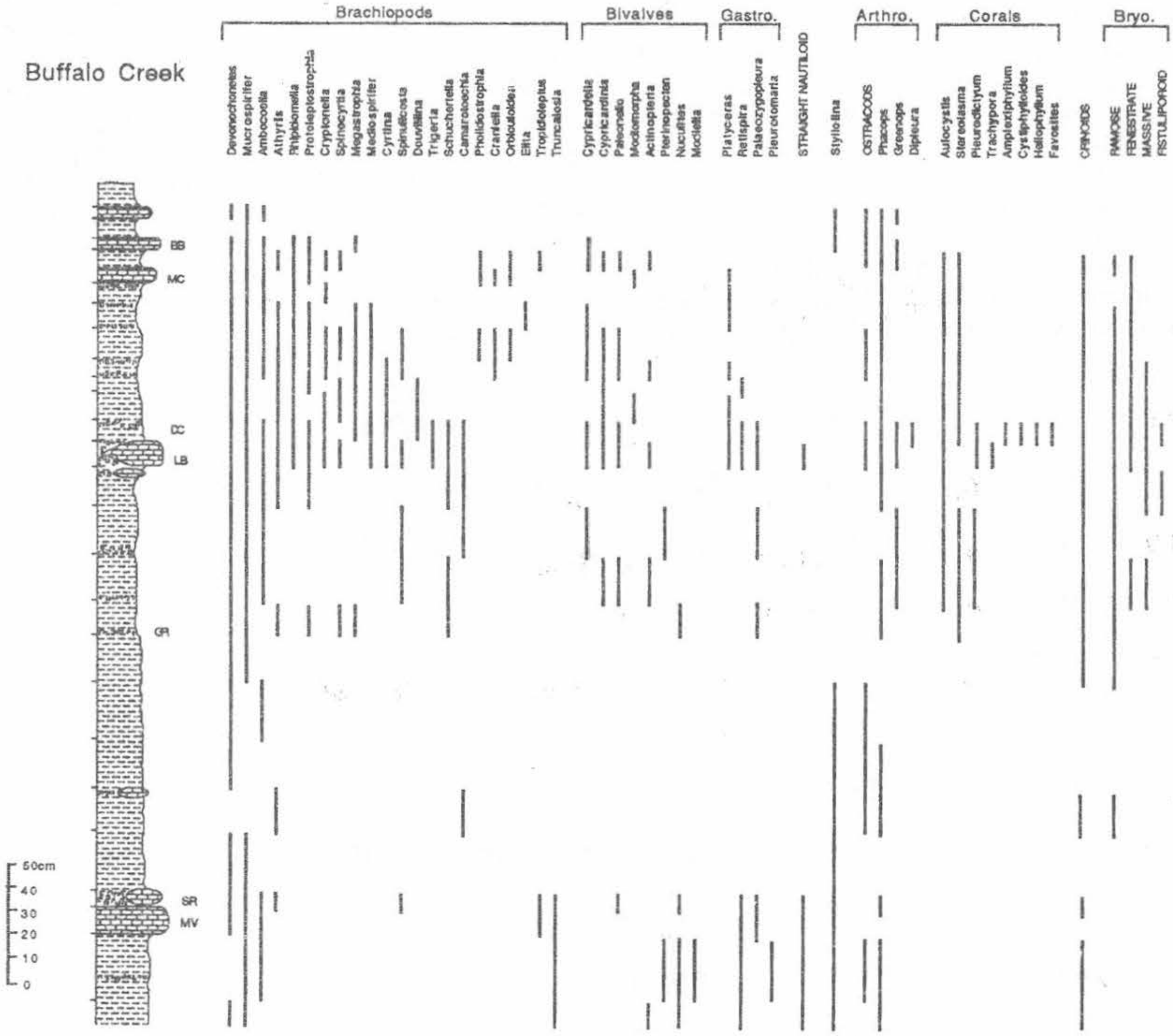


Figure 12. Stratigraphic column of Buffalo Creek section with bed-by-bed presence/absence macrofaunal data. Note the highly condensed transgressive portion of the cycle at this locality.

STOP 2 - BUFFALO CREEK AT GIRDLE ROAD BRIDGE

A continuous cut bank on the south side of Buffalo Creek to the west of the Girdle Road bridge exposes the entire Darien Center Submember for a distance of over 500 meters. At this locality, the Darien Center cycle is only 3 meters thick and the transgressive half is highly condensed. The presence/absence faunal data of Figure 12 shows this asymmetry and also reveals the sudden diversity changes associated with the traceable key beds. These distinctive marker horizons represent subcycle boundaries and are the basis for correlation across western and central New York State.

At the base of the Darien Center Submember is a peculiar discontinuity surface which is especially well displayed at this locality. This discontinuity is represented by the undulatory Stolle Road Bed which repeatedly separates from and merges with the underlying Mt. Vernon Bed defining a series of mudstone lenses of varying amplitude and width (see Figs. 8 and 9 in text). The mudstone lenses, or what I call "swells," are elongate features trending at a high angle to depositional strike. Dense shell layers dominated by the tiny brachiopod *Ambocoelia* occupy the broad flat "channel" areas between the swells. These shell layers extend up the flanks of the swells, becoming thinner and progressively losing fossil taxa. Over the tops of the swells the discontinuity is often represented only by a horizon of pyritic burrow tubes and pyrite nodules. The geometry of the swells and channels, and their approximately onshore-offshore orientation, suggest an analogy with modern sedimentary furrows produced by helical flow cells within episodically flowing bottom currents. This irregular erosion surface was formed during the early regressive phase of the Darien Center cycle, and may record an accentuated regressive pulse produced by a superimposed subcycle.

33.6	0.0	Continue south on Girdle Road
34.4	0.8	Turn right (west) on Bullis Road
38.9	4.5	Turn left (south) on Transit Road (Rt. 20/78)
39.2	0.3	Interchange with Rt. 400
39.7	0.5	Rt. 78 turns off. Continue on Transit Road
40.6	0.9	Bear right on Rt. 20
43.9	3.3	Cross Union road (Rt. 277)
45.5	1.6	Cross Southern Expressway (Rt. 219)
48.2	2.7	Pass under New York Thruway
48.6	0.4	Cross Rt. 62
50.4	1.8	Cross Rt. 75
56.2	5.8	Turn right (west) on South Creek Road after crossing Eighteenmile Creek
56.5	0.3	Pass through North Evans
57.4	0.9	Pass under railroad bridge
58.6	1.2	Cross Rt. 5

58.9 0.3 Turn right (northeast) on Lake Shore Road. After only a few hundred feet (before the bridge) turn right onto gravel drive and park in privately owned gravel parking area.

Walk across the bridge over Eighteenmile Creek and then proceed toward the lake shore along the north bank. At the mouth of the creek walk north along the lake shore

STOP 3 - LAKE ERIE SHORE NORTH OF EIGHTEENMILE CREEK

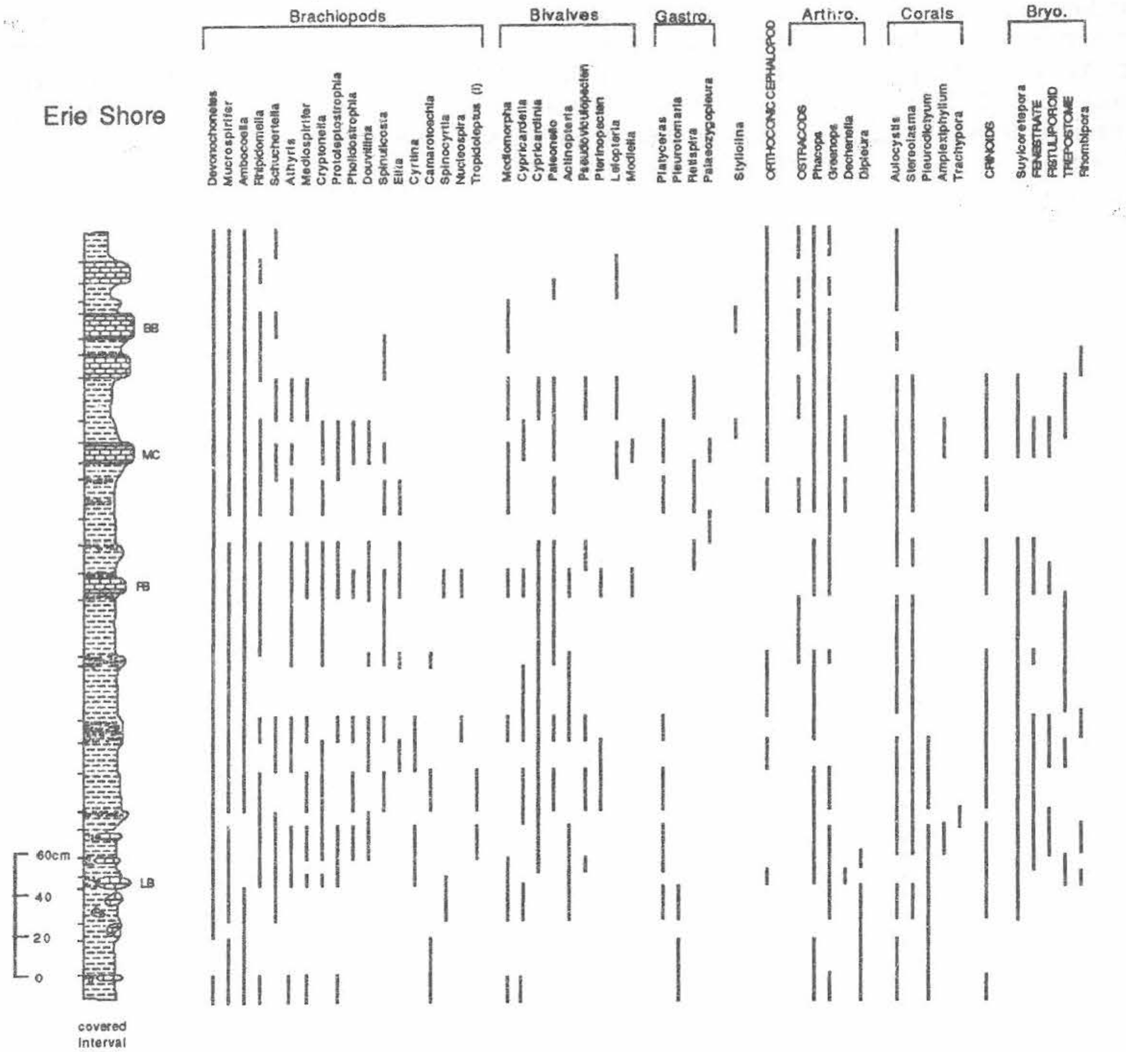


Figure 13. Stratigraphic column of Lake Erie shore section with bed-by-bed presence/absence macrofaunal data. Most of the regressive portion of the cycle is below the lake level at the Stop 3 locality.

This beautiful cliff exposure provides access to the upper transgressive half of the Darien Center Submember which alone is about 3 meters thick. The lower portion of the section is below lake level at this locality. The closely spaced shell beds can be easily examined and sampled here, and fossil collecting is excellent. Figure 13 shows the presence/absence faunal data for these lake shore cliffs.

The major key beds are easily recognized here. Especially well displayed at and near the lake level are the Lakeview Bed and the underlying shell beds containing abundant *Pleurodictyum* corals. Exposed bedding plane surfaces of these beds reveal well preserved *in situ* fossil assemblages on shell bed tops. Articulated and life position brachiopods are abundant, and complete trilobites and camerate crinoids can often be found. These fossils represent smothered epifaunal communities which had colonized winnowed shell hash layers. The shales overlying the shell beds are nearly barren of fossils and probably represent muds rapidly deposited by storm events. These muds were likely responsible for the preservation of the underlying shell beds by removing them from the sediment/water interface where taphonomic loss is great, and by stimulating early diagenetic pyritization and carbonate concretion formation.

SHALLOW WATER REEFS OF THE MIDDLE DEVONIAN EDGECLIFF MEMBER OF THE ONONDAGA LIMESTONE, PORT COLBORNE, ONTARIO, CANADA

THOMAS H. WOLOSZ
Center for Earth & Environmental Science
SUNY College at Plattsburgh
Plattsburgh, N.Y. 12901

INTRODUCTION

The reefs of the Edgecliff member of the Onondaga Formation are a well known part of the geology of New York State and the Niagara Peninsula in Ontario, Canada. Over the past thirty to forty years, the Onondaga and its reefs have been the subject of a number of Master's theses and Doctoral dissertations (see Wolosz, 1984, for references), and yet there are still a large number of unanswered questions regarding these reefs and the Onondaga in general.

The Edgecliff has generally been assumed to represent a warm, tropical, shallow water environment due to its abundant and diverse coral fauna. However, the lack of stromatoporoids and calcareous algae is a notably unusual feature for Devonian reefal limestones. The absence of these organisms lead Kissling and his students (Kissling, 1987; Kissling and Coughlin, 1979; Cassa and Kissling, 1982) to interpret these reefs as deep water structures. They have also pointed to the absence of peritidal facies in the Onondaga as added support for this interpretation, suggesting that the preserved Onondaga represents a deep water facies, the shallow water facies which rimmed the basin having been removed by erosion.

Studies of other Edgecliff reefs have yielded results which do not support this interpretation. The Thompson's Lake bioherm has been interpreted as having grown into the surf zone (Williams, 1980), while the LeRoy Bioherm has been interpreted as a shallow water deposit by Poore (1969), Lindemann (1988), and Wolosz (1988). Wolosz (1984, 1985) presented evidence of coral breakage and overturning to argue for shallow water conditions during reef growth at Roberts Hill and Albrights Reefs.

This field trip will examine the Ridgemount and Quarry Road bioherms, which are being interpreted by the author as the shallowest water reefs known from the Onondaga.

STRATIGRAPHY

Our current understanding of the Onondaga is due, in large part, to the work of Oliver (see Oliver, 1976, for extensive reference list), who used gross lithology and fauna to subdivide the Onondaga into the Edgecliff, Nedrow, Moorehouse and Seneca members. Lindholm (1967) later subdivided the Onondaga into four petrographic microfacies which in some areas deviate greatly from Oliver's stratigraphy (Figure 1), while Coughlin (1980) and

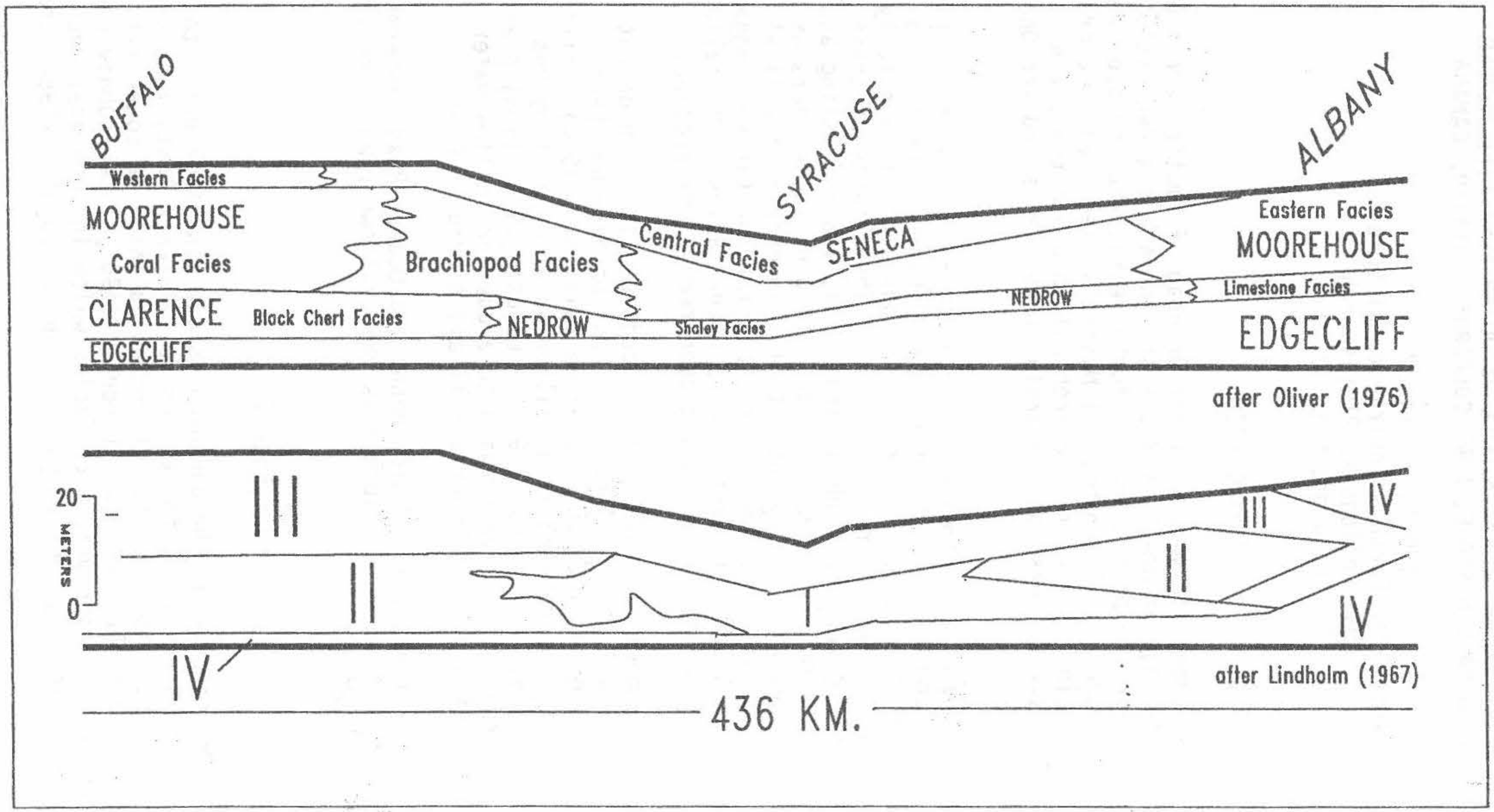


Figure 1. Onondaga stratigraphy and facies distribution. Upper cross-section illustrates Oliver's (1976) designation of members and facies based on biostratigraphy and megascopic rock characteristics. Thin Cl in east not shown. Lower cross-section illustrates Lindholm's (1967) microfacies distributions (I=fossiliferous calcisiltite with about 25% clay and less than 10% fossils; II=fossiliferous calcisiltite with about 5% clay and less than 10% fossils; III=bio-calcisiltite with 10-50% fossils; and IV=biosparite and bio-calcisiltite with greater than 50% fossils.)

Cassa (1980) presented interpretations of Onondaga subsurface stratigraphy across much of central and western New York State.

The Onondaga ranges up to approximately 34 meters in thickness in the eastern part of New York, but with the exception of the basal 2 meter "C1" micrite, it consists of crinoidal packstone/grainstone which is divisible into members only on the basis of biostratigraphy. In central New York, the formation thins to roughly 21 meters, but is easily divided on lithologic grounds into Oliver's four members, with the Edgecliff a massive, biostromal, very coarsely crystalline limestone from about 2.5 to 7.5 meters thick; the Nedrow a thin bedded, very fine grained shaley limestone; the Moorehouse a very fine grained limestone with chert and shaley partings; and the Seneca similar to the Moorehouse lithologically, but with a different fauna. Near Buffalo the formation reaches a thickness of 43 meters with only a very thin Edgecliff unit (about 1.5 meters). The Nedrow equivalent in this area is a sparsely fossiliferous, fine grained, chert-rich limestone which Ozol (1963) named the Clarence member. Both Lindholm (1967) and Messolella (1978) identified the central New York Onondaga as representing the most basinal facies exposed at the surface, and located the topographic axis of the basin through that area (Fig.2A).

The basal contact of the Onondaga is marked by a widespread unconformity (Rickard, 1975). In the east, the contact with the underlying Schoharie Formation has alternatively been interpreted as gradational (Goldring and Flower, 1942) or disconformable, with the presence of a glauconitic sand bed cited as evidence of a period of nondeposition (Chadwick, 1944). In the central part of the state, the base of the Onondaga is marked by the "Springvale Sand" which overlies either the patchily distributed Lower Devonian Oriskany Sandstone, or the older Helderberg limestones. The underlying units continue to be variable to the west, where the Onondaga rests upon either the Lower Middle Devonian Bois Blanc Formation or Silurian dolomites.

In the field trip area, the stratigraphy of the Onondaga is somewhat variable. In the Ridgemount Quarry the base of the Onondaga is marked by the presence of the green/gray fine "Springvale" sand which ranges from about 0.3 to 0.6 meter thick. The "Springvale" appears to be absent to the east of the coral mounds, where the contact is marked by a gray limestone with common solitary rugosans and intraclasts.

Above the basal unit, the northeast side of the quarry is characterized by coral beds, the north side by crinoidal grainstone/packstone interbedded with the "Springvale", and the west and south quarry walls by a roughly 0.3 to 0.6 meter clayey layer. The remainder of the exposed Onondaga along the south and west quarry walls, and the west portion of the north quarry wall consists of a gray-green biostromal limestone with abundant blue-black chert nodules. This unit is also present above the coral beds, where it is somewhat more shaley and roughly 2.7 to 3.6 meters thick; but here it is capped by a coarse bioclastic carbonate sand with abundant coral, which can be seen to pinch out along the north wall of the quarry.

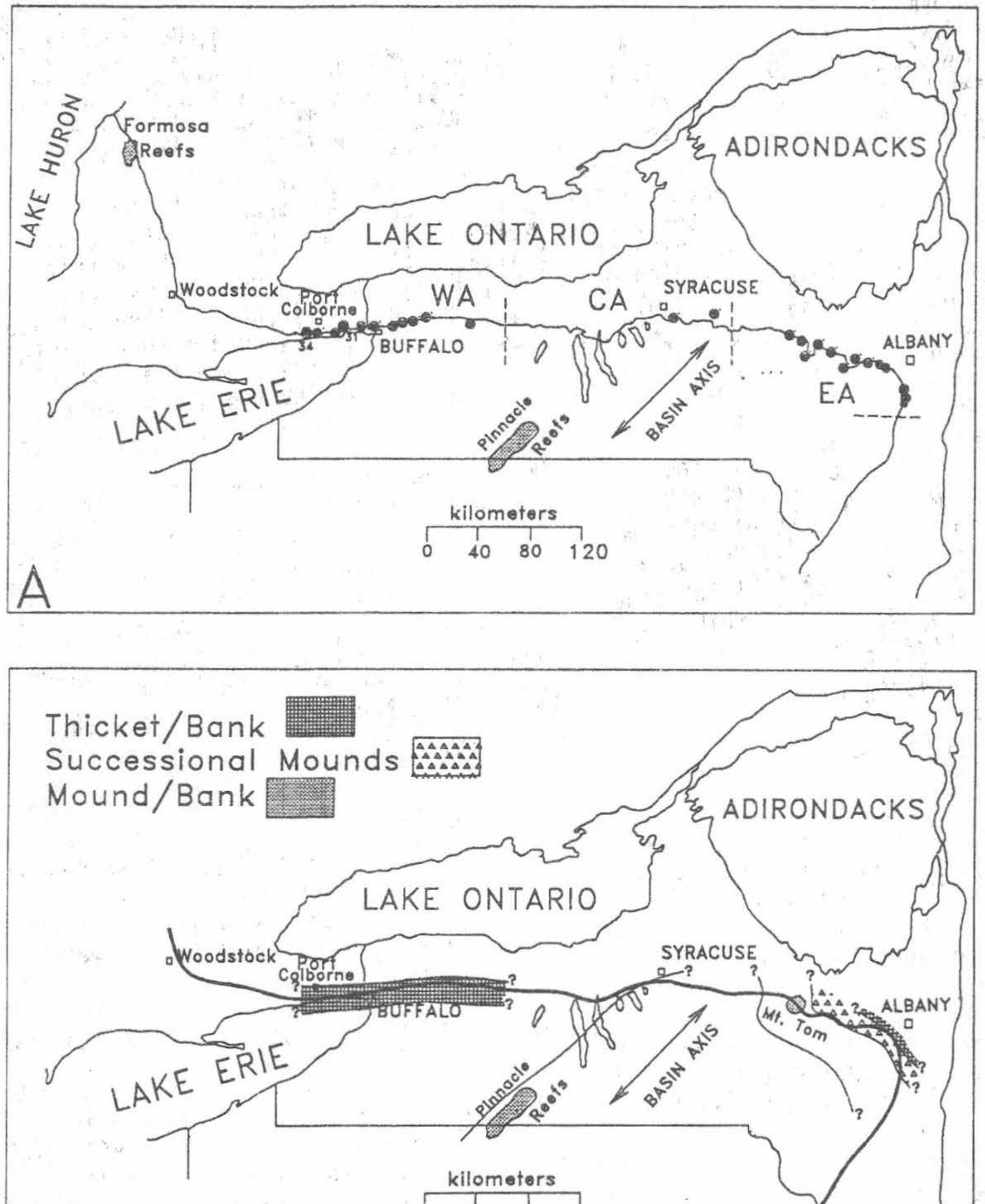


Figure 2A. Distribution of reefs along Onondaga strike belt (dots) in New York and Canada. Refer to Oliver (1976) for locality data. Bioherm #31 is Ridgemount Bioherm, #34 is Quarry Road Bioherm. Pinnacle reefs are subsurface. Formosa reefs are Edgecliff equivalents. Note position of basin axis. 2B. Distribution of Edgecliff reefs by type.

To the southwest, at the Quarry Road exposure and the active quarry on the north side of Route 3 (former Law Quarry), the basal Onondaga is marked by roughly 1.8 to 2.1 meters of greenish, shaley limestone with clay seams and sparse to common coral, followed by from 1.0 to 2.4 meters of limestone which varies from dark gray to light gray to buff with varying densities of clay seams, but characterized by abundant *Cystiphylloides*, a solitary rugose coral. At the former Law Quarry, and along the west wall of the West Quarry at Quarry Road, the remaining Onondaga is a chert rich, biostromal, micritic limestone with colonial coral. Along the east wall of the West Quarry, the *Cystiphylloides* biostrome is capped by a grainstone bed with abundant coral and stromatoporoids. At the south end of the quarry a number of small coral mounds are exposed which are stratigraphically above this grainstone bed and grade laterally into the chert rich, micritic Onondaga facies.

REEF COMMUNITIES

Most Edgecliff reefs include two distinct communities - the Phaceloid Colonial Rugosan Community and the Favositid/Crinoidal Sand Community.

The phaceloid colonial rugosan community is made up almost exclusively of colonial rugosans. Common genera include *Acinophyllum*, *Cylindrophyllum*, and *Cyathocylindrium*; with *Eridophyllum*, *Synaptophyllum*, and possibly phaceloid colonies of *Heliophyllum* as accessories. The dense growth of these rugosan colonies appears to have restricted most other organisms to only minor roles, with favositids (both domal and branching) being small and rare, brachiopods uncommon, and bryozoans mainly fragmentary encrusters.

The favositid/crinoidal sand community displays a much higher diversity than the rugosan community. This community is more biostromal than biohermal. Large sheet-like to domal favositids are abundant, but never form a constructional mass. Solitary rugose corals are also extremely abundant as are fenestrate bryozoan colonies. Single colonies of the mound building phaceloid rugosans are occasionally found. Brachiopods and other reef dwellers are also common although never extremely abundant. Stromatoporoids and massive colonial rugosans, while extremely rare in the Edgecliff reefs, when found are part of this community. The crinoids were the greatest contributor to this community - ossicles making up the bulk of the rock and indicating abundant growth of these organisms - but complete calyces are never found.

REEF CLASSIFICATION

The Edgecliff reefs represent a continuum of growth pattern in which the two communities described above (Phaceloid Colonial Rugosan and Favositid/Crinoidal Sand) are the pure end members. Hence, a simple classification of these reefs is as follows:

Mounds - distinct high relief mounds of the Phaceloid Colonial Rugosan Community. Subdivided into:

1) Successional Mounds - rugosan mounds up to roughly 15 meters thick which display an internal succession of mound building colonial rugosan genera.

2) Small Mounds - small monogeneric to mixed fauna buildups. Generally not more than 1 - 3 meters thick. Commonly found in protected back-reef areas. At least five such structures are known from the field trip area.

Composite Structures - structures formed through interbedding or intergrowth of the two communities. The term "bank" follows the definition of Nelson, et al. (1962): "a skeletal limestone deposit formed by organisms which do not have the ecologic potential to erect a rigid, wave resistant structure." Subdivided into:

1) Mound/Bank - large structures resulting from the repetitive intergrowth of rugosan mounds and the favositid/crinoidal sand facies. Subsurface pinnacle reefs reaching up to 60 meters in thickness represent this type of structure.

2) Ridge/Bank - Colonial rugosans and occasional large favositids form a series of small mounds which coalesce laterally to form an elongate, ridge-like structure. The Ridgemount Bioherm is the only known example of this reef type.

3) Thicket/Bank - The favositid/crinoidal sand facies makes up the main mass of these buildups in the form of gently dipping (5 to 12 degrees), bedded packstone and grainstone with abundant large sheet to domal favositids. The colonial rugosans occur as thickets roughly 0.3 meters thick, which cover the entire bank, and are now interbedded with the biostromal deposits. The resultant structure is a low relief, shield shaped mound up to approximately 300 meters in diameter and 15 meters thick.

Biostrome - bedded Favositid/Crinoidal Biostrome, typical bedded Edgecliff, with no evidence of relief above sea-floor. Banks of pure Favositid/Crinoidal Community have not been recognized, although some Thicket/Bank structures are, volumetrically, very close to this state.

PATTERNS OF EDGECLIFF REEF GROWTH AND THEIR GEOGRAPHIC DISTRIBUTION

Oliver (1976) has presented location data for both the known Edgecliff reef exposures and subsurface reefs (Figure 2A). When the reefs are described using the classification outlined above, and this data added to the reef distribution map, the following patterns emerge (Figure 2B).

The large mound/bank structures rim the axis of major basinal subsidence. Wolosz (1984, 1985, 1989a, 1989b) has argued that the dominant reef community in Edgecliff Reefs was controlled by the level of water turbulence at the crest of the reef. Following this model, Wolosz and Paquette (1988) suggested that the shifts between the Favositid-Crinoidal Sand Community and the Colonial Rugosan Community mark "catch up/fall back" growth cycles as the reef community attempted to maintain itself at a constant water depth during basinal subsidence. The balance of growth versus subsidence resulted in the great thickness of these reefs.

In the east, an off-shore to on-shore trend from mound/bank through successional mounds to thicket/banks is notable (Figure 2B). This is interpreted as a shift from deeper to shallower waters, based upon both geographic considerations - the Adirondack Mountains having been a low land mass to the north, while deeper water lay to the south - and by basinal facies patterns - the mound/bank and successional mound reefs are rooted in the micritic, deeper water Cl Edgecliff facies, while the thicket/bank structures are rooted in the typical grainstone/packstone of the Edgecliff. Further, the thicket/bank structures are characteristic of the assumedly shallow water, western transgressive facies of the Edgecliff from the vicinity of Rochester, westwards into Ontario, Canada (Figure 2B).

The Ridgemount Quarry is the only known example of a ridge/bank reef. Its location roughly 3 kilometers to the north (up dip and shoreward) of the main trend of thicket/bank reefs from Fort Erie to west of Port Colborne suggests a shallower water environment than that in which the thicket/bank reefs formed.

Figure 3 illustrates the water depth relationship among the various reef types.

THE EDGECLIFF REEFS - COOL WATER STRUCTURES?

As mentioned in the introduction, Kissling and his students have pointed to the lack of stromatoporoids and calcareous algae, in conjunction with the absence of clear peritidal deposits to suggest that the Edgecliff reefs may have been deposited in deep water. An alternative hypothesis to the deep water model is for the Edgecliff reefs to have been deposited under cool water conditions. Wolosz and Paquette (1988) suggested a cool water environment for the Edgecliff, as have Koch and Boucot (1982), based on the Edgecliff brachiopod fauna, and Blodgett, et al. (1988), based on gastropod faunas.

The model presented by Wolosz and Paquette (1988) suggests a westward current flow across New York and on into Canada, with water temperatures gradually increasing due to solar warming. Evidence of this trend may be found in the increase in stromatoporoid abundance across New York State and into Canada. In eastern New York stromatoporoids are extremely rare and generally small. Near Rochester (LeRoy Bioherm) they are larger (up to about 0.3 meter in diameter, but only a few centimeters thick), but still uncommon. In the Quarry Road exposures, stromatoporoids are quite common in the carbonate sand facies, still thin, but up to 0.9 to 1.5 meters in diameter.

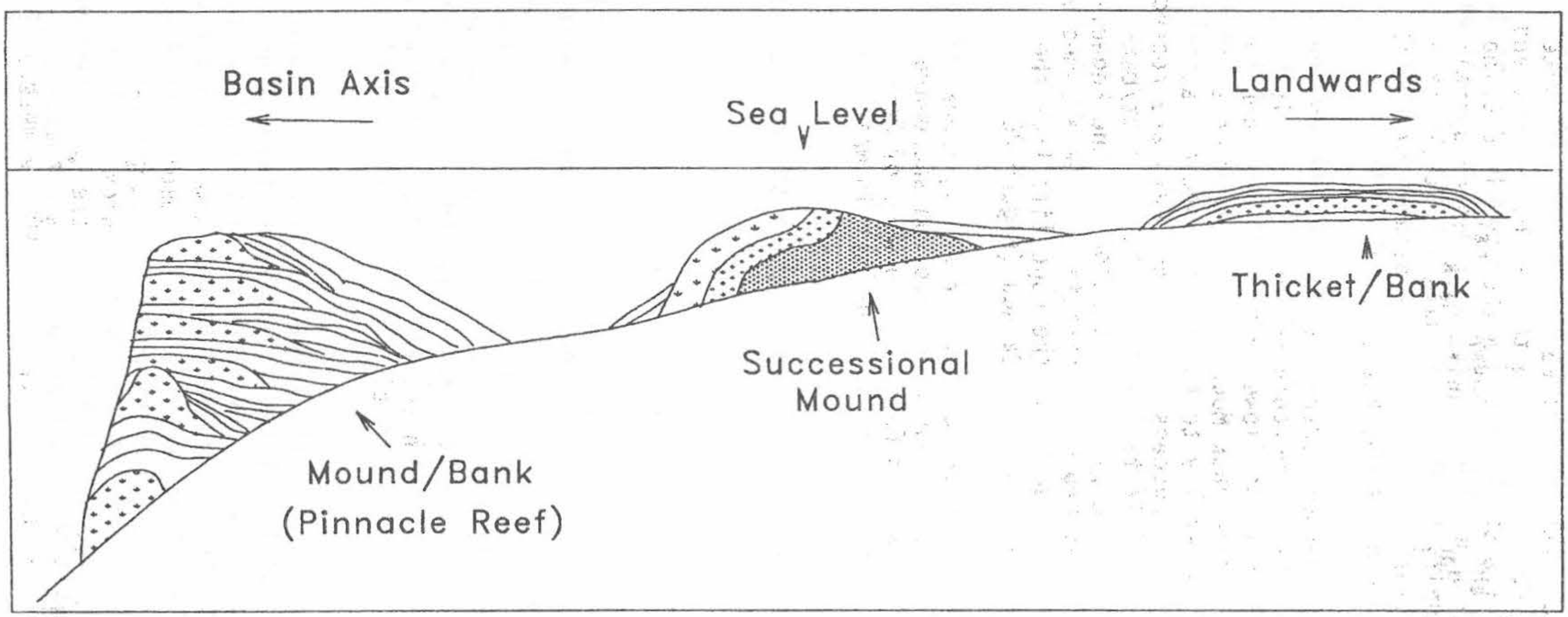


Figure 3. Water depth relationship of reef types. Mound/bank structures indicate areas of subsidence, with active mound growth only at top of structure. Final reef structure is made up of stacked mounds interbedded with favositid dominated, crinoidal grainstone/packstone flank beds. Successional mounds are characteristic of areas of minimal subsidence, although sea-level fluctuation would control coral succession. Flank beds surround the mound. Thicket/bank structure represents shallowest water conditions of three reef types. Colonial rugosan occur only as thickets interbedded with favositid dominated crinoidal sand flanks. Reefs are not drawn to scale.

Finally, the lack of peritidal facies may simply reflect our over-reliance on the tropical Bahamian model for limestone deposition. Lees (1975) illustrated world trends in carbonate deposition, illustrating the fact that temperate water limestones (Foramol Association) do not occur with algal mats, ooids or any of the other tropical carbonate facies.

FIELD TRIP STOPS

Ridgemount Bioherm (STOP 1).

Figure 4A is a map of the Ridgemount Quarry. Note that reef facies are restricted to the northeastern quarter of the quarry. The north wall of the quarry exposes a number of bioherms (Satellite bioherms), while a single bioherm can be examined in the east wall, just north of the water filled section of the quarry (Bioherm A). The northeast wall of the quarry displays abundant coral in a bed which pinches and swells from about 0.7 to roughly 2 meters thick (Coral Beds).

Cassa and Kissling (1982) interpreted the Edgecliff in this quarry as consisting of numerous small bioherms, which is, to a large degree, correct. However, it is important to note the dip of the beds which overlie this coral bed (in particular the quarry walls surrounding tank cars on Upper Level). These exposures are above and to the east of the Coral bed exposure, but they are already dipping westwards, indicating that these small mounds existed behind a larger structure to the east. While the dip has to some degree been enhanced by compaction (especially around Bioherm A), a pinch out of these beds can be observed along the north wall of the quarry, indicating a primary depositional dip. Further, examination of the beds behind the tank cars reveals numerous thickets of *Acinophyllum* and small branching tabulate corals. These thickets have not been observed elsewhere in this quarry, suggesting colonization of a topographic high on the sea-floor. Finally, note the top of the quarry wall (good exposures may be found along the quarry rim above the tank cars, also above Bioherm A) which consists of a clean biostromal sand containing large phaceloid rugosan colonies (*Cyathocylindrium*?) and abundant solitary rugosans and favositids. This facies appears to be restricted to the general vicinity of the stratigraphically lower coral buildups, also supporting the presence of a topographic high on the sea-floor.

Examination of the coral buildups (Bioherm A, Satellite bioherms, coral beds) reveals little internal structure. Unlike the well developed rugosan thickets commonly found in thicket/bank structures or the dense growth of colonial corals within mound structures (as will be seen at the Quarry Road bioherms), these buildups are most commonly characterized by debris and overturned and broken coral colonies (large overturned favositids are especially notable). Bioherm A (Figure 5) is easily accessible for examination. Note the absence of a well developed core facies in this structure. This unusual lack of internal structure is interpreted as indicative of very shallow water conditions during mound development. Intermittent high energy periods (storms) prevented the continuous coral

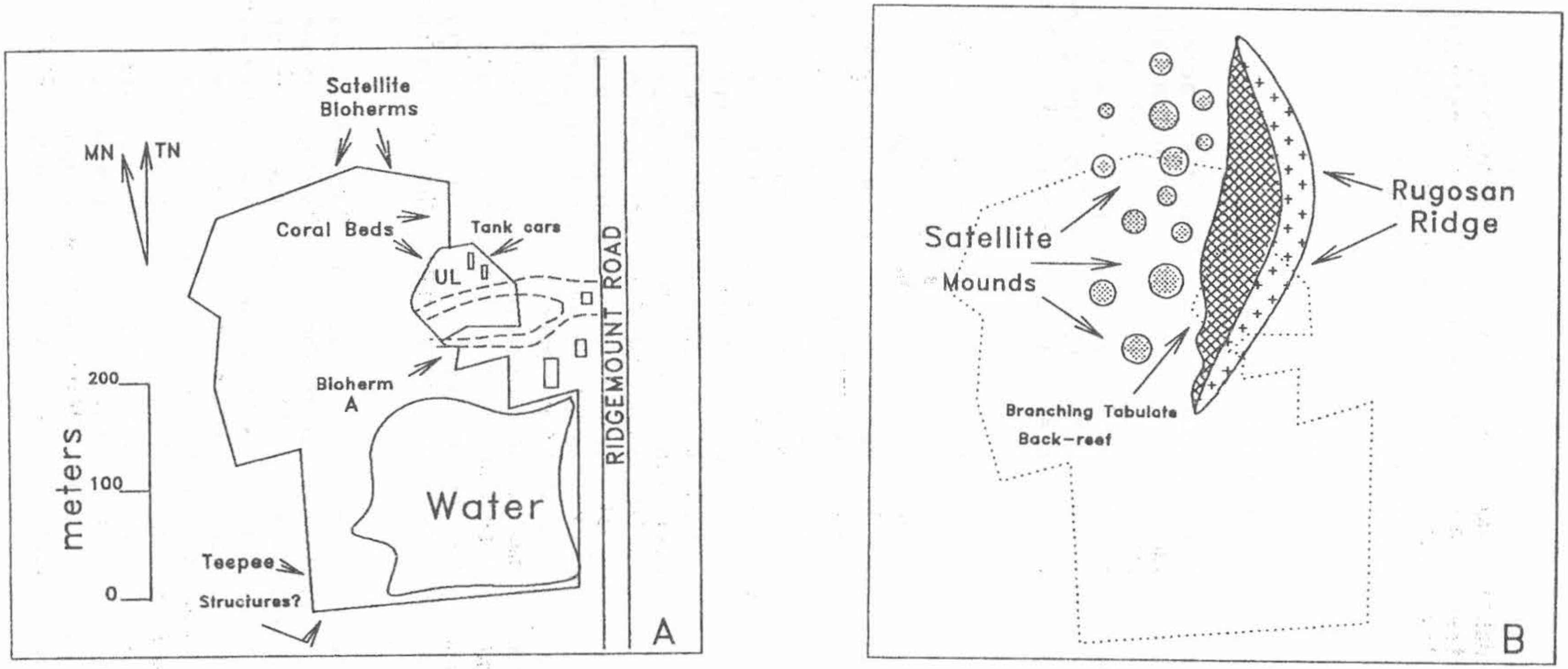


Figure 4A. Ridgemount Quarry. Coral Horizons restricted to the northeast quarter of the quarry. UL = Upper Level. Upper carbonate sand horizon is well exposed at top of quarry wall above Coral Beds and Bioherm A. Siluro-Devonain contact well exposed in western quarry wall and in floor of quarry near possible Teepee Structures. 4B. Interpretation of Ridgemount Coral buildup. Note outline of present quarry (dots). Rugosan Ridge is mainly debris pile formed through coalescence of small mounds. Branching tabulate back-reef presently exposed in quarry walls (Coral Beds, Fig. 4A). Presence of satellite mounds extrapolated from location of Satellite Bioherms in north quarry wall.

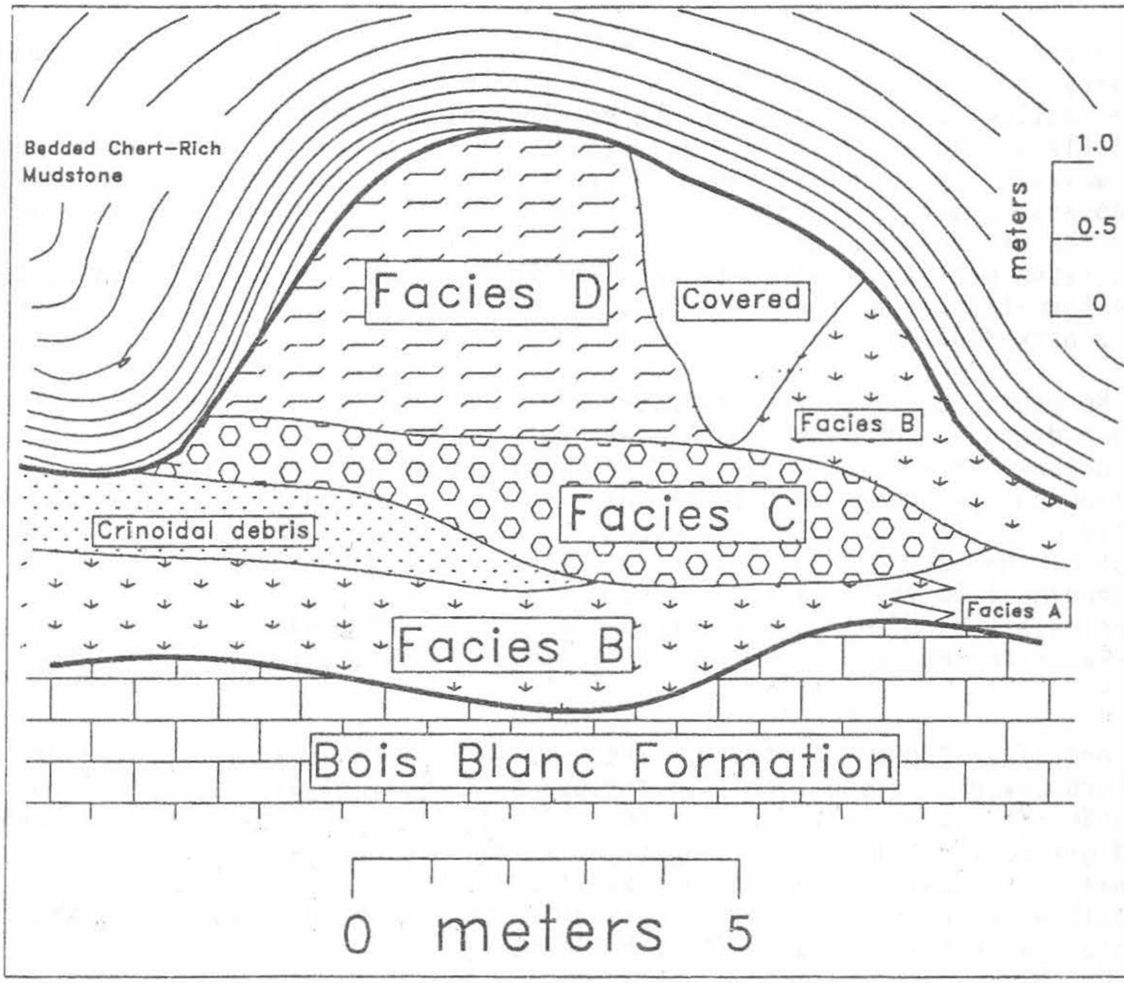


Figure 5. Bioherm A. Facies A - biostromal packstone with clasts of shale and underlying Bois Blanc Formation. Facies B - Acinophyllum rich biostrome with abundant small favositids and solitary rugose corals. Facies C - Acinophyllum, large phaceloid colonial rugosans and large solitary rugosans. Some possibly in-place. Facies D - Mainly coral debris, including Acinophyllum colonies, large branching tabulate corals and solitary rugosans. Most are overturned and fragmental. Overlying bedded chert-rich mudstone contains abundant large crinoid columnals characteristic of theEdgecliff. Dip of these beds has been exaggerated by compaction.

development seen elsewhere in the Edgecliff, and resulted in the debris-mound facies seen in this quarry.

A proposed model of the coral mounds and beds present in the Ridgemount Quarry is presented in Figure 4B. In this model, initial colonization of the shallow sea-floor lead to the development of small debris mounds (little in-place growth is preserved due to the battering of these mounds by waves). These debris mounds acted as sites for further colonization by rugosans, and as protection for tabulate corals colonizing the sea-floor to the lee of these mounds. Gaps between mounds would fill with storm generated debris, and then be recolonized by coral, eventually leading to the formation of an elongate "Coral Ridge", with small satellite bioherms in a back-ridge biostrome.

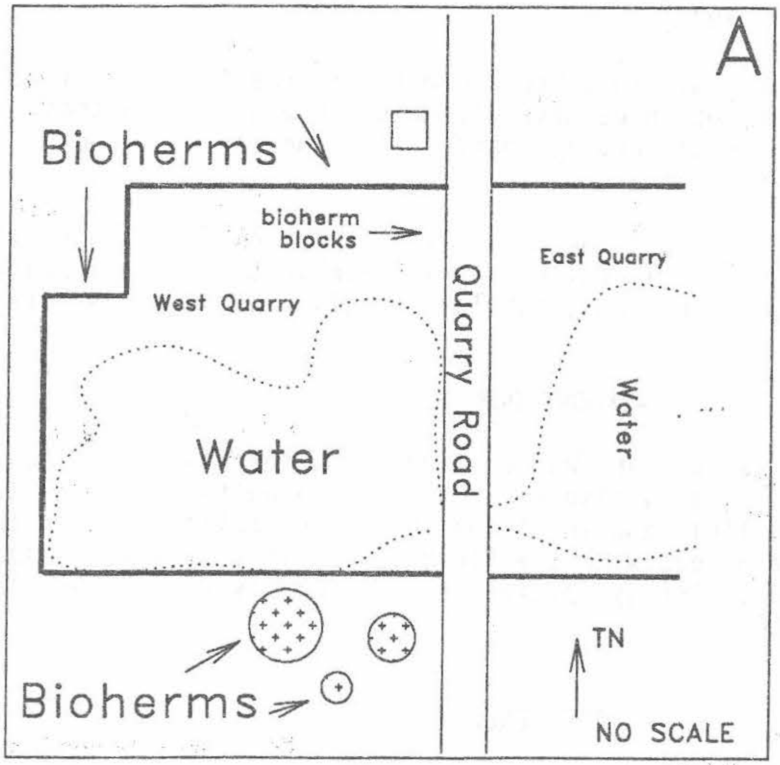
No other Edgecliff reef exposure exhibits such extensive evidence of storm damage, suggesting a shallower water depth than any other known structure. Further, evidence of the early growth of this structure following transgression over the exposed Bois Blanc (and therefore shallow water conditions) can be seen along the north wall of the quarry, to the west of the last satellite bioherm. There, the carbonate sand facies extending from the satellite bioherm interfingers with the basal "Springvale Sand", indicating that the growth of the satellite mound was coeval with the deposition of the "Springvale" only a few meters to the west.

Ancillary Topics at Stop 1. The west wall of the quarry exposes the Siluro-Devonian disconformity described by Kobluk, et al. (1977). This contact is characterized by borings which have been filled by quartz sand and glauconite. The underlying Silurian dolomite contains brecciated horizons. Excellent samples are available for class use. Finally, possible teepee structures are present in the Bois Blanc just below the contact with the Onondaga (see map Figure 4A).

Quarry Road Bioherm (Stop 2).

The West Quarry (Figure 6A) offers examples of two types of Edgecliff Reefs. Along the north wall two small, mono-generic mounds and some quarried blocks of mound facies are available for examination. The easternmost mound (and the bioherm blocks) consist of *Acinophyllum* colonies, while the western mound is predominantly *Syringopora*. Unlike the bioherms at Ridgemount these small structures exhibit dense, in-place growth of colonial coral.

At the south end of the quarry, exposed on a glaciated surface, are three small thicket/bank structures (Figure 6B) the largest of which is approximately 26 meters in diameter. These small mounds are made up of thickets of phaceloid colonial rugosans interbedded with biostromal beds of small branching tabulate corals. Again, compare the in-place preservation of coral in these small structures to the debris mounds at the Ridgemount Quarry, keeping in mind that we are now roughly 3 kilometers down depositional dip from the Ridgemount structures. Also note that these structures are quite small when compared to other thicket/bank reefs such as the roughly 245 meters diameter North Coxsackie Reef south of Albany and



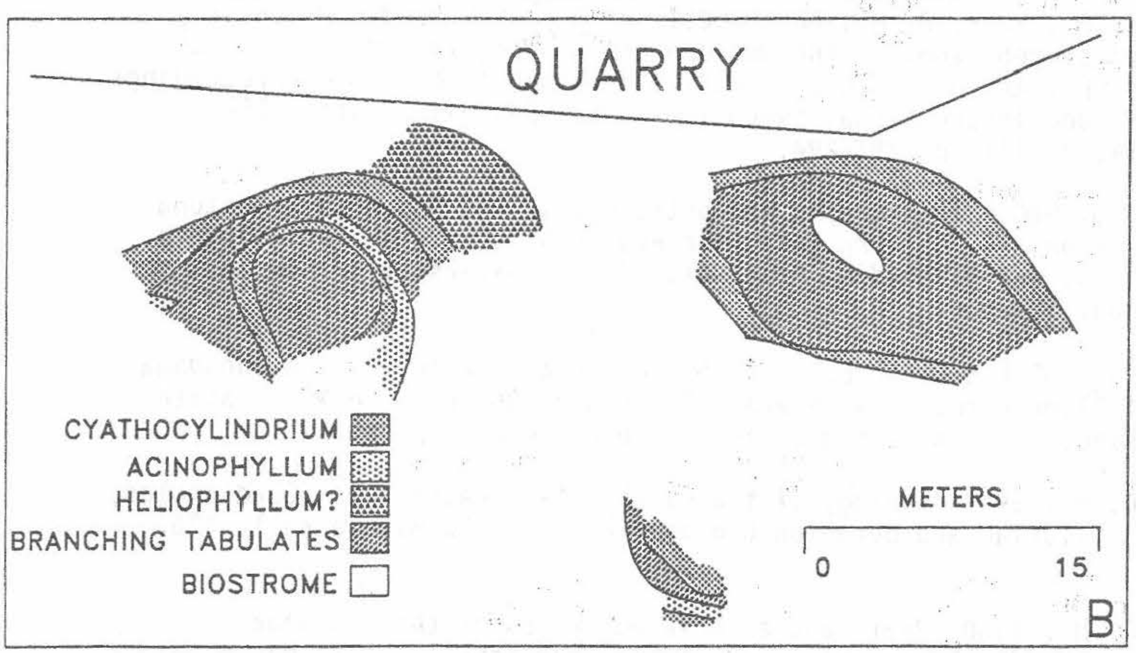


Figure 6A. Quarry Road Quarries.
6B. Detail of three small thicket/bank structures exposed at the southern end of the West Quarry.

the approximately 300 meters diameter Buffalo Country Club Reef (bioherm #'s 3 and 28 of Oliver, 1976).

Finally, examine the grainstone bed along the northeast and east wall of the West Quarry. Stromatoporoids have become quite abundant in this facies, and can also be occasionally found in the underlying, muddy *Cystiphylloides* biostrome.

Ancillary Topics at Stop 2. The Bois Blanc Formation floors both the East and West Quarries. Fossil collecting is excellent, with the quarry floor littered with loose rugosans, tabulate corals, and other fossils.

ACKNOWLEDGMENTS

The author would like to thank Walker Industries for allowing access to the Ridgemount Quarry. Thanks also to Douglas E. Paquette for his valuable assistance both in the field and in discussions of Edgecliff reefs. Thomas Mooney and James Quinn helped with the field work. This work was supported by The U.S. Department of Energy Special Research Grants Program Grant #DE-FG02-87ER13747.A000.

REFERENCES

BLODGETT, R. B., ROHR, D. M., AND BOUCOT, A. J., 1988, Lower Devonian gastropod biogeography of the western hemisphere, *in* McMillan, N. J., Embry, A. F., and Glass, D. J., eds., Devonian of the World, Proceedings of the Second International Symposium on the Devonian System, CSPG Memoir 14, v. III, p. 281-294.

CASSA, M. R., 1980, Stratigraphy and petrology of the Onondaga Limestone (Middle Devonian), Eastern Lake Erie Region of New York, Pennsylvania, and Ontario: Unpublished M. A. thesis, State University of New York at Binghamton, 108 p.

CASSA, M. R., AND KISSLING, D. L., 1982, Carbonate facies of the Onondaga and Bois Blanc Formations, Niagara Peninsula, Ontario, *in* N. Y. State Geol. Assoc., 54th Ann. Mtg., Field Trip Guidebook, p. 65-97.

CHADWICK, G. H., 1944, Geology of the Catskill and Kaaterskill Quadrangles, Part II: Silurian and Devonian Geology: N. Y. State Museum Bull. 336, 251 p.

COUGHLIN, R. M., 1980, Reefs and associated facies of the Onondaga Limestone (Middle Devonian), West Central New York: Unpublished M. A. thesis, State University of New York at Binghamton, 194 p.

GOLDRING, W. AND FLOWER, R. H., 1942, Restudy of the Schoharie and Esopus Formations in New York State: Amer. Jour. Sci., v. 240, p. 673-694.

KISSLING, D. L., 1987, Middle Devonian Onondaga pinnacle reefs and bioherms, Northern Appalachian Basin *in* Second International Symposium on the Devonian System, Calgary, Alberta, Canada, Program and Abstracts, p. 131.

KISSLING, D. L., AND COUGHLIN, R. M., 1979, Succession of faunas and frameworks in Middle Devonian pinnacle reef of south-central New York: Geol. Soc. Amer., Abstr. with Program, v. 11, no. 1, p. 19.

KOBLUK, D. R., PEMBERTON, S. G., KAROLYI, M., AND RISK, M. J., 1977, The Siluro-Devonian disconformity in southern Ontario: Bull. Can. Petrol. Geol., v. 25, p. 1157-1187.

KOCH, W. F., II, AND BOUCHT, A. J., 1982, Temperature fluctuations in the Devonian Eastern Americas Realm: Jour. Paleo., v. 56, p. 240-243.

LEES, A., 1975, Possible influences of salinity and temperature on modern shelf carbonate sedimentation: Mar. Geol., v. 19, p. 159-198.

LINDEMAN, R. H., 1988, The LeRoy Bioherm, Onondaga Limestone (Middle Devonian), Western New York, *in* Geldsetzer, H. H. J., James, N. P., and Tebbutt, G.E. (eds.), Reefs - Canada and Adjacent areas: CSPG Memoir 13, p. 487-491.

LINDHOLM, R. C., 1967, Petrology of the Onondaga Limestone (Middle Devonian), New York: Unpublished Ph. D. dissertation, Johns Hopkins University, 188 p.

MESOLELLA, K. J., 1978, Paleogeography of Some Silurian and Devonian reef trends, Central Appalachian Basin: AAPG Bull., v. 62, p. 1607-1644.

NELSON, H. F., BROWN, C. W., AND BRINEMAN, J. H., 1962, Skeletal limestone classification, *in* Ham, W.E., ed., Classification of Carbonate Rocks, A Symposium: AAPG Memoir 1, p. 224-252.

OLIVER, W. A., JR., 1976, Noncystimorph colonial rugose corals of the Onesquethaw and Lower Cazenovia Stages (Lower and Middle Devonian) in New York and adjacent areas: U. S. Geol. Survey Prof. Paper no. 869, 156 p.

OZOL, M. A., 1963, Alkali reactivity of cherts and stratigraphy and petrology of cherts and associated limestones of the Onondaga Formation of Central and Western New York: Unpublished Ph. D. dissertation, Rensselaer Polytechnic Institute, 228 p.

POORE, R. Z., 1969, The LeRoy Bioherm: Onondaga Limestone (Middle Devonian) western New York: Unpublished M. S. thesis, Brown University, 69 p.

RICKARD, L. V., 1975, Correlation of the Silurian and Devonian Rocks in New York State: N. Y. State Museum and Science Service, Map and Chart Series, no. 24.

WILLIAMS, L. A., 1980, Community succession in a Devonian patch reef (Onondaga Formation, New York) - Physical and Biotic Controls: Jour. Sed. Petrol., v. 50, p. 1169-1186.

WOLOSZ, T. H., 1989a, Water turbulence - the controlling factor in colonial rugosan successions within Edgecliff reefs: Geol. Soc. Amer., Abstr. with Programs, v. 21, no. 2, p. 77.

WOLOSZ, T. H., 1989b, Thicketing events - a key to understanding the ecology of the Edgecliff reefs (Middle Devonian Onondaga Formation of New York): Geol. Soc. Amer., Abstr. with Programs, v. 21, no. 2, p. 77.

WOLOSZ, T. H., 1988, The LeRoy Bioherm: a reactivated Devonian reef (Edgecliff Member, Onondaga Formation): Geol. Soc. Amer., Abstr. with Programs, v. 20, no. 1, p. 80.

WOLOSZ, T. H., 1985, Roberts Hill and Albrights Reefs: faunal and sedimentary evidence for an eastern Onondaga sea-level fluctuation, *in* N. Y. State Geol. Assoc., 57th Annual Meeting, Field Trip Guidebook, p. 169-185.

WOLOSZ, T. H., AND PAQUETTE, D. E., 1988, Middle Devonian Reefs of the Edgecliff Member of the Onondaga Formation of New York, *in* McMillan, N.J., Embry, A.F., and Glass, D.J., eds., Devonian of the World, Proceedings of the Second International Symposium on the Devonian System, CSPG Memoir 14, v. II, p. 531-539.

WOLOSZ, T. H., 1984, Paleoecology, sedimentology, and massive favositid fauna of Roberts Hill and Albrights Reefs: Unpublished Ph. D. dissertation, State University of New York at Stony Brook, 391 p.

ROAD LOG FOR SHALLOW WATER REEFS OF THE MIDDLE DEVONIAN
EDGECLIFF MEMBER OF THE ONONDAGA LIMESTONE,
PORT COLBORNE, ONTARIO, CANADA

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Toll both at Peace Bridge.
0.75	0.75	Canadian Customs Booths. Bare to right leaving customs booths.
0.95	0.2	Fort Erie Exit. Turn right at end of exit ramp.
1.35	0.4	Right turn at light onto Route 3.
5.65	4.3	Right turn onto Ridgemount Road.
6.45	0.8	STOP 1. Entrance to Ridgemount Quarry on left side of road. Park at quarry entrance. HARDHATS REQUIRED. Please beware of overhangs and loose rock. NOTE: Visitors should sign in at the quarry office 0.8 miles to the north.
7.25	0.8	Return to Route 3. Turn right.
22.95	15.7	Pass through town of Port Colborne. Turn right onto Quarry road.
23.25	0.3	STOP 2. Abandoned quarries on both sides of road. Park along side of road.

A FRESH LOOK AT THE SEDIMENTS WITHIN THE TERRACES AND CLIFFS THAT PARALLEL THE LAKE ERIE SHORE IN ERIE COUNTY, PENNSYLVANIA

DAVID THOMAS and RAYMOND BUYCE
(Mercyhurst College)

INTRODUCTION

The lake plain adjoining the coastal bluffs of Lake Erie in Pennsylvania is comprised of two to three major southwest-northeast trending terrace systems. They form a stair-step arrangement of very gently sloping terraces each bounded on the north by slightly to precipitously steeper escarpments or cliffs. The edges of these terraces have been referred to as "beach ridges" by several early workers (e.g. Leverett, 1892). Extensions of the terrace margins in Ohio and elsewhere are actually ridge-like although their beach origin has been questioned. Each relatively flat terrace level is reliably attributed to deposition associated with a corresponding lake level of a Pleistocene precursor to Lake Erie. It is still an open question whether the environment of deposition for any particular terrace or part thereof can be justifiably ascribed to beach deposition or whether a braid-plain (sandar), delta, or some other setting might have been responsible. Some of the early work relied heavily upon topographic expression to seek answers to the questions of origin but on this trip we will have the opportunity to look at two fine exposures within the terraces to see if the sedimentological evidence is more definitive.

Exposed in the gravel pit at the first stop is the uppermost of the two terraces present along the southeastern part of the Harborcreek 7 1/2 minute quadrangle. The upper terrace is attributed to sediment deposition associated with Lake Whittlesey (elevation: 740 feet, age: 13,000 yr B.P.).

Our second stop is on the shore of the modern Lake Erie where the upper approximately 70 feet of the bluff exposes the sands and gravels which underlie the lowermost of the two terraces. These sediments are ascribed to deposition associated with Lake Warren which was both lower and later (elevation: 675. age: 12,000-13,000 yr B.P.) Also exposed underlying the sands and gravels in the bluff at this location is more than 100 feet of glacial till (periglacial? diamict).

To enable you to better orient yourself geologically a brief description of the geology of the field trip area is provided below.

PHYSIOGRAPHY

The major physiographic divisions within the field trip area are, from north to south, the Eastern Lake Section of the Central Lowland Province and the Glaciated Section of the Appalachian Plateaus Province (Figure 1).

The Eastern Lake Section of the Central Lowland Province occupies a band 2 to 5 miles wide extending from Lake Erie to the base of an escarpment that rises southward onto the Glaciated Section of the Appalachian Plateau. It consists of three major east-west trending Pleistocene lake terraces delimited by gentle escarpments and the northernmost precipitous Lake Erie bluff. The terraces are dissected by streams that flow northward off the plateau to the lake. Abandoned stream channels with less than 3 m (10 ft) of relief also cut the terraces.

The Glaciated Section of the Appalachian Plateaus Province consists of relatively flat-lying beds of sedimentary rock that have been erosionally truncated to form a north-facing escarpment with up to 60 m (200 ft) of local relief at the northwest margin. The plateau and escarpment are covered by glacigenic deposits of irregular thickness and distribution.

The courses of most major streams draining the Plateau in the area begin by flowing westward, then flow northward for a short distance and then westward again until they reach the escarpment. They then flow northward off the escarpment, through the lake terraces and into Lake Erie.

BEDROCK GEOLOGY

The oldest rocks exposed at the surface in northwestern Pennsylvania are those along the Lake Erie shore. The westernmost 2 miles of the shoreline are underlain by the Girard Shale. The Northeast Shale underlies the remainder of the shoreline up to the New York State line (Berg and others, 1980). Both are Upper Devonian in age and correlate with portions of the Chagrin Shale to the west and the Brallier Formation, Bradford Group, and Catskill Formation elsewhere in Pennsylvania (Berg and others, 1986). The regional dip is gentle and to the south.

The Northeast Shale is a series of alternating gray shales and thin layers of gray siltstone and fine-grained sandstone. Fossils are uncommon in this unit in Pennsylvania, but fucoids have been noted (Tomikel and Shepps, 1967). Cone-in-cone structures have also been documented in these rocks. The maximum thickness of the Northeast Shale in eastern Erie County, PA is about 475 feet. The overlying Girard Shale is an

FIGURES

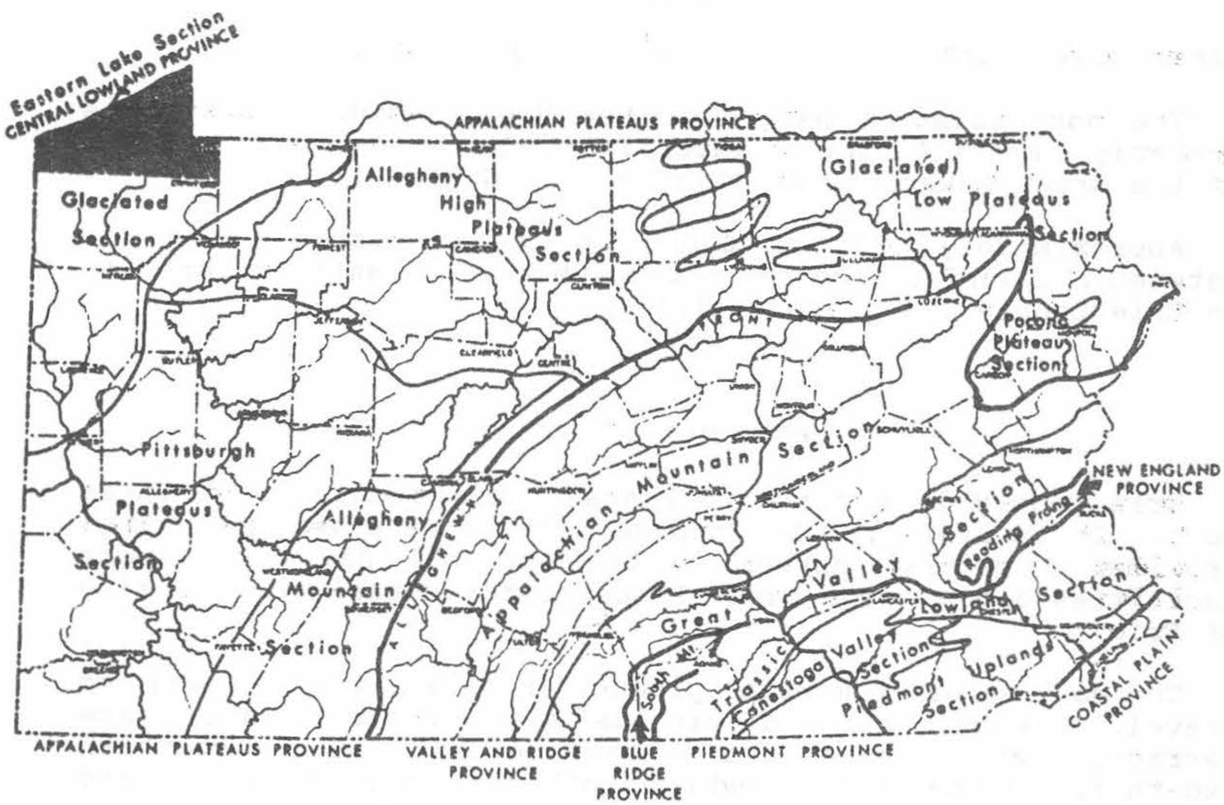
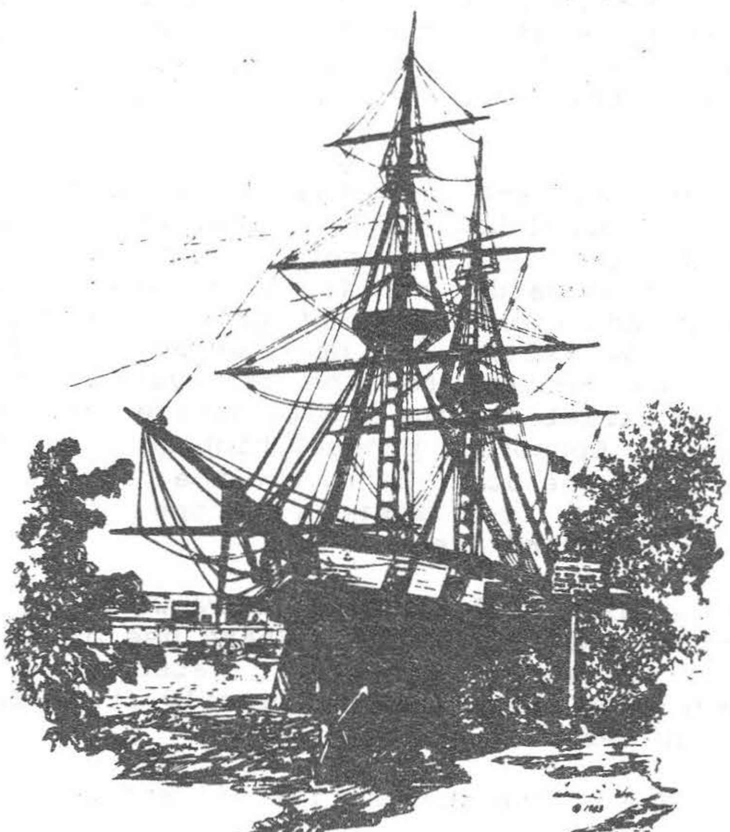


Figure 1. Physiographic provinces of Pennsylvania and location of field trip area (Erie County with diagonal pattern).



ashen gray, flaky shale with rare marine fossils.

The non-resistant nature of the Upper Devonian shales is probably a major factor in determining the locations and extent of the Great Lakes basins (Figures 2 and 3).

Approximately 1830 m (6,000 ft) of Paleozoic rocks lie between the surface and the metamorphosed Precambrian basement in Erie County, PA (Lapham, 1975).

ECONOMIC GEOLOGY

Erie County lays claim to Pennsylvania's only Great Lakes port. It is protected by Presque Isle, a recurved spit which provides a natural harbor in Presque Isle Bay. The port facilities are a large part of the economic base for the city of Erie.

The most valuable mineral product in Erie County is sand and gravel. Numerous pits of various sizes exploit kame, kame terrace, glaciodeltaic, and outwash deposits in the southern two-thirds of the county, and glaciofluvial, glaciodeltaic, and glaciolacustrine sediments at Lake Wittlesey, Arkona, and Warren I and II levels in the northern third. These materials are used for road and building construction, concrete block manufacture, and as nourishment material for the beaches of Presque Isle. In addition, sand is dredged from Lake Erie for concrete and masonry sand. The total production in Erie County, PA exceeds 500,000 tons/year (Tomikel and Shepps, 1967). The field trip will visit one pit at the Lake Whittlesey level.

The well-drained sand and gravel soils along the eastern Lake Section of the Central Lowland Province, which is commonly referred to as the Lake Plain, provide excellent soil for nursery stock, fruit orchards, and fruit and vegetable farms. This soil, plus 15 additional days of growing season in the fall due to the slow release of heat energy by the lake, provides an ideal environment for the cultivation of grapes. There are two major varieties. The fruity native North American grapes, most commonly Concords, produce high quality jelly and juice. Most wines, on the other hand, are the product of vines that have been brought from western Europe. We will be travelling through vineyard areas.

Lake Erie has a plentiful, high quality water supply for adjacent communities and industries. Where use of lake water is not practical, industries and public water supplies must derive their water from subsurface, primarily from glacigenic sand and gravel aquifers.

Erie County is one of the most active oil and gas producing

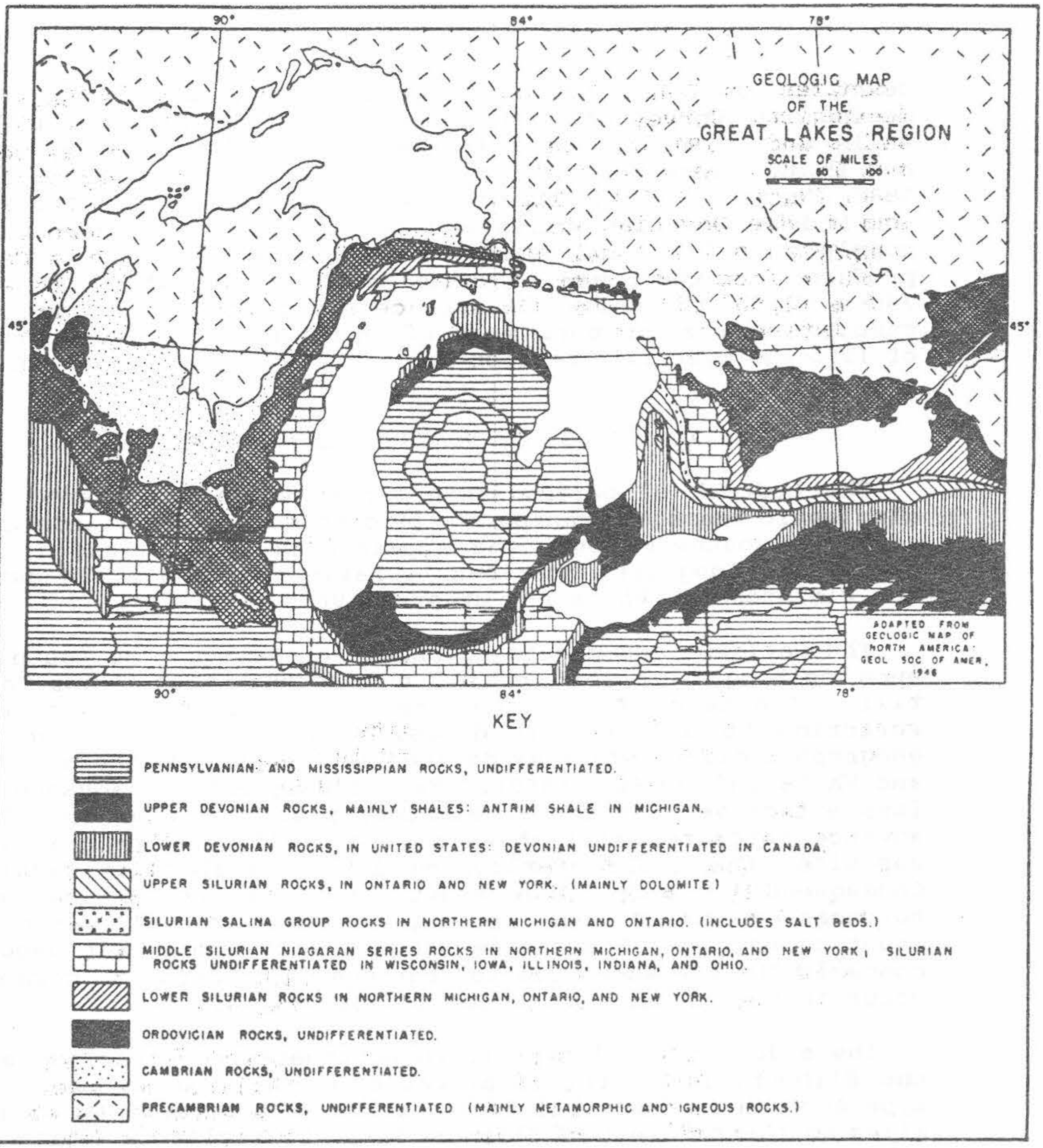


Figure 2. Geologic map of the Great Lakes Region (from Hough, 1958).

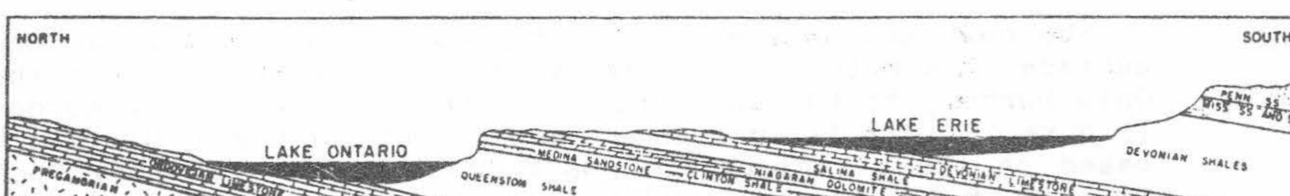


Figure 3. Cross section through the western end of the Ontario basin and the eastern end of the Erie basin, showing relations of the basins to weak shales (from Hough, 1958).

counties in Pennsylvania. In June 1987, the Pennsylvania Geological Survey had well records on file for 406 shallow wells and 2,191 deep wells in the county. Most are gas wells, but 15,825 barrels of oil were produced from 121 deep wells in 1986 (Harper, 1987). The shallow wells produce from the Upper and Middle Devonian shales and are typically non-commercial, supplying a few local users. Most deep wells in Erie County produce from the Lower Silurian Medina Group, which is about 760 m (2,500 ft) below the surface along the lake shore. The cumulative total deep gas production for Erie County at the end of 1985 was over 118 trillion cubic feet of gas (Harper, 1986).

GLACIAL GEOLOGY OF NORTHWESTERN PENNSYLVANIA

The bedrock of northwestern Pennsylvania is covered with glacigenic deposits transported by continental ice sheets which advanced southward onto the Appalachian Plateau and to the southwest along the Ontario-Erie Basin numerous times during the Pleistocene (White and others, 1969).

The earliest ice sheets advanced farthest to the south and upon melting back left behind the southernmost deposits of till. The term till is used here in a stratigraphic sense referring to all glacial deposits of a specified age and geographic distribution as defined in Shepps and others (1959) and White and others (1969). Each subsequent ice advance was less extensive than the immediately preceding advance. Each advance deposited till that only partially covers the older deposits, the whole overlapping in a shingle-like fashion. Consequently, each till sheet outcrops in an elongate northeast-southeast band roughly parallel to its former ice front or margin. The overlapping of each subsequent deposit conceals the preceding drifts, but the older deposits commonly occur in the subsurface (White and others, 1969).

The oldest glacial deposit in northwestern Pennsylvania is the Slippery Rock Till of probable pre-Illinoisan age. The type section is described from a subsurface occurrence about 3 miles north-northwest of Slippery Rock, Pennsylvania (White and others, 1969). The presumed age is based upon its stratigraphic position below the Mapledale Till.

The next youngest unit, the Mapledale Till, occurs at the surface in a belt 1 to 5 miles wide from Beaver County at the Ohio border, to Warren County near the New York State border (Figure 4). The Illinoisan age assignment for the Mapledale is based on its stratigraphic position above the Slippery Rock Till and below the Titusville Till, and because of the degree of weathering of a paleosol on its surface.

The Titusville Till is exposed at the surface along a belt 0 to 10 miles wide extending from northern and western Warren County at the New York border to the Ohio border in

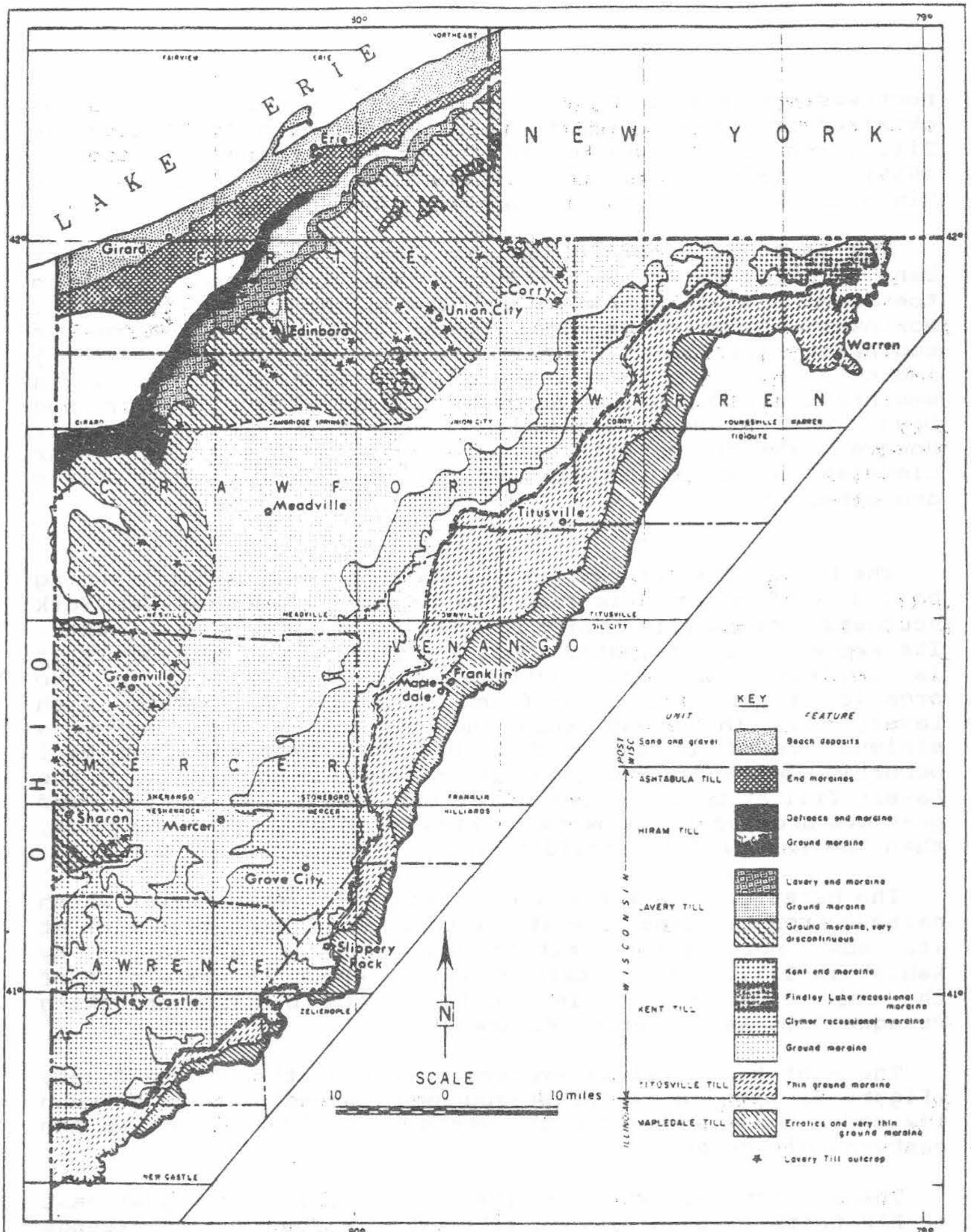


Figure 4. Glacial map of northwestern Pennsylvania showing distribution of Illinoian and Wisconsinan tills, ground moraine, recessional moraine, end moraine, and sand and gravel beach deposits. Neither the Slippery Rock Till nor the Girard moraine are shown on this map (from White and others, 1969).

northwestern Beaver County (Figure 4). Radiocarbon dates obtained from peat deposits in gravel beneath the Titusville Till range from 35,000 to 40,500 yr B.P. (White and others, 1969). This places the Titusville Till within middle Wisconsinan time (Dreimanis and Goldthwait, 1973).

The Kent Till deposits extend along a northeast-southwest band 30 miles wide from the Ohio border in Lawrence County to the New York State border in northern Warren County (Figure 4). Northwest-southeast oriented linear subglacial landforms in southeast Erie County and northern Crawford County (Figure 5) indicate that the movement of the Kent ice was in a southeasterly direction. No organic material for ^{14}C dating has been found in Pennsylvania in association with Kent Till. However, lacustrine material associated with Kent Till near Cleveland, Ohio has yielded a ^{14}C date of 24,000 yr B.P. (White and others 1969).

The Lavery Till is exposed in a northeast-southwest trending belt 1 to 5 miles wide from the western border of New York southeast across Erie County into northwestern Crawford County. Its exposure is terminated about 10 miles east of Ohio where it is completely overlapped by the Hiram Till (Figure 2). No organic material has been found in direct association with Lavery Till in Pennsylvania. However, a ^{14}C date having a minimum age of 14,000 yr B.P. has been obtained from marl occurring beneath a peat deposit in front of the margin of the Lavery Till at Corry (White and others, 1969). If the marl and peat are proglacial deposits related to ablation of Lavery ice, then the Lavery Till should be older than 14,000 yr B.P.

The Hiram Till is exposed in a triangular-shaped area which narrows from 14 miles wide at the Ohio border to 1 mile wide at its most northeastern extent where it is overlapped by Ashtabula Till. The eastern extent is about 23 miles east of the Ohio line and 5 miles south of the bluff overlooking Presque Isle Bay in Erie (Figure 2).

The Ashtabula Till is exposed across northern Erie County (Figure 2) along a northeast-southwest trending belt between 1/2 and 5 miles wide from the western New York border to the eastern Ohio border.

The westernmost extent of the Girard Till occurs just east of Elk Creek in Lake City, northwest Erie County. It extends eastward as a band 1/2 to 1-1/2 miles wide to about 1 mile south of the Borough of North East where it appears to coalesce with the Astabula Till.

Except for the Titusville, each of the Wisconsinan-age tills has an identified end moraine (Figure 2). The end moraine associated with the Hiram Till is named the Defiance end moraine. Each of the other end moraines takes the same name as

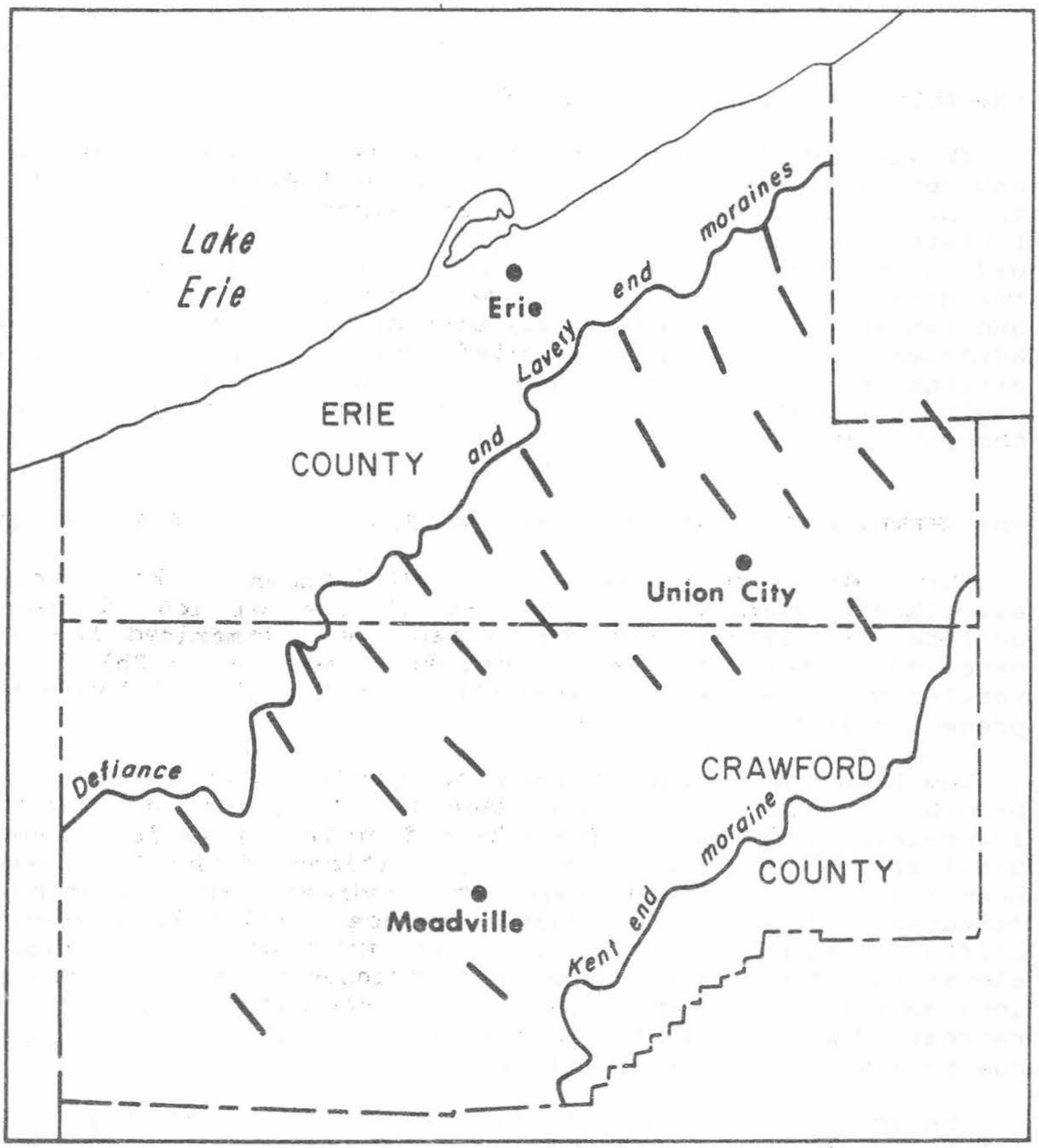


Figure 5. Orientation of linear landforms in northwestern Pennsylvania. Each linear-trend line represents an average of up to 40 orientation measurements. Orientation measured on 1:50,000 scale topographic maps of Erie and Crawford Counties. Only hilltop landforms were measured. Presumably the landforms represent both constructional and erosional features, but no field verification has been done. Note that the linear landforms occur only in a ground moraine area between the Kent and Defiance-Lavery end moraines. Figure prepared by W. D. Sevon, Pennsylvania Geological Survey, 1987.

the till with which it is associated.

In summary, it appears that there were at least 8 advances and retreats of glacial ice into northwest Pennsylvania within the pre-Illinoisan, Illinoisan, and Wisconsinan Stages of the Pleistocene. Each left behind till. Each succeeding advance did not encroach as far south as the previous advance although the Hiram ice did completely override part of the Lavery Till and Ashtabula ice moved across part of the Hiram Till. These advances and retreats resulted not only in large-scale alterations of the topography and surface drainage of the region but must have been the major excavators and shapers of the Lake Erie basin.

THE HISTORY OF PLEISTOCENE LAKE LEVELS IN THE LAKE ERIE BASIN

The following discussion of Pleistocene lake levels, associated advances and retreats of glacial ice, drainage outlets, and radiocarbon dates has been summarized from an excellent compilation by Calkin and Feenstra (1985). The version presented here is a simplification and is intended to present only the major points.

The Lake Erie basin has been occupied by a series of major proglacial ice or moraine-dammed lakes and non-glacial lower-level lake phases from the end of the Hiram ice advance until today. The positions and elevations of the lakes have been identified by their associated sediments and topographic features such as beach ridges, terraces, and relic wave-cut cliffs. Among the factors that influenced the various elevations of the lake stands were advances and retreats of ice lobe margins, opening of lower drainage channels during ice retreat, downcutting of outlet channels, and crustal warping due to glacial unloading and reloading.

The geologic history of the lakes of interest here are related to the advances and retreats of three ice lobes in two Great Lake basins. The earlier glacial ice that extended well south of Lake Erie excavated the Huron and Ontario-Erie basins. The basin topography largely controlled the movements of subsequent ice lobes that moved across the area during the time of the glacial and non-glacial lakes. The Huron ice lobe and the Saginaw ice lobe flowed within the Lake Huron basin and the Ontario-Erie lobe flowed within the Lake Ontario-Lake Erie basin. As the ice lobe margins retreated lakes occupied the vacated basins and as the margins readvanced the lakes moved in response.

The lakes have been named Lakes Maumee I (earliest), II, and III; Lake Arkona; Lake Ypsilanti; Lake Whittlesey; Lakes Warren I, II, and III; Early Lake Erie; and present Lake Erie (latest). The sediments and topographic features to be seen during the field conference are those related to Lakes

Whittlesey, Warren I, Arkona, and Lake Erie.

Lake Maumee I

The first of the proglacial Great Lakes in the Lake Erie basin developed in northwestern Ohio and northeastern Indiana following the culmination of the Hiram ice readvance and deposition of the Defiance end moraine during the Late Wisconsinan (Figure 6). Named Lake Maumee I, it stabilized at a present-day elevation of 800 feet and drained into the Wabash River through two outlets cutting the Fort Wayne Moraine. Maumee I beaches have been traced in northwestern Ohio and southeast Michigan. A ^{14}C date for the minimum age for Hiram ice recession from the Defiance Moraine is 14,500 ± 150 yr B.P. This should serve as a minimum age for Lake Maumee I.

Lakes Maumee II and III

As the ice of the Huron Lobe melted back to the north along the Lake Huron basin and the Ontario-Erie Lobe retreated northeast along the Lake Erie basin from the Defiance Moraine, a lower drainage channel was either uncovered or downcut by escaping water north of Imlay, Michigan causing abandonment of the Maumee I outlet (Figure 7). Consequently, the Maumee II and III phases were stabilized at 780 feet and 760 feet respectively. There is considerable confusion about the sequence of the two latter Maumee levels and as to the location of the drainage channel for Maumee II. Nevertheless, it appears that Maumee III stabilized, after a readvance or standstill of ice that formed the Tillsonburg Moraine in Ontario, the Ashtabula Moraine in Pennsylvania, and an as yet unlocated moraine below the present Lake Erie. A ^{14}C date for Maumee III is 13,700 ± 220 yr B.P.

Lake Arkona

Subsequent retreat of the Huron and Saginaw ice lobes northward along the Lake Huron basin and the northeast melt back of the Ontario-Erie Lobe in the Lake Erie basin to the Paris Moraine in Ontario and the Girard Moraine in Pennsylvania caused the waters in the Lake Huron and Lake Erie basins to subside and at the same time to join forming an early phase of Lake Arkona (Figure 8). This lake merged with early Lake Saginaw and drained westward via the Saginaw Bay outlet through the Grand River valley to Lake Chicago. Three different Arkona phases have been identified: Highest Lake Arkona, 710 feet; Middle Lake Arkona, 700 feet; and Lowest Lake Arkona, 695 feet. Beaches of the two higher lakes have been traced to 7.4 miles east of the Ohio-Pennsylvania border by Totten (1982; 1985). A ^{14}C date places Lake Arkona at 13,600 ± 500 yr B.P.

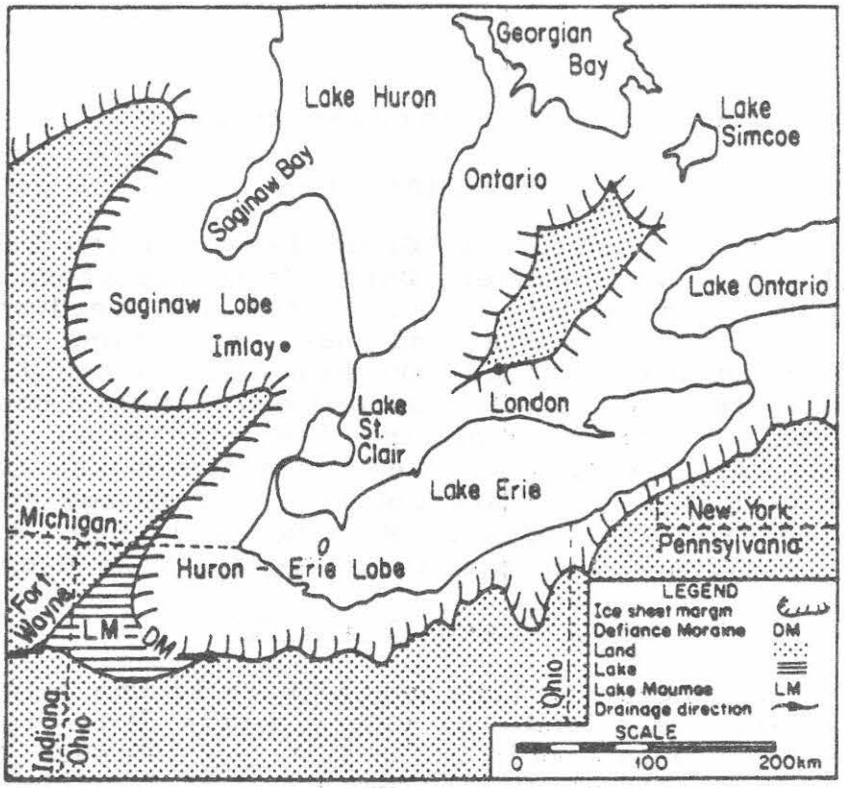


Figure 6. Locations of Maumee I, the western outlet at Fort Wayne, Indiana, and the Defiance End Moraine. Note the positions of Lakes Huron, Erie, and Ontario, and the Huron-Erie Lobes (from Calkin and Feenstra, 1985).

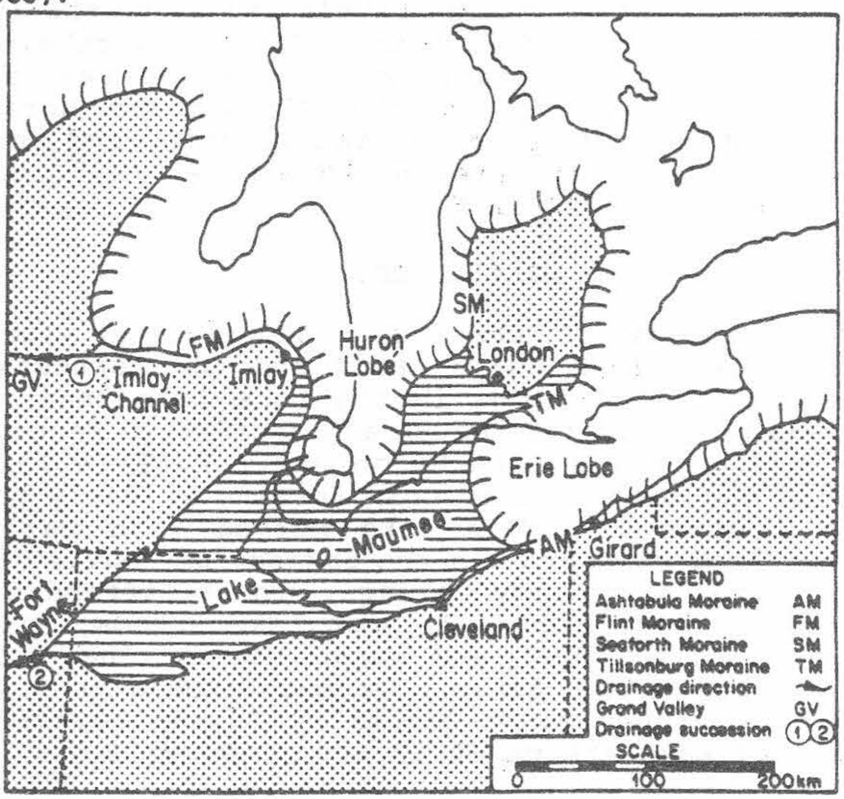


Figure 7. Locations of Lake Maumee III, the western outlet north of Imlay, Michigan, the Grand River valley crossing Michigan, and the Tilsonburg and Ashtabula end moraines. Note the positions of the Saginaw, Huron, and Erie Lobes (from Calkin and Feenstra, 1985).

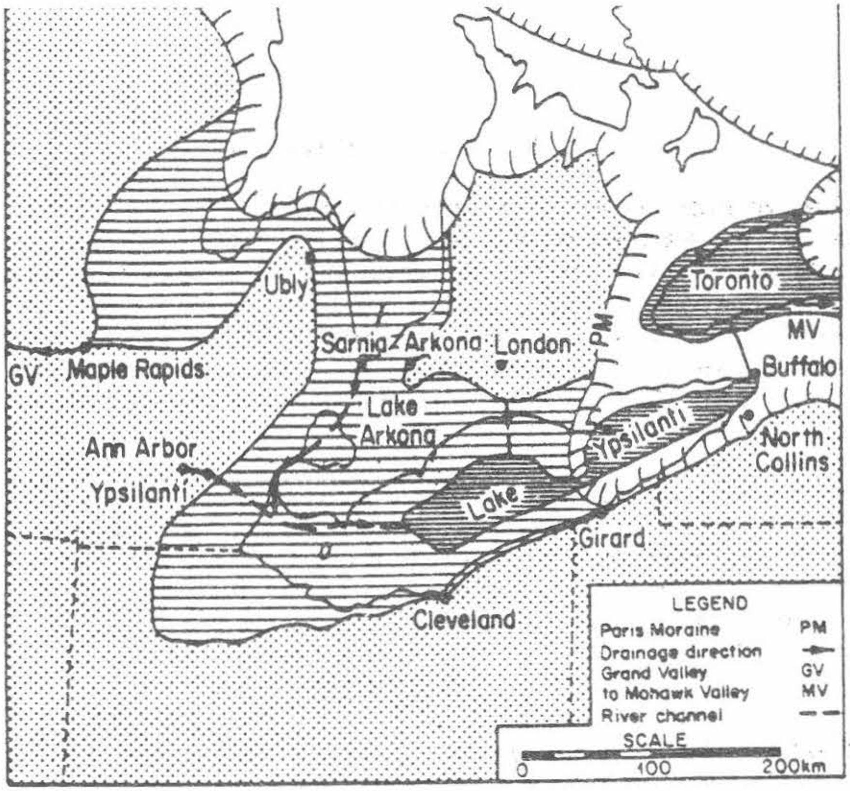


Figure 8. Locations of Lake Arkona, the western outlet through the Grand River valley in Michigan, and the Paris and Girard Moraines. Note the positions of the Saginaw, Huron, and Erie Lobes at the east end of Lake Arkona and the margin of the ice and the eastern outlet through the Mohawk River valley during the existence of Lake Arkona (from Calkin and Feenstra, 1985).

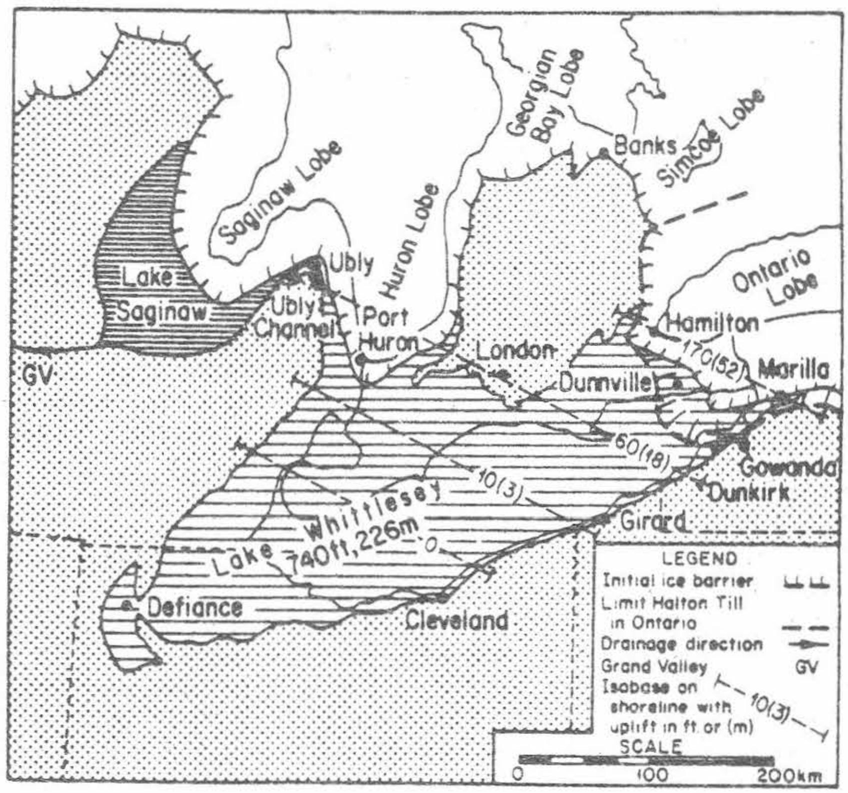


Figure 9. Locations of Lake Whittlesey, the western outlet north of Ubley, the Grand River valley in Michigan, and the Hamburg Moraine in western New York. Note the positions of the Saginaw, Huron, and Ontario Lobes (from Calkin and Feenstra, 1985).

Lake Ypsilanti

The Huron ice lobe continued to retreat to the present location of Georgian Bay while the Ontario-Erie ice lobe melted out of the Erie basin, across the Niagara Escarpment, beyond Toronto to the north, and into the northeast portion of the Lake Ontario basin. This opened an eastward drainage channel from the basins of Lakes Michigan, Huron, and Erie into a lower lake in the Ontario basin which, in turn, drained eastward through the Mohawk River valley and then through the Hudson River valley. The lower lake that developed in the Lake Erie basin was the nonglacial Lake Ypsilanti (Figure 8). Ypsilanti sediments have been found at elevations between 677 feet and 300 feet reflecting a fairly rapid lowering during this time. ^{14}C dates of organic material that relate to Lake Ypsilanti range between 12,600 ± 440 yr B.P. and 13,360 ± 440 yr B.P.

Lake Whittlesey

As the Saginaw and Huron Lobes readvanced south through the Lake Huron basin into Michigan and as the Ontario-Erie Lobe readvanced southwest within the Ontario basin across the Niagara Escarpment and into the northeastern-most part of the Lake Erie basin, waters in the Lake Erie basin rose to form Lake Whittlesey (Figure 9). Westward drainage was re-established through a spillway at Ubley, Michigan to Lake Saginaw which then drained through the Grand River valley to Lake Chicago in the Lake Michigan basin. The farthest extent of the readvance of the ice into the Erie basin is marked by the Hamburg Moraine near Buffalo, New York (Figure 10). Proglacial Lake Whittlesey stabilized against this moraine at the elevation of 740 feet. ^{14}C dates give a maximum age for Lake Whittlesey of 13,000 yr B.P.

Lake Warren

As the ice margin retreated from the Port Huron Moraine north of Ubley, Michigan (Figure 9) and from the Hamburg Moraine to the Alden Moraine south of Buffalo, New York (Figure 10), high discharges into Lake Saginaw and Lake Whittlesey produced downcutting of the western drainage channel through the Grand River valley and into Lake Chicago. This resulted in the lowering of Lake Whittlesey (740 feet) to the highest Lake Warren level (685 feet). As a result of either continued downcutting of the channel in Michigan or retreat of the ice margin in western New York State, three phases of Lake Warren were established: Warren I at 685 feet, Warren II at 675 feet, and Warren III at 670 feet. Warren II has been identified only locally. The various Warren phases were the last and most extensive of the major great glacial lakes to occupy the Erie basin (Figure 11). ^{14}C dates range from 13,050 yr B.P. on wood beneath Warren I deposits to 12,000 yr B.P. on organic material believed to be post-Warren.

Lakes Grassmere and Lundy

The northward retreat of the ice margin from the Alden Moraine (which had restricted Warren III) was interrupted by brief standstills that are reflected by the Fort Erie-Buffalo Moraine, Niagara Falls Moraine, and the Vinemount-Barre Moraine (Figure 10). The drainage at this time was eastward through the Mohawk Valley which resulted in lower-level lakes named Lake Grassmere at 640 feet and Lake Lundy at 620 feet. These lakes are represented by discontinuous and low relief (less than 2 to 6 feet) shore features.

Early Lake Erie

The final ice margin retreat that affected the Lake Erie basin caused the lake surface to fall below the elevation of the Niagara Escarpment. This initiated the first phase of modern Lake Erie known as Early Lake Erie (Figure 12). The final lowering occurred as a channel incised northward across the Fort Erie-Buffalo and Niagara Falls Moraines and down to the underlying Onondaga limestone at Buffalo, New York. This channel became the southern part of the Niagara River. The Onondaga limestone now serves as a threshold for present Lake Erie. Rather than continuing to flow north, the water spilled into the east-west trending Lake Tonawanda which subsequently spilled through drainage channels that opened across the Niagara Escarpment east of Buffalo as the ice retreated farther north into the Ontario Basin. Because of the loading of glacial ice, the Onondaga limestone threshold at that time was 120 feet below its present level and that of the present Lake Erie datum of 570 feet. This places the level of Early Lake Erie over 120 feet below present Lake Erie! ^{14}C dates from organic material found in sediments associated with Early Lake Erie places its age between 12,650 ± 170 yr B.P. and 12,080 ± 300 yr B.P.

Present Lake Erie

Glacio-isostatic uplift of the Onondaga threshold successively raised lake levels to about 12 feet below Lake Erie datum of 570 feet by 3,500 yr B.P. Continuous glacio-isostatic response of that threshold has resulted in the present datum of 570 feet (Figure 13).

Some Points of Controversy in Lake Erie Basin History

1. The eastern margin of Lake Maumee III may have been the Ashtabula Moraine (Fullerton, 1980; Totten, 1982) or the Girard Moraine (Schooler, 1974).
2. Lake Arkona may have occurred before Lake Whittlesey rather than after.
3. Lake Whittlesey and Lake Warren strands may have been

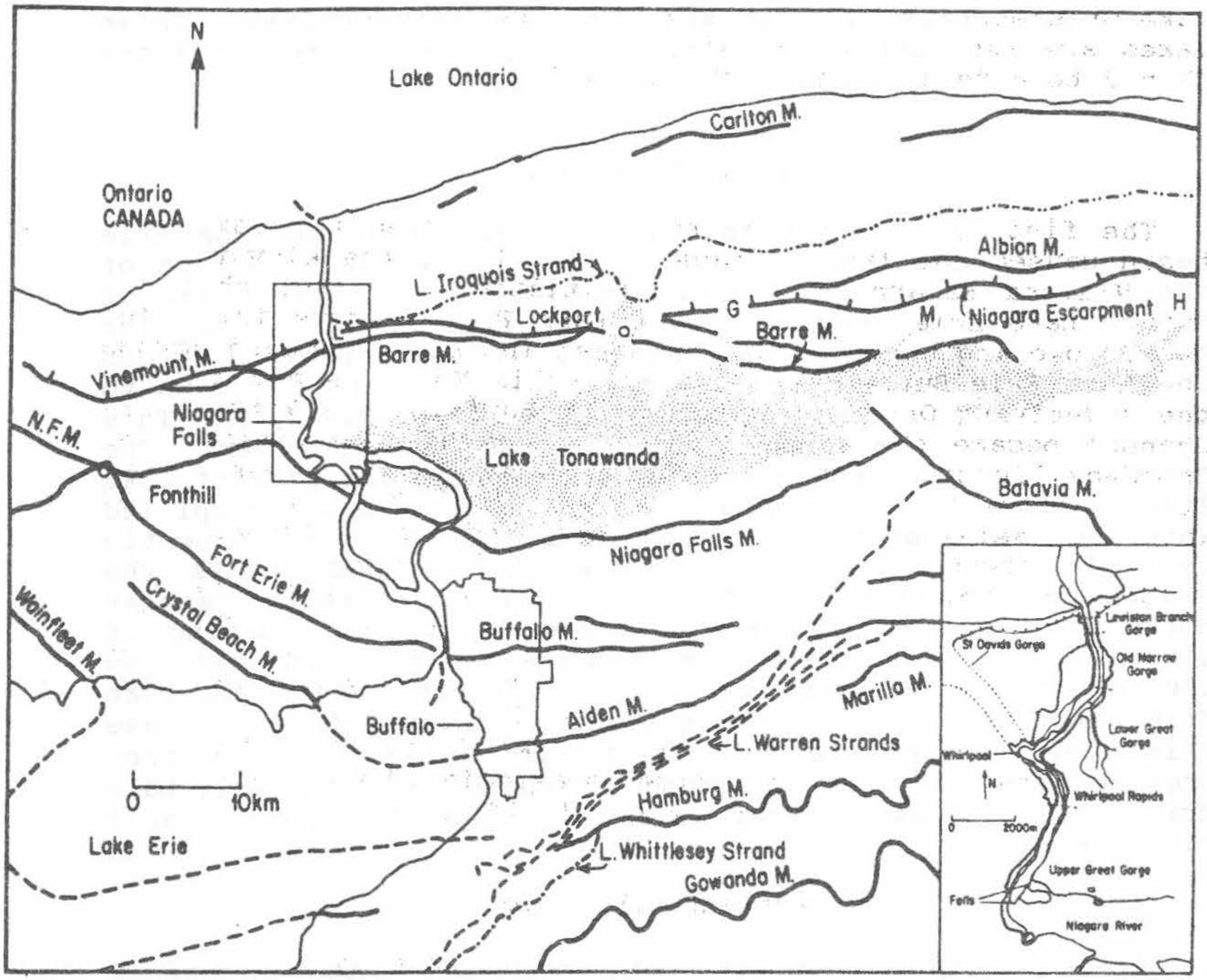


Figure 10. Locations of Lakes Whittlesey and Warren strands and the related Hamburg and Alden Moraines. Note the positions of the Buffalo-Fort Erie and Niagara Falls Moraines that were breached during the formation of Lakes Grassmere and Lundy, and of Lake Tonawanda that formed as a result of the breaching. Also note the northern outlet channels that drained Lake Tonawanda at H, M, O, and L into the Lake Ontario basin (from Calkin and Feenstra, 1985).

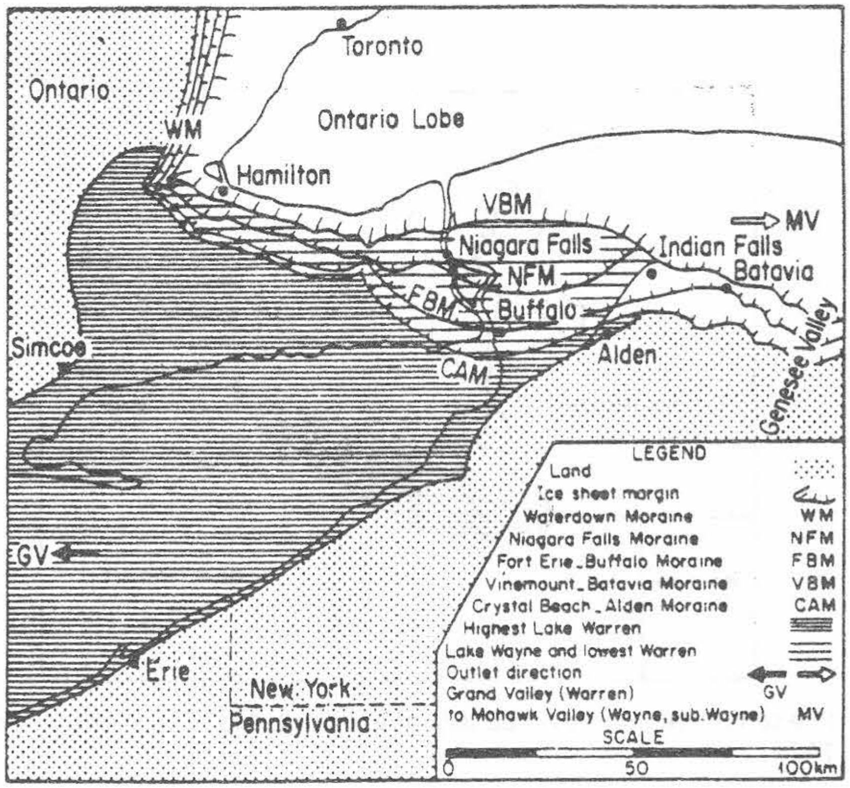


Figure 11. Location of Lake Warren and the position of the Alden Moraine (from Calkin and Feenstra, 1985).

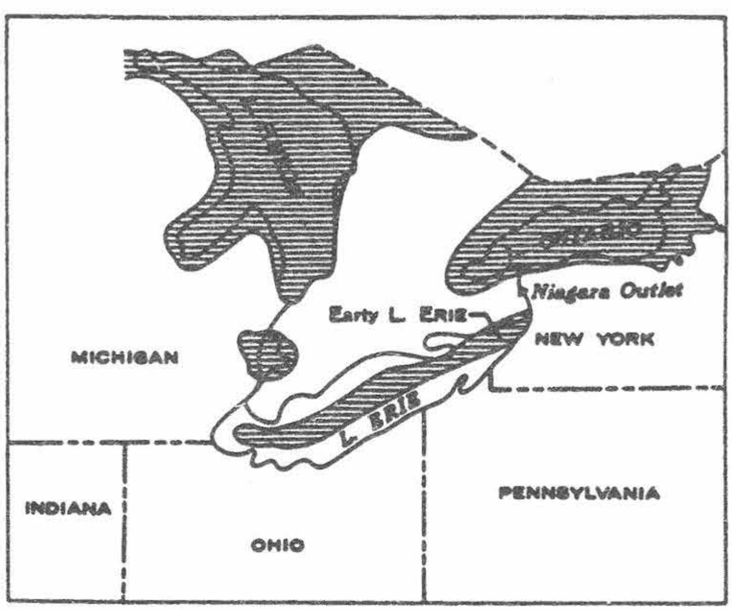


Figure 12. Location of Early Lake Erie. The Niagara River outlet was at least 120 feet lower than today. Note the possible position of the ice margin in Ontario, Canada (from Schooler, 1974).

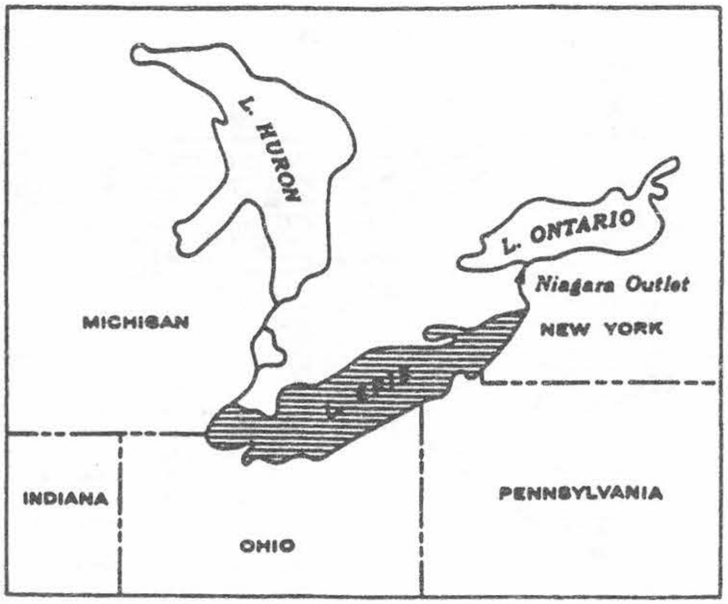


Figure 13. Present Lake Erie. Datum approximately 570 feet (from Schooler, 1974).

mapped upon sediments that are not beach deposits.

4. Lake Warren III (lowest) may have drained eastward along the Mohawk Valley rather than westward through the Grand River Valley in Michigan.
5. Some evidence suggests that Lake Erie may have been higher than its present level within the last few thousand years (Coakley and Lewis, 1985; Barnett, 1985; Larsen, 1985).

ROAD LOG AND STOP DESCRIPTIONS

The road log begins at Exit 10: Harborcreek-PA Rt. 531 of Interstate 90 in Pennsylvania.

Cumm. Miles	Miles from last point	Description
0.0	0.0	Stop sign at intersection of Harborcreek exit ramp and PA 531. Turn right (North) onto PA 531 and proceed for about .1 mile to intersection with Davison road.
0.1	0.1	Intersection of PA 531 and Davison Road. Bear right onto Davison Road. We will cross Seven-Mile creek, an essentially north-flowing creek that empties into Lake Erie. Seven-Mile refers to the distance of the creek and the city of Erie to the west, and not the length of the creek.
0.15	.05	Intersection of Davison Road and Dugan Road on the right. Continue North on Davison Road. You will be traveling across the Ashtabula Morainic System. The Girard and Ashtabula Moraines which are easily distinguished south of Erie, PA cannot be recognized at the two separate ridges east of Erie, Ashtabula till Halburn, PA.
0.6	.045	Overlook. Ahead is the steeply descending North-facing slope of the Ashtabula Moraine and the flatter lying terraces of lakes Whittlesey, Warren I-II and Warren III.
0.7	0.1	Intersection of Davison Road and McGill Road. Continue north on Davison Road. We are continuing to descend the north-facing slope of the Ashtabula ice marginal moraine.
1.4	0.7	Stop sign at intersection of Davison road and Belle road. Stop and turn right (east)

		onto Belle road. We will be traveling east on and parallel to the north facing slope of the Ashtabula Ice Marginal Moraine.
2.4	1.7	Intersection of Belle road and Rohl road to the right. Continue east on Belle road.
2.8	0.4	Intersection of Sidehill road (Belle road becomes Sidehill road at this point) and Mooreheadville road. Continue east on Sidehill road.
3.5	0.7	Intersection of Sidehill road and Brickyard road. Turn right (north). We will be descending a much gentler north slope of the Ashtabula Moraine.
4.4	0.9	Intersection of the Brickyard road and Law road. Continue north on Brickyard road. We'll remain on Ashtabula deposits.
4.9	0.5	We cross a west-flowing tributary of Twelve-mile creek. The stream has cut into stratified, subangular to subrounded cobbles with an occasional glacial erratic. The material has not been extensively worked.
5.1	0.6	Railroad tracks. There are two separate sets of tracks with two separate gates. Use extreme caution. Trains often move past here at very high speeds.
5.2	0.1	We will descend a relic "wave cut" cliff of Lake Whittlesey and proceed north across much flatter terrain. The terrain is underlain by gravel of a variety of sizes.
5.4	0.2	Intersection of Brickyard road and US Route 20. Stop sign. Turn left (west) and continue on US Route 20. To our right is the lake Warren I terrace.
5.7	0.3	The cemetery to our left illustrates the true topography. Large amounts of gravel have been quarried from both sides of the cemetery.
6.0	0.3	Entrance to Central Sand and Gravel Co. pit. Turn Left.

Stop 1. CENTRAL SAND AND GRAVEL PIT
(Stop description by: Charles Carter, Univ. of Akron from Thomas and others, 1987)

Introduction

This pit is part of an extensive (several kilometers long and up to a few kilometers wide) sand and gravel deposit that trends southwest-northeast, roughly parallel to the present Lake Erie shore (Figure 14). The existing pit is nearly mined out as are adjacent pits to the west and south. The sand and gravel are used for road construction, the mining is done with front-end loaders, and the processing is accomplished by a hydraulic sorter. The deposit is at least 8 m (26 ft) thick and mining has been done to 5 or 6 m (16 or 20 ft) below the original, prestripped surface. Foreman Scott Wassink said that the deposit was coarser to the east so that although the sediments exposed along the western face are largely granule- and pebble-sized, the overall grain size for the pit was larger.

A small, about 4 m (13 ft) long, normal, nearly bedding-plane parallel fault occurs near the center of the face. The apparent fault plane is curved and has a strike of about east-west and an overall dip of about 23°N although siltstone cobbles are oriented nearly 90° from the horizontal at the north end. Whether the origin of the fault is tectonic or isostatic is unknown.

The elevation of the surface of this deposit (about 780 to 790 ft), and the correlation of this surface both to the east and the west, has led to the mapping of this feature and elevation as the Whittlesey strandline (Schooler, 1974, p. 11, 12). The deposits underlying this surface in this area of Pennsylvania have been interpreted as "beach deposits" (most recently by Schooler, 1974, p. 19) mainly on the basis of geomorphic evidence. However, she also recognized a "channel-like feature" at one location that included imbricated gravel that she interpreted as a "delta" (p.22). Schooler's work is typical of most of the work that has been done on the "beach ridges," in that the interpretations of these deposits have been made largely on topographic expression and texture with little consideration given to the internal characteristics and geometry of the deposits. Because of this, University of Akron students and faculty are doing sedimentologic studies of the ridges in order to improve our knowledge of their origin and evolution.

Sedimentology and Paleogeography

The pit lies near the western apex of a triangular-shaped terrace that opens to the east (Figure 15). A north-south face with a length of about 350 m (1148 ft) and a mean height of

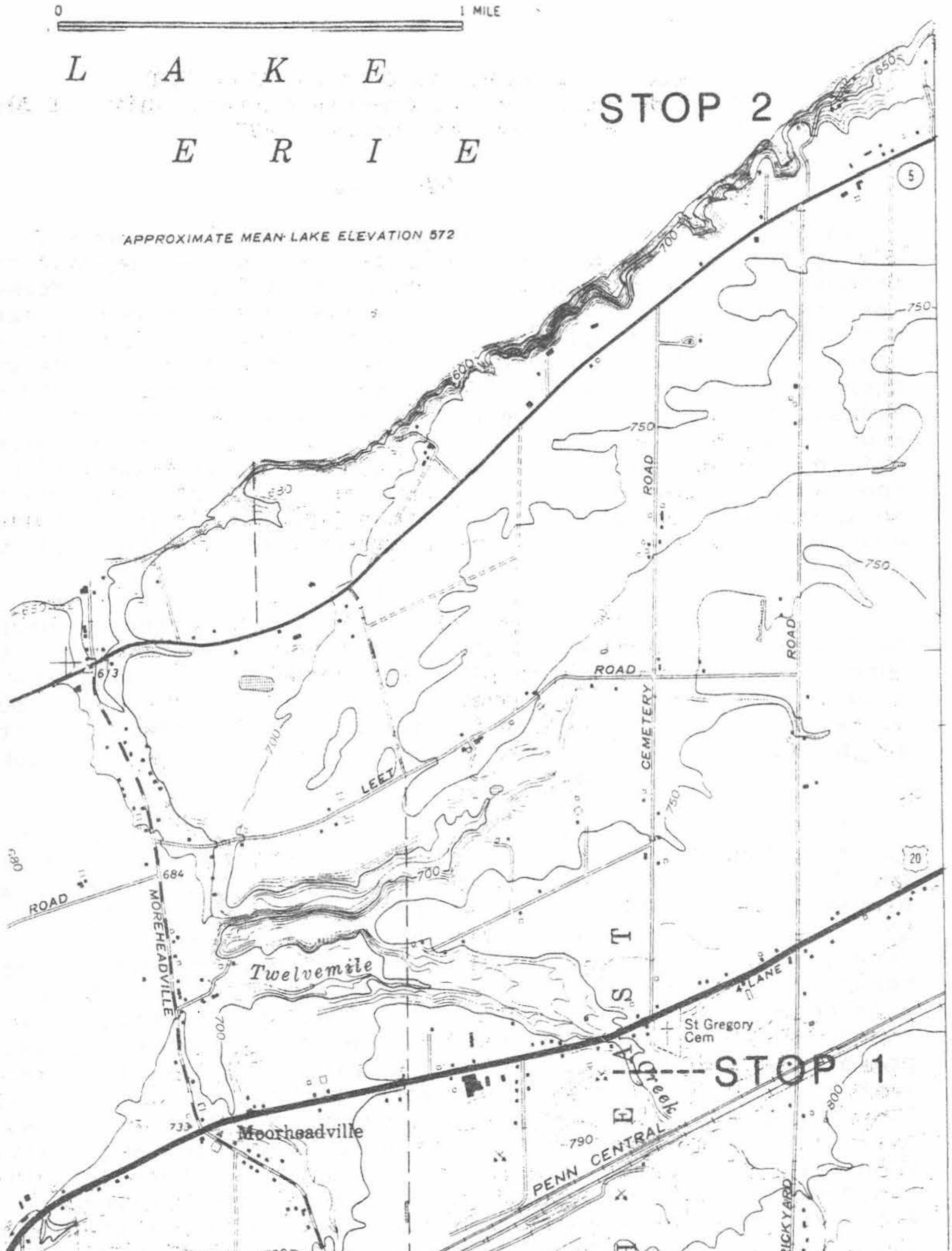


Figure 14. Location map for Stops 1 and 2 (from the Harborcreek, PA 7.5-minute topographic map).

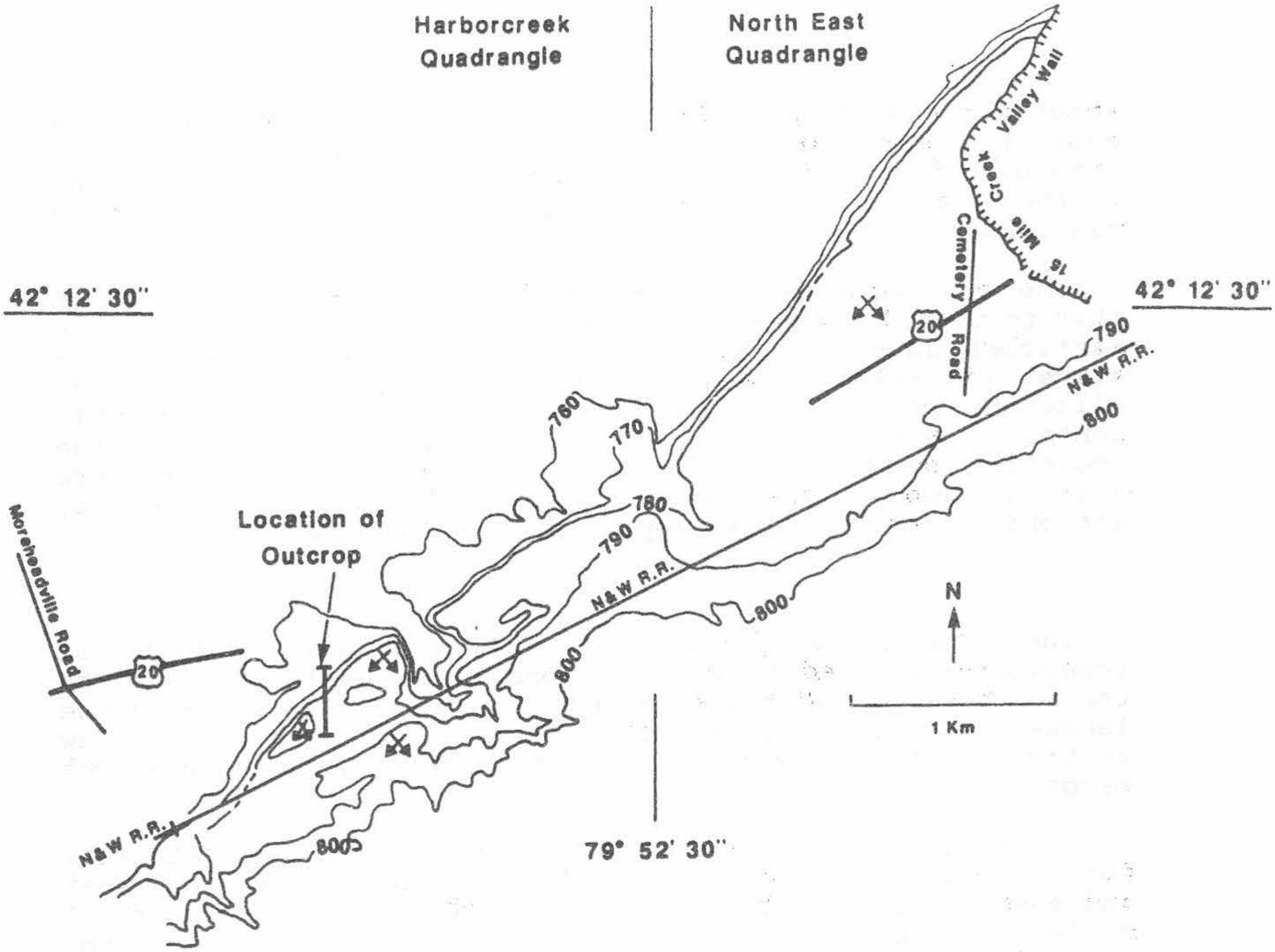


Figure 15. Location map of Central Sand and Gravel pit and plan view of deposit.

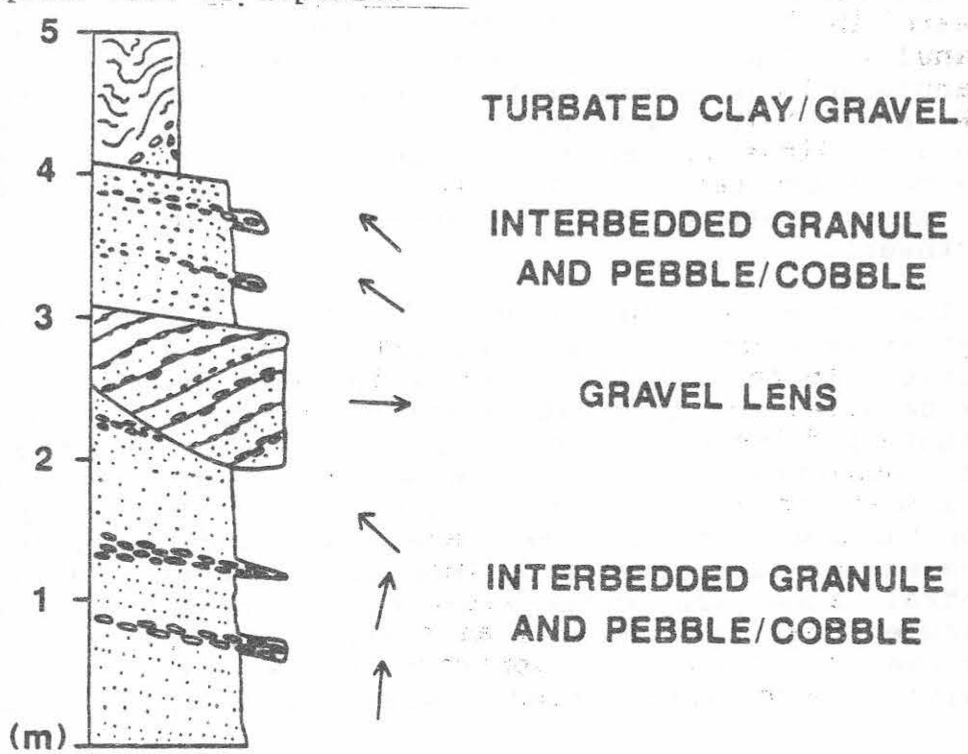


Figure 16. Generalized sequence with facies and paleoflow directions.

about 4 m (13 ft) provides an excellent cross section of the deposit. There are three facies exposed in the face: an interbedded granule and pebble/cobble facies (about 80 percent of the face), a gravel lens facies (5 percent), and a turbated clay/gravel facies (15 percent) (Figure 16).

The interbedded granule and pebble/cobble facies consists of thin to thick beds and lenses that dip gently to the north The particles making up the individual beds or lenses consist almost entirely of well sorted, well rounded, disc-shaped, siltstone clasts. The individual beds and lenses of granule contain laminations and very thin beds that parallel the enclosing bedding planes. The laminations and beds are distinguished on the basis of slight textural differences, although internally they appear to lack grading.

The individual lenses of pebble/cobble consist of framework-supported clasts with long axes oriented parallel to the northerly dip of the surrounding beds. The majority of the lenses are made up of siltstone pebbles no more than a few centimeters thick and thin to single pebble thicknesses at the margins.

In general, the beds and lenses can be traced laterally for 10 to 20 m (33 to 66 ft). The granule beds are thickest and most continuous, but the pebble/cobble lenses are up to 25 m (82 ft) and 20 cm (0.6 ft) thick. In places, the pebble/cobble lenses truncate the underlying granule at low angles. Overall, there is little if any lateral change in this facies from north to south, but vertically the facies fine upward in a 2-3 m (6-10 ft) section from fine pebbles to granules. The true dips of the sloping surfaces in both the granule and the pebble/cobble lenses are oriented between 305' (NW) and 15' (NE) with dips from 4' and 17' (the dips of 6 of the 7 readings lie between 4' and 7'). The dip orientations are not scattered. The 2 lowermost orientations are just east of north whereas the 5 uppermost orientations are to the northwest.

The gravel lens facies consists of flat to convex-up lenses that enclose cross-stratified beds of granules, pebbles, and cobbles In this facies, some of the long axes of the gravels lie parallel to the stratification and others do not. In two flat-topped lenses near the south end of the face, the shape of the cross-stratification resembles lateral accretion surfaces. In a major convex-up lens located about one-third the distance from the southern end of the face, the central part of the lens consists of tightly packed, imbricated gravels. Flanking the central zone are cross-stratified cobbles, pebbles, and granules that fine as well as steepen (11' to 27') away from the center. There is no apparent change in the texture of the individual cross-stratified layers. A second, much smaller but coarser (mostly cobbles), convex-up lens lies about 30 cm (1 ft) above the larger lens. This lens lacks the cross-stratification of the underlying lens. In both lenses the siltstone cobbles dip uniformly in a westerly direction whereas the stratification flanking the major lens dips to the south.

The gravel-size clasts, as in the interbedded facies, are framework supported, although in both facies there are numerous discrete cobble and even siltstone boulders that occur here and there in the section, parallel to the stratification of the enclosing granules, and commonly in the same planes. The pebbles and cobbles in the lenses are mostly siltstone. However, at three scattered places, well sorted and rounded, more spherically-shaped, gravel-sized clasts of mixed rock types are found. A pebble count of 69 clasts from one of these gravel lenses showed the following: 53 percent siltstone, 23 percent sandstone, 20 percent igneous and metamorphic, and 4 percent carbonate clasts.

The gravel lenses can be traced laterally for 10 m (33 ft) or more with the major convex-up lens having a lateral extent of at least 30 m (98 ft). The lenses dip gently to the north and although there are no vertical changes in the lenses, nearly all of the lenses show an updip (south) decrease in grain size. The true dip orientations of the "cross-stratification" in the lenses lie between 160' (SE) and 190' (SW) with dips from 11' to 27'. The true dip orientation of the cobbles in one imbricated zone is 275' (NW) with a dip of 10', and the unmeasured orientation of other imbricated zones is also to the west.

The turbated clay/gravel facies consists of pods of clay separated by contorted zones of granule/gravel. Most of the facies has been removed, but before the surface was stripped there was about 15 cm (0.5 ft) of soil underlain by about 75 cm (2.5 ft) of clay (Scott Wassink, personal communication).

In terms of facies association and order, the interbedded granule and pebble/cobble facies is eroded by the gravel lens facies, with a relief as much as 1 m (3 ft). There are about 6 gravel lenses along the face with the lenses occurring throughout the section in a vertical sense, and along the southern two-thirds of the face. The turbated clay/gravel facies caps the two underlying facies.

Interpretation

The northward-dipping, interbedded granule and pebble/cobble facies is interpreted as proximal Gilbert-type deltaic foresets. The gently dipping lenses represent deposition from grain flow and avalanching on the delta foreset because of flow expansion and deceleration at the river mouth. The coarser gravel lenses may have been segregated during avalanching and/or may represent higher energy, episodic events. Rapid deposition took place during homopycnal flow mixing (Nemec and Steel, 1984). The incomplete foreset thickness of 3 to 4 m (10 to 13 ft) indicate minimum water depths of 3 to 4 m (10 to 13 ft) at the river mouth. The overall fining-upward sequence in these deposits may be due to decreased flow strengths caused by a gentler gradient.

The cross-cutting, cross-stratified sands are interpreted as channel deposits. Incisions of the delta slope by distributary channels during episodic drops in water level, and subsequent rises in water level, may have caused this scour and fill type of structure.

The northerly paleoflows are consistent with a delta/coastal alluvial fan origin. Northward flowing streams coming off the steep isostatically uplifted glaciated uplands debouched into glacial Lake Whittlesey with little modification by waves. The high rate of bedload coupled with low wave energy allowed the deltas to build directly to the north.

Alternate hypotheses could include sandbar or barrier systems. However, the geometry of the deposit and the thickness and lateral extent of the interbedded granule and pebble/cobble facies make these hypotheses unlikely.

ROAD LOG AND STOP DESCRIPTIONS

Cumm. Miles	Miles from last point	Description
6.0	0.0	Exit Central Sand and Gravel Co. pit. Intersection of Gravel pit entrance and US Route 20. Turn right (east) onto Route 20 and continue east.
6.5	0.5	Intersection of US Route 20 and Brickyard road. Turn left (north) and proceed north on brickyard road. We will descend a short, steep slope onto an area of relatively low relief. The elevation along Route 20 is about 780 feet. This is the edge of the Law-Whittlesey Terrace. The elevation at the bottom of the steep slope is 760 feet. This is the general elevation of the Lake Warren I-II level in the eastern section of Erie county. The slope is thought to be a relic wave cut cliff cut by the waves of Lake Warren I and can be seen running parallel to US Route 20 about 40 feet above it.

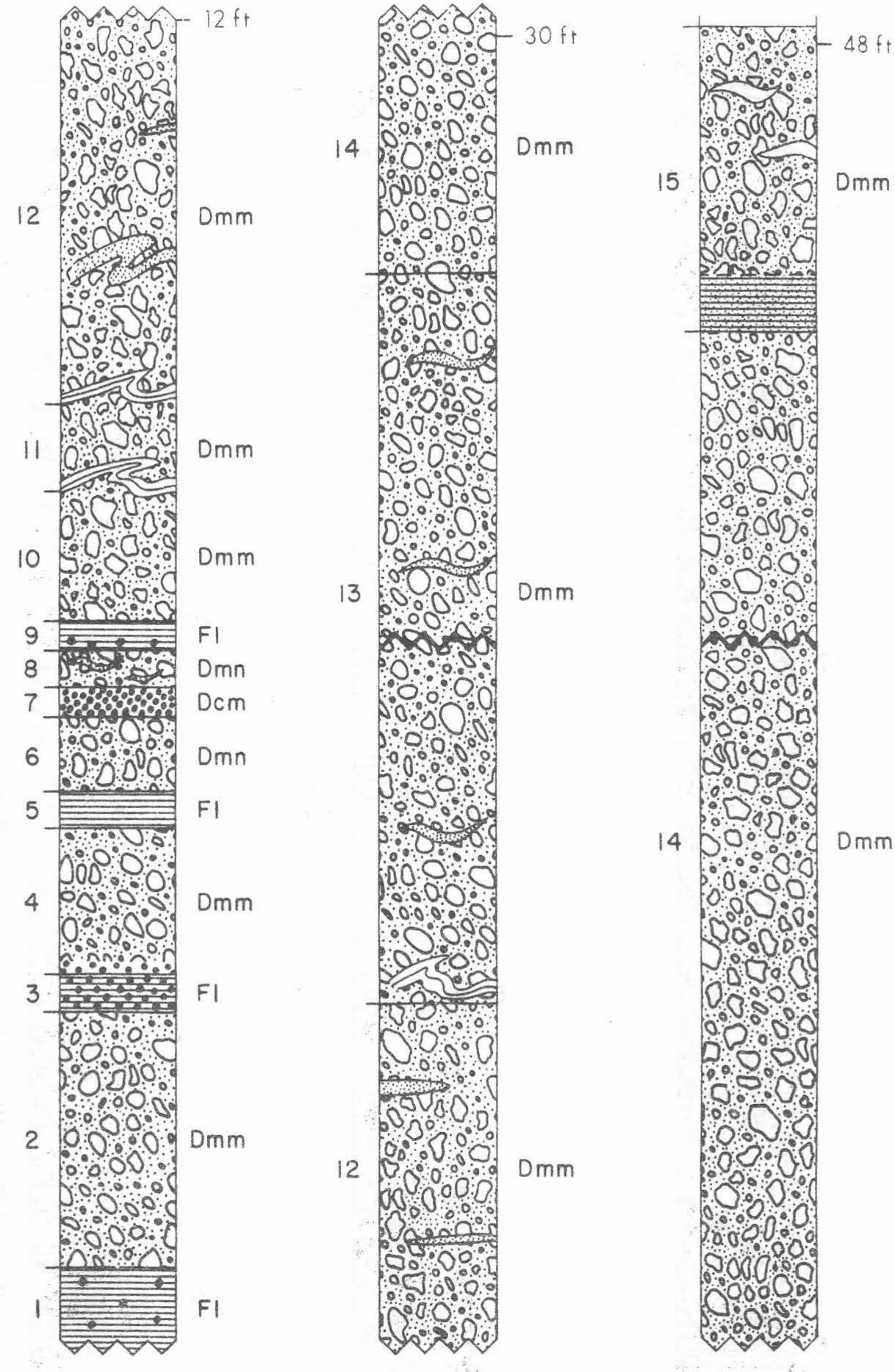
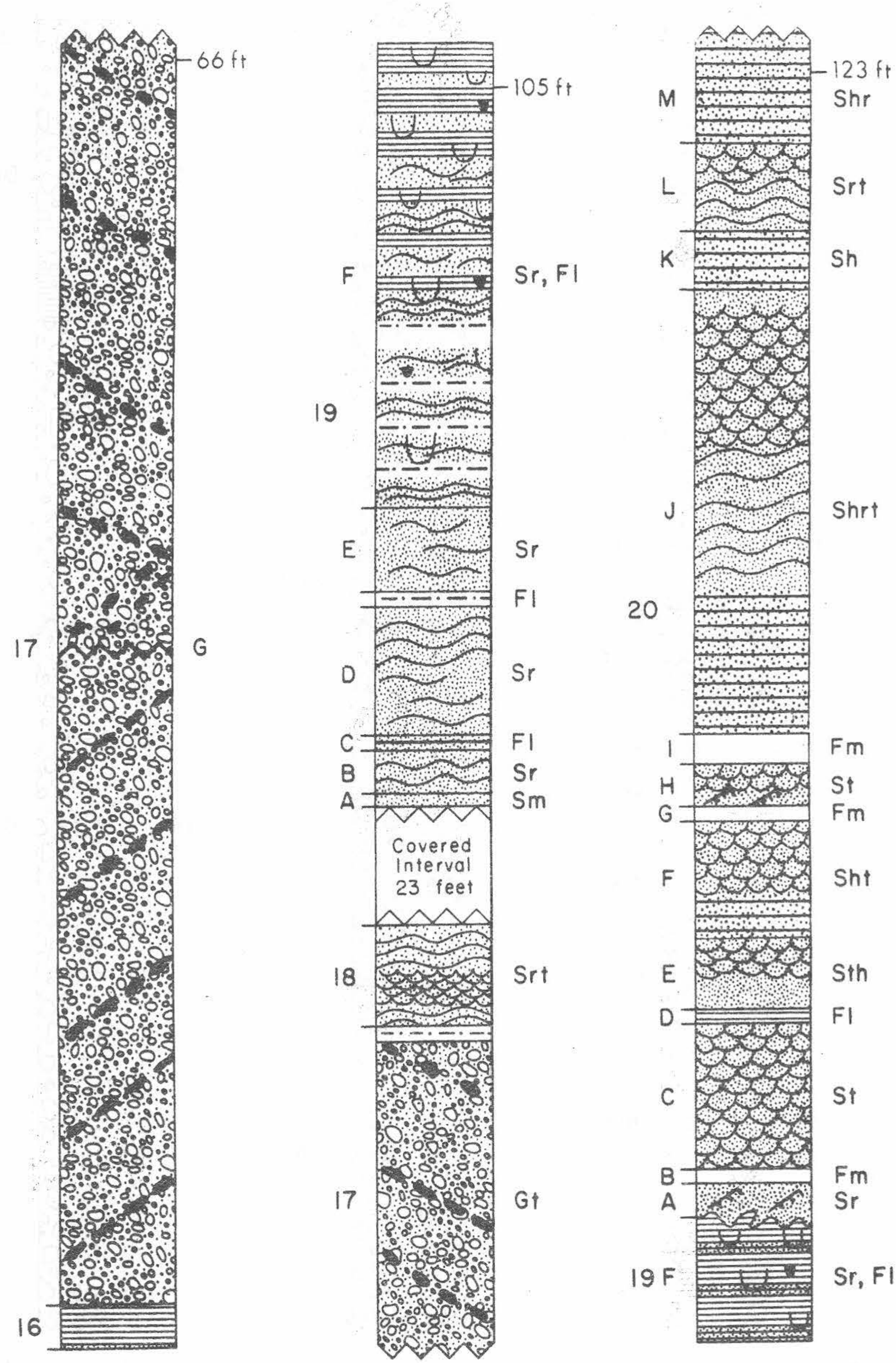
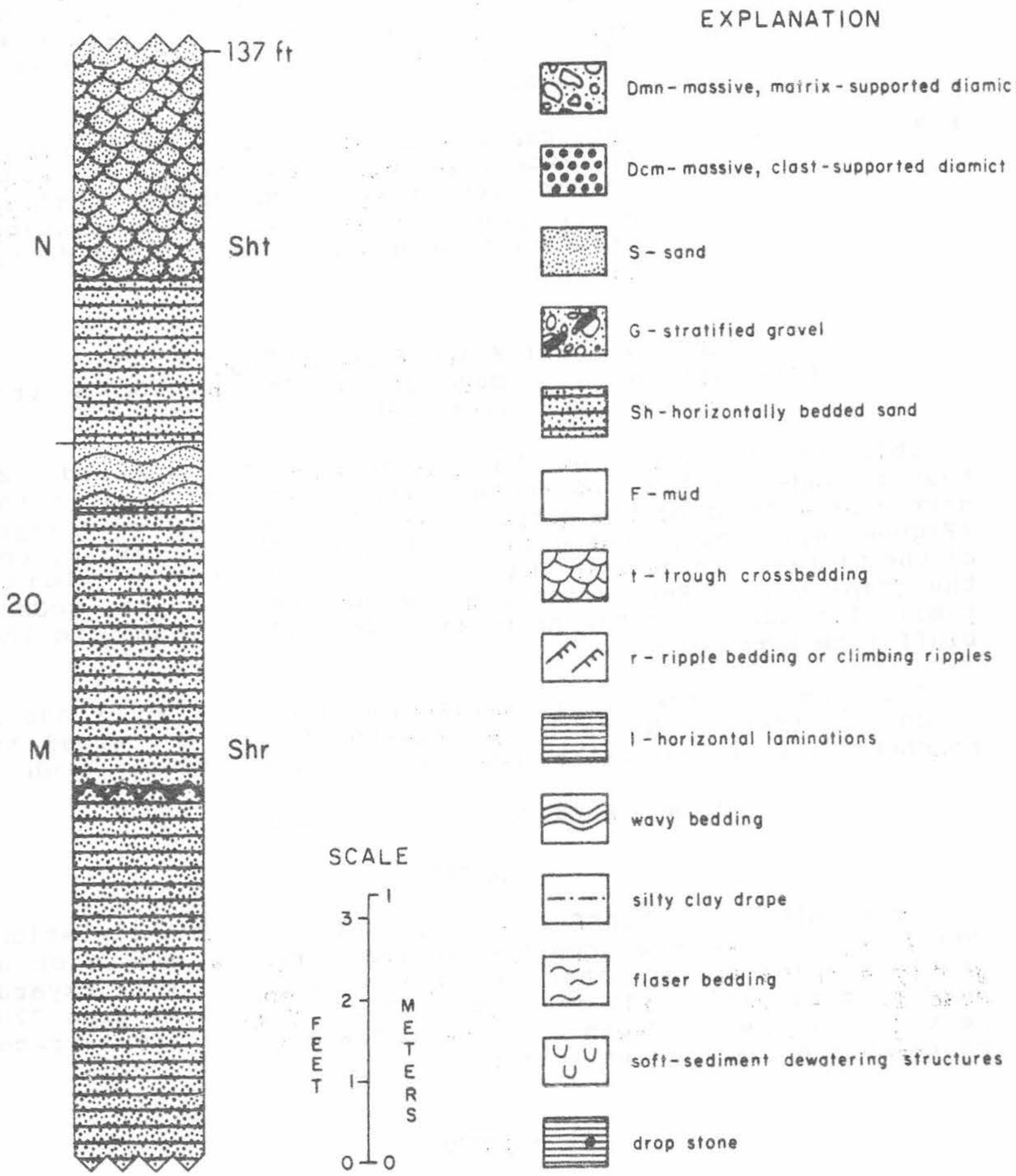


Figure 17. Graphic log of sediments occurring in the Brickyard Road bluff face. See Table 1 for measured section description


66 ft
17
G
16
105 ft
F
Sr, Fl
19
E
Sr
Fl
D
Sr
C
Fl
B
Sr
A
Sm
Covered
Interval
23 feet
18
Srt
17
Gt
123 ft
M
Shr
L
Srt
K
Sh
J
Shrt
20
I
Fm
H
St
G
Fm
F
Sht
E
Sth
D
Fl
C
St
B
Fm
A
Sr
19 F
Sr, Fl

Sun. F28



137 ft
N
Sht
20
M
Shr
SCALE
3
2
1
0
FEET
1
0
METERS
EXPLANATION
Dmn - massive, matrix - supported diamict
Dcm - massive, clast - supported diamict
S - sand
G - stratified gravel
Sh - horizontally bedded sand
F - mud
t - trough crossbedding
r - ripple bedding or climbing ripples
l - horizontal laminations
wavy bedding
silty clay drape
flaser bedding
soft-sediment dewatering structures
drop stone

Sun. F29

7.1	0.6	We are crossing a channeled west flowing tributary stream to Twelve-mile Creek.
7.8	0.7	Intersection of Brickyard road and West Middle road (Leet road on the topographic map). Continue north on Brickyard road. We continue to cross the lake Warren I Terrace.
8.9	1.1	Intersection of Brickyard road and US Route 20. Stop sign. Park vans and unload. Route 20 carries very high speed traffic; participants should cross with caution. Walk north on gravel road along side the vineyard.

STOP 2. BRICKYARD ROAD SECTION.
(Description by: Dave Thomas and Ray Buyce from Thomas and others, 1987)

This stop is located at the foot of a private, unpaved road that extends north of US Route 5 from Brickyard Road in the northeast section of the Harborcreek, PA 7.5-minute quadrangle (Figure 14). The property is owned by the McCord family, one of the pioneer vintners in the Pennsylvania-New York section of the grape belt. Permission must be obtained from the McCord family for access to the property. We will be examining the bluff-face exposures.

The summit of the bluff overlooking Lake Erie about 300 m (1000 ft) from US Route 5 is dangerous! It is capped by noncohesive sand that fails under very small additional loads.

DO NOT WALK TO THE BLUFF EDGE!

Landform

The summit of the bluff at this location is about elevation 700 feet. It is the location of the northern margin of a gently sloping terrace that can be traced south on Brickyard Road from elevation 730 feet at US Route 5 to elevation 770 feet just below US Route 20. We believe that this terrace represents a Lake Warren I level.

Stratigraphy

General. Wave erosion and a variety of mass wasting processes have exposed approximately 50 m (165 ft) of glacigenic deposits represented by 2 distinct facies. The lower part of the section is a diamict facies which comprises 17.4 m (57 ft) or 35 percent of the total bluff height. The upper 33 m (108 ft) or 65 percent of the total bluff height is composed of a gravel and sand facies. Figure 16 presents a graphic display and Table 1 the detailed stratigraphic description of the measured section.

Diamict Facies. This facies may be divided into 2 distinct lithologic types based upon thickness and grain size distribution. The lower segment consists of 11 units (1 through 11, Measured Section) ranging in thickness from 0.06 m (0.2 ft) to 0.6 m (1.75 ft). These 11 units occur as 2 different lithologies. Six of the units are massive, matrix-supported diamicts composed of olive gray, clayey silt to very fine sand supporting angular to subrounded pebbles and rarer gray and reddish gray soft clay clasts. Thicknesses range from 0.07 to 0.53 m (0.25 to 1.75 ft). The other 5 units are laminar-bedded, clayey silt, silt, or very fine sand supporting angular to subrounded pebbles and rarer gray and reddish gray soft pebble-size clay clasts. The thicknesses range from 0.06 to 0.07 m (0.20 to 0.25 ft). These 2 lithologies consistently alternate up through the entire 2.2 m (7.25 ft) of this segment of diamicts. At the lower contact of the uppermost unit (11) are alternating dark and light gray, clayey silt laminae with a roll-up structure and folds overturned westward.

The upper diamict segment is 12.75 m (41.85 ft) thick and consist of 4 units (12 through 15) ranging in thickness from 1.3 to 5.3 m (4.25 to 17.25 ft). These 4 units are composed of olive gray, massive, clayey silt or very fine sand matrixes supporting angular to subrounded pebbles, and fewer cobbles and rare boulders. The angular cobbles are hard siltstones or very fine sandstones whose origins may be found in the Upper Devonian bedrock to the immediate east. The subangular to subrounded cobbles are either Lower Paleozoic sedimentary rocks plucked from the Ontario-Erie basin farther to the east or Precambrian crystalline igneous and metamorphic rocks transported from the Canadian Shield. The lower 2 units (12 and 13) show, at their lower contacts, roll-up structures and folds in laminar clayey silt beds that are overturned toward the west. In addition, there is a normal fault with 6 mm (0.25 in.) displacement in the rolled-up laminar beds near the base of Unit 12.

Gravel/Sand Facies. The diamict and gravel/sand facies is separated by a 0.15 m (0.5 ft) thick laminar-bedded clay with a thin sand bed at the base and top. The contact between the clay and gravel facies appears erosional. The gravel segment of the gravel/sand facies (Unit 17) is 6.6 m (21.65 ft) thick. It consists of angular to subrounded pebbles and interstitial sand and clayey silt. The lower part of the unit has 2 cut and fill channels. The beds of the lowest channel dip northwest, while the beds of the upper channel truncate those of the lower and dip northeast. Bedding is not apparent in the remaining upper part of the gravelly unit; however, discontinuous fine-grain drapes occur throughout. A continuous 2.5 to 5 cm

(1 to 2 in.) drape separates the gravel from the sandy units above.

The sandy segment of the gravel/sand facies is about 20 m (66 ft) to the summit of the bluff. The upper 4.6 to 6 m (15 to 20 ft) were not measured because of steepness and instability of the slope. The sand units are composed of very fine, fine, and medium sand with relatively thin interlayers of clayey silt or very fine sand. The clayey silts or very fine sands occur as drapes.

The lowest unit (18) is composed of 0.46 m (1.5 ft) of very fine laminated sand displaying horizontal, trough, wavy, and flaser bedding and beds of horizontally laminated clayey silt.

The overlying unit (19) is 10.6 m (35 ft) thick. The lower 7 m (23 ft) is covered with sandy colluvium; however, exposures east and west of the Measured Section indicate lithologies consistent with the measured upper 3.7 m (12 ft) of the unit. The upper 3.7 m (12 ft) is composed of alternating beds of very fine sand and clayey silt beds. The laminar clayey silts are bedded horizontally while the very fine sand units display horizontal, trough, wavy, and flaser bedding. Pebble-size drop stones are sparsely scattered through the unit. The thicker sand beds in the lower part of the unit yield to clayey silt beds upward through the unit. This fining upward sequence is contorted by soft sediment deformation structures with amplitudes of 3 ft or more in its upper part. The remainder of the measured section (Unit 20) which is 20.1 m (66 ft) thick is composed of sand and clayey silt or very fine sand beds. The sand beds display laminar and thin horizontal, wavy, trough, and flaser bedding and climbing ripples. The clayey silt or very fine sand occurs as relatively thin drapes.

QUESTIONS

Questions to be asked at this stop are basically the same as those at Stop 2:

1. What process (es) formed the diamicts?
2. Is there evidence of sublacustrine deposition and if so what mechanisms were involved?
3. Is there more than one age of diamict here? What approach could be utilized to resolve the problem?
4. How many different environments of deposition may be represented by these sediments?

Table 1: Measured Section: Brickyard Road Bluff Face.

Stratigraphic section measured on bluff just east of valley which provides access to the beach. The upper 4.6 to 6 m (15 to 20 ft) were not described because of the steepness and instability of the slope.

Unit	Thickness Meters (feet)	Description
20-N	1.49 (4.9)	Sand, light brown, laminar, fine to medium. Trough and horizontal bedding.
20-M	3.0 (10)	Sand, light brown, laminar, fine to medium. Horizontal bedding. Ripple cross laminated unit near top.
20-L	0.36 (1.2)	Sand, light brown, laminar. Wavy and trough crossbedding.
20-K	0.24 (.8)	Sand, fine, light brown, very thinly bedded to thinly bedded. Horizontal bedding.
20-J	1.9 (6.2)	Sand, fine, light brown laminar coarsening upward to medium. Horizontal, wavy and trough crossbedding.
20-I	0.15 (.5)	Clay, gray
20-H	0.18 (.6)	Sand, fine, light brown, laminar. Trough crossbedding and climbing ripples. Slit bed at top.
20-G	0.03 (.1)	Clay, gray, massive
20-F	0.34 (1.6)	Sand, fine, light brown, laminar. Horizontal and trough crossbedding.
20-E	0.34 (1.13)	Sand, fine, and silt; light brown, laminated. Ripple crossbeds with wave lengths of about 6.5 cm (2.5 in.) and amplitudes of 2.5 cm (1 in.) Some flaser bedding.
20-D	0.048 (.16)	Clay, silt, gray, laminar.

20-C	0.6 (2.0)	Sand, light brown, laminar to very thin bedded. Trough bedding throughout.
20-B	0.045 (.15)	Clay, silt, gray, laminar.
20-A	0.09 (.3)	Sand, fine to medium, light brown, laminar. Horizontal and flaser bedding. Climbing ripples with lee and stoss sides preserved.
19	10.74 (35.25)	Sand, very fine, light to medium gray, alternating with silt or clay beds in upper 3.7 m (12 ft). Trough, wavy, flaser, and horizontal bedding in the sand units. Clay or silt laminations are horizontally bedded. Rare subangular to subrounded dropstones. Silt or clay beds become more apparent upward through the silt. Loading structures become more apparent upward with amplitudes greater than 1.5 m (5 ft) in the uppermost part. Lower 7 m (23 ft) is covered with sandy colluvium.
18	0.46 (1.5)	Sand, laminated, olive gray to light olive gray and yellowish brown, with clayey silt. Trough, wavy, and flaser bedding.
17	6.6 (21.65)	Gravel, light brown angular to subrounded pebble-size with silt and sand interstices. A 2.5 cm (1 in) clay drape extends across the top of the unit separating it from Unit 18. Middle bed truncates both the lower and middle beds about 0.6-0.9 (2-3 ft) above the contact with Unit 16. No bedding is evident in the upper 6 m (20 ft) of the unit, but several discontinuous clay drapes occur throughout. Lowest bed dips about 23' northwest.
16	0.1-0.15 (.33-.5)	Sand, fine to very fine, laminated, alternating with clay. Some lenses of pebbles. Sand is rippled at the top of the unit. 5 cm (2 in), laminar, medium gray and dark gray clay near center of unit. Bottom 2.5 cm (1 in) is fine to medium sand.
15	0.38-1.3	Diamict, massive, matrix supported.

	(1.25-4.25)	Olive gray, clayey silt supporting pebble-size angular, hard siltstone and subangular to subrounded sedimentary, igneous, and metamorphic rock fragments with maximum diameters of 2.5 cm (1 in). Large number of cobble-size, angular, hard siltstones and subangular to subrounded sedimentary, igneous, and metamorphic rocks. Cobbles become more abundant and smaller upward.
14	5.26 (17.25)	Diamict, massive, matrix supported. Olive gray, clayey silt supporting pebbles of angular, hard siltstone and subangular to subrounded sedimentary, igneous, and metamorphic rock fragments up to 2.5 cm (1 in) in diameter.
13	3.0 (10)	Diamict, massive, matrix supported. Silt and clay supporting angular to subrounded pebbles with 1.3-5.0 cm (.5-2.0 in) in diameter. Rarer angular siltstone and subrounded to rounded sedimentary, igneous, and metamorphic cobbles and one .6 m (2 ft) subrounded boulder. Cobbles increase in size and number eastward. 3 wavy sand layers, 2, 6, and 25 mm (.15, .25, and 1 in) thick, with horizontal laminations occur within the unit. Roll-up structure overturned toward the west occurs near the base. Lower contact is not distinct, but probably erosional. Undulating intercalated sand bodies along east end of contact. One .6 m (2 ft) cobble and angular, flat siltstone cobbles positioned on Unit 12 at western end.
12	2.97 (9.75)	Diamict, massive, matrix-supported. Olive gray, clayey silt supporting angular, subangular, and subrounded pebbles between 6mm and 5 cm (.5 and 2 in) in diameter with cobble-size angular siltstones showing striations; Subangular to subrounded sedimentary and crystalline igneous and metamorphic rocks with 7-13 cm (3-5 in) diameters in lesser amounts. One 25 cm (10 in) subrounded boulder. Sparsely occurring beds of very fine

		to fine sand between 6 and 25 mm (.25 and 1 in) thick and from .45 to 1 m (.18 to 45 in) long with light and dark laminae below and above a massive core. A roll-up structure and folds overturned to the west and normal faults with apparent dips of 75' west are displaced 6 mm (.25 in) along the fault plane within part of the roll-up structure 0.8 m (2.6 ft) above the lower contact.
11	.13-.4 (.42-1.44)	Diamict, massive, matrix supported. Silt with scattering of angular to subrounded pebbles with a maximum diameter of 2.5 cm (1 in). Lower part shows alternating dark and light laminae. Roll-up structures at bottom contact with Unit 10 are deformed from east to west.
10	.38 (1.25)	Diamict, massive, matrix-supported. Olive gray, clayey silt or very fine sand supporting a scattering of angular to subrounded pebbles with a maximum size of 2.5 cm (1 in) and a few cobbles 8 cm (3 in) or more in diameter.
9	0.06 (.21)	Sand, very fine to fine, occurring as 1-2 mm alternating light gray and light gray laminae. Rip-up clasts from Unit 8 in bottom 6 mm (.25 in) of unit. Dark gray layer of mud laminae in upper 1.4 cm (.5 in). Rippled very fine sand bodies with amplitudes and lengths of 3 mm (.1 in).
8	0.08 (.27)	Diamict, massive, matrix-supported. Silt with scattering of angular to subrounded pebbles. Wavy laminae 2 mm (.05 in) thick of very fine sand occur within one unit.
7	0.008 (.27)	Gravel, pebble size. Maximum diameter of 2.5 cm (1 in) with very fine interstatial sand, silt and clay. No bedding or imbrication observed.
6	0.15 (.5)	Diamict, massive, matrix-supported. Brownish gray silt or fine sand matrix supporting a scattering of angular to subrounded pebbles and

		rare grayish red and medium gray soft clay clasts.
5	0.08 (.25)	Sand, very fine, thin to medium laminae, light brownish gray alternating with dark gray silt or clay laminae.
4	.15 (.5)	Diamict, massive, matrix-supported. Olive gray to brownish gray silt or fine sand supporting a scattering of angular to subrounded pebbles and rare grayish red soft clay clasts with olive gray interstatial silt or very fine sand.
3	0.08 (.25)	Silt or fine sand, alternating in thin to medium laminae of pale red and grayish red with dense scattering of angular to subrounded pebbles of up to 2.5 cm (1 in) in diameter.
2	.53 (1.57)	Diamict, massive, matrix supported. Olive to brownish gray silt or very fine sand supporting a scattering of angular to subrounded pebbles up to 2.5 cm (1 in). Pebbles become more sparse upward. Slight evidence of bedding when dampened with water.
1	0.08 (1.75)	Silt and/or very fine sand in alternating bands of light and darker olive gray laminar bedded, supporting a sparse scattering of subrounded to rounded pebbles and rare clasts of grayish red clay. BASE OF SECTION

ROAD LOG

Cumm. Miles	Miles from last point	Description
8.9	0.0	Leave Stop 2. Proceed south on Brickyard road. We will be travelling across the Warren I-II Terrace from elevation 740 feet to elevation 760 feet over the next 1.8 miles, a gradient of about 11 feet per mile.
10.0	1.1	Intersection of Brickyard road and West

		Middle road to the right. Continue south on Brickyard road. US Route 20 is straight ahead (south) running along the edge of Lake Whittlesey Terrace, just above the Warren I scarp (Wave Cut?).
10.7	0.7	Stop sign at intersection of Brickyard road and US Route 20. Turn left (east) onto US Route 20 and continue east. We are now travelling on the Lake Whittlesey Terrace with the Warren I-II Terrace to our left.
11.1	0.4	Intersection of US Route 20 and Williams road to the right. Continue east on US Route 20.
12.2	1.1	Intersection of US Route 20 and Cemetery road. Continue east on US Route 20.
13.0	0.8	Traffic signal at intersection of US Route 20 and South Mill street. Continue east through the town on Northeast, Pennsylvania.
13.4	0.4	Traffic Signal at intersection of US Route 20 and PA Route 89. Continue east on US Route 20.
14.0	0.6	Yellow flasher intersection of US Route 20 and Orchard Beach road. Continue east on US Route 20.
14.1	0.1	Boundary sign of Northeast Township. Also beginning of divided highway. Continue east on US Route 20.
16.4	2.3	Intersection of US Route 20 and Eastbound entrance ramp of Interstate 90. Proceed onto entrance ramp and on to Fredonia, New York.

REFERENCES CITED

Barnett, P.J., 1985, Glacial retreat and lake levels, north central Lake Erie basin, Ontario, in Karrow, P.F. and Calkin, P.E., eds, Quaternary evolution of the Great Lakes: Geological Association of Canada Special Paper 30, p. 185-194.

Berg, T.M., and others, 1980, Geologic map of Pennsylvania: Pennsylvania Geologic Survey, 4th ser., Map no. 1.

Berg, T.M., and others, 1986, Stratigraphic correlation chart of Pennsylvania, revised edition: Pennsylvania Geologic Survey, 4th ser., General Geology Report 75.

Calkin, P.E., and Feenstra, B.H., 1985, Evolution of the Erie-Basin Great Lakes in Karrow, P.F., and Calkin, P.E., eds., quaternary Evolution of the Great Lakes: Geological Association of Canada Special Paper 30, p. 147-170.

Coakley, J.P. and Lewis, C.F.M., 1985, Postglacial lake levels in the Erie basin in Karrow, P.F. and Calkin, P.E., eds., Quaternary evolution of the Great Lakes: Geological Association of Canada Special Paper 30, p. 195-212.

Dreimanis, A., and Goldthwait, R.P., 1973, Wisconsin glaciation in the Huron, Erie, and Ontario lobes, in Black, R.F., Goldthwait, R.P., and Willman, H.B., eds., The Wisconsin Stage: Geological Society of America Memoir 136, p. 71-106.

Fullerton, D.S., 1980, Preliminary correlation of post-Erie Interstadial events (16,000-10,000 radiocarbon years before present), central and eastern Great Lakes region, and Hodsonm, Champlain, and St. Lawrence lowlands, United States and Canada: U.S. Geological Survey Professional Paper 1089, 52 p.

Harper, J.A., 1986, Oil and gas developments in Pennsylvania in 1985: Pennsylvania Geologic Survey, 4th ser., Progress Report 199, 112 p.

Harper, J.A., 1987, Oil and gas developments in Pennsylvania in 1986: Pennsylvania Geologic Survey, 4th ser., Progress Report 200.

Hough, J.L., 1958, Geology of the Great Lakes: University of Illinois Press, Urbana, 313 p.

Lapham, D.M., 1975, Interpretation of K-Ar and Rb-Sr isotopic dates from a precambrian basement core, Erie County: Pennsylvania Geological Survey, 4th ser., Information Circular 79, 26 p.

Larsen, C.E., 1985, A stratigraphic study of beach features on the southwestern shore of Lake Michigan: new evidence of Holocene lake level fluctuation: Illinois Geological Survey, Environmental Geology Notes 112, 31 p.

Leverett, F., 1892, On the correlation of moraines with raised beaches of Lake Erie: American Journal of Science, 3rd Series, 43, p. 281-301.

Nemec, W and Steel, R.J., 1984, Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits, in Koster, E.H., and Steel, R.J., eds., Sedimentology of gravels and conglomerates: Canadian Society of Petroleum Geologists Memoir 10, p. 1-31.

Schooler, E.E., 1974, Pleistocene beach ridges of northwestern Pennsylvania: Pennsylvania Geologic Survey, 4th ser., General Geology Report 64, 38 p.

Shepps, V.C., White, G.W., Droste, J.B., and Stitler, R.F., 1959, Glacial geology of northwestern Pennsylvania: Pennsylvania Geologic Survey, 4th ser., Bulletin G 32, 54 p.

Thomas, D.J. and others, 1987, Pleistocene and Holocene geology of a dynamic coast. Guidebook for the 52nd annual field conference of Pennsylvania geologists.: Pennsylvania Geological Survey, 88 p.

Tomikel, J.C., and Shepps, V.C., 1967, The geography and geology of Erie County, Pennsylvania: Pennsylvania Geological Survey, 4th ser., Information Circular 56, 64 p.

Totten, S.M., 1982, Pleistocene beaches and strandlines bordering Lake Erie: in White, G.W., ed., Glacial geology of northeastern Ohio: Ohio Geological Survey Bulletin 68, p. 52-60.

White, G.W., Totten, S.M., and Gross, D.L., 1969, Pleistocene beaches and bars, Presque Isle, Erie, Pennsylvania (6/22/74-8/10/74): unpublished M.S. thesis, State University of New York College at Fredonia, 157 p.

PADDLING UP A MELTWATER CHANNEL: A LATE-WISCONSINAN ICE-MARGINAL CRUISE NEAR FREDONIA, NEW YORK

Randy J. Woodbury and Michael D. Jensen, Department of Geosciences, State University of New York College at Fredonia, Fredonia, N.Y. 14063.

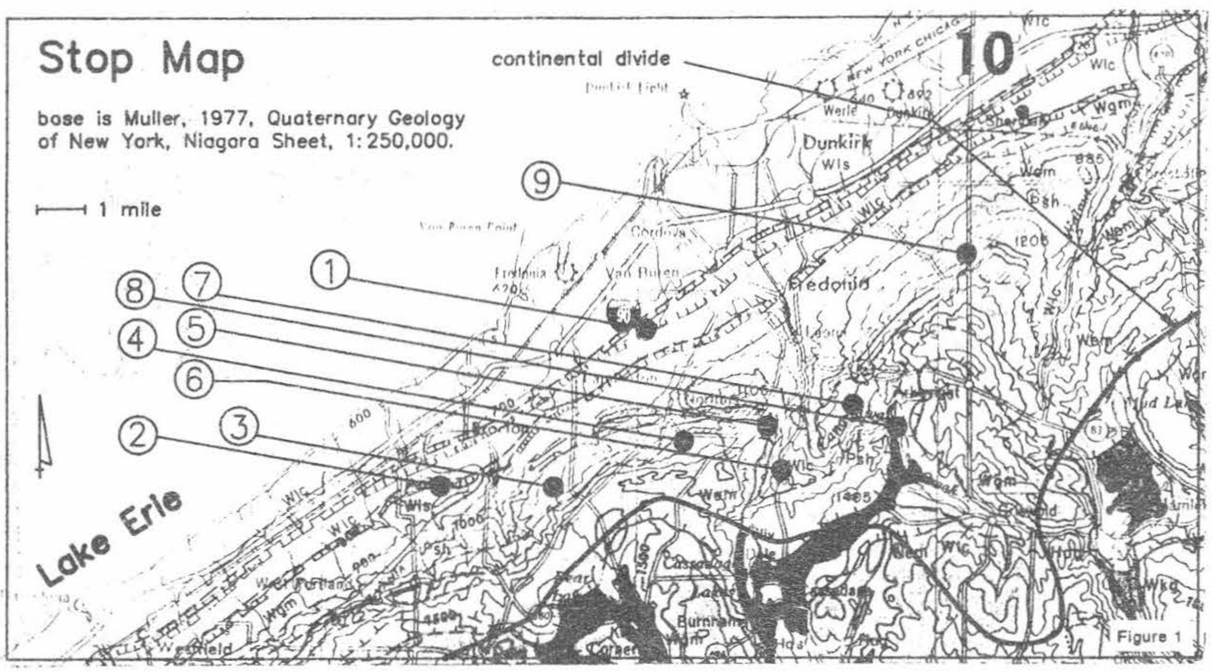
INTRODUCTION

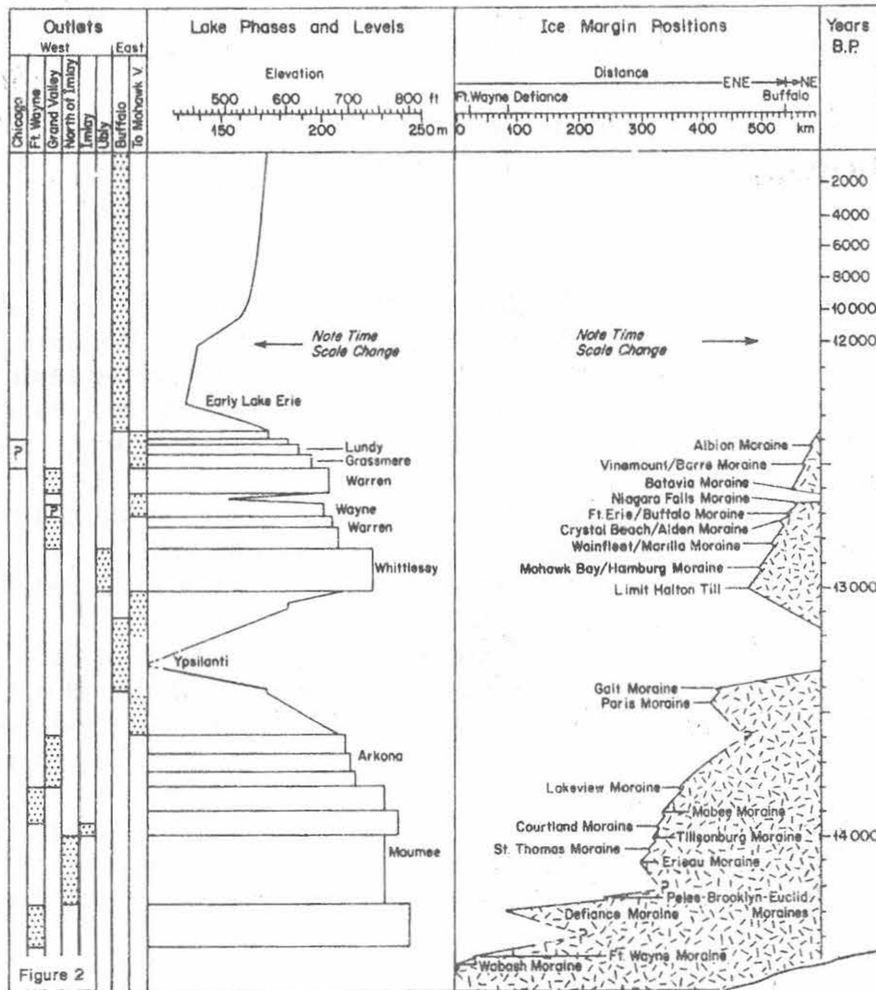
Although the area near Fredonia was glaciated several times during the Pleistocene (see, for example: Leverett, 1902; Fairchild, 1907; Muller, 1963, 1977; Muller and Fahnestock, 1974; Schooler, 1974; and Cadwell, 1988), it was the most recent Late Wisconsinan ice margins, active here between 14,000 and 12,000 years B.P. (Calkin and Feenstra, 1985), that left some of the more dramatic erosional and depositional marks on the landscape.

Staying within eight miles of the college and on the Lake Erie side of the nearby St. Lawrence - Mississippi continental divide, this trip will climb the hydraulic gradient of a glacial meltwater system that incised deep channels through Devonian bedrock, depositing sediment as deltaic gravels and lake clays.

We will begin at beaches of two of the large ancestral pro-glacial lakes of the Erie basin that are locally 170 feet and 240 feet, respectively, above the water level of present Lake Erie. The trip will ascend another 300 feet, then another 100 feet, as we follow a sub-marginal meltwater channel to the levels of two smaller lakes that were imponded along the ice margin.

Stops 1, 2, and 7, on the map below, will allow hands-on inspection of glacially-derived gravels, and stop 6 will be in lake clay. The rest of the stops will be to view a channel that is now relatively dry, but would have been actively carrying meltwater during one or more glacial phases. The diagrams on the next two pages offer background on local glacial history and a schematic view of the subject channel.





LEFT: The preferred chronology from Calkin and Feenstra, 1985, for the succession of ancestral lakes in the Erie Basin.

Elevations on the preferred chronology differ from those found in western New York, as this area was isostatically depressed by the weight of the glaciers. Our field trip will consider elevations to be relative, ignoring local isostatic differences (only about one contour interval from Stop 2 to Stop 9).

1. Carlton
2. Albion
3. Barre
4. Batavia
5. Niagara Falls
6. Buffalo
7. Alden
8. Lackawanna
9. Marilla
10. Hamburg
11. Rock Glen
12. Gowanda
13. Springville-Nunda
14. Lake Escarpment
15. Eagle
16. Farmersville
17. Findley Lake
18. Clymer
19. Randolph

RIGHT: Generalized Ice Margins for western New York, younger south to north (Cadwell, 1988).

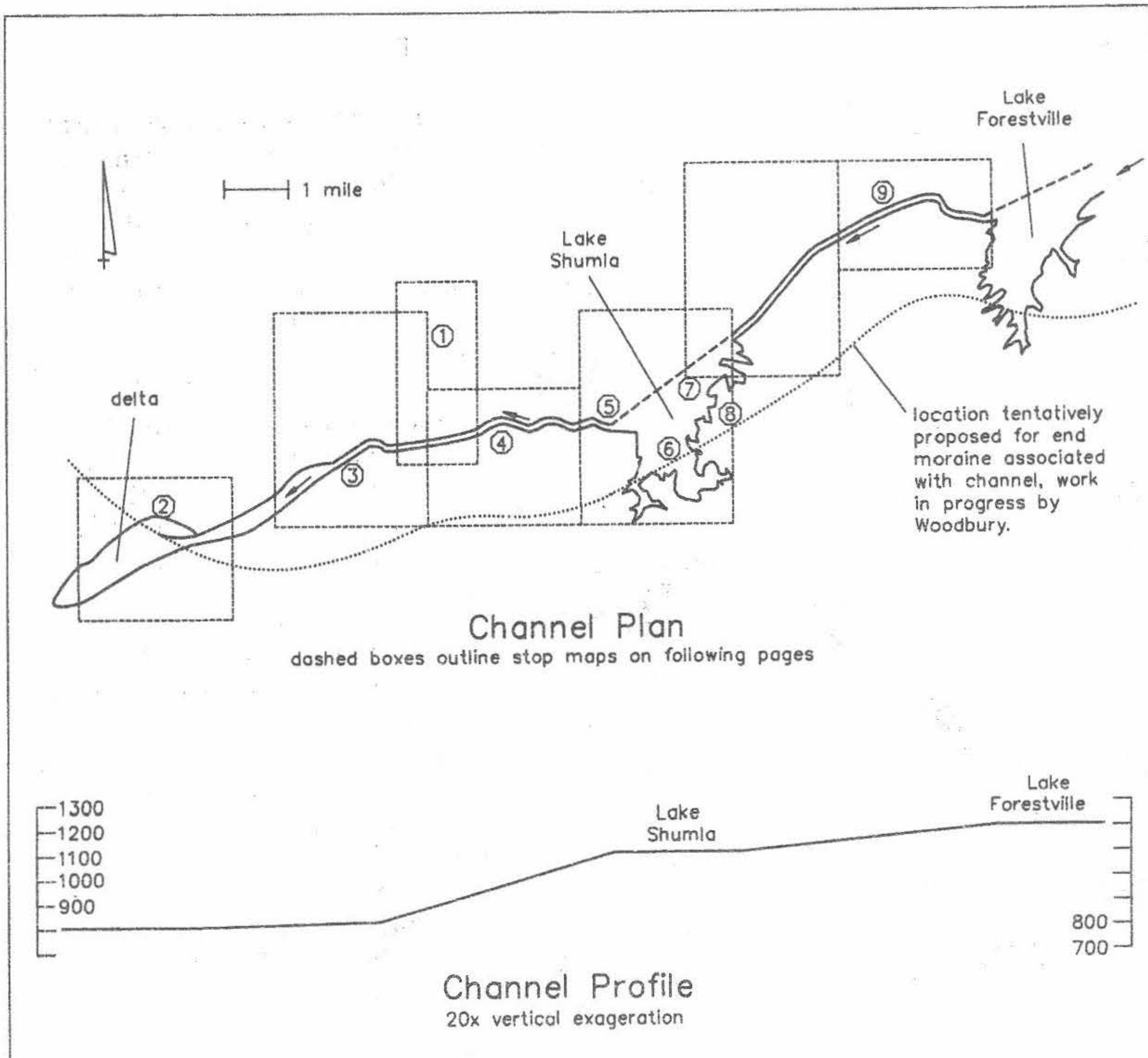
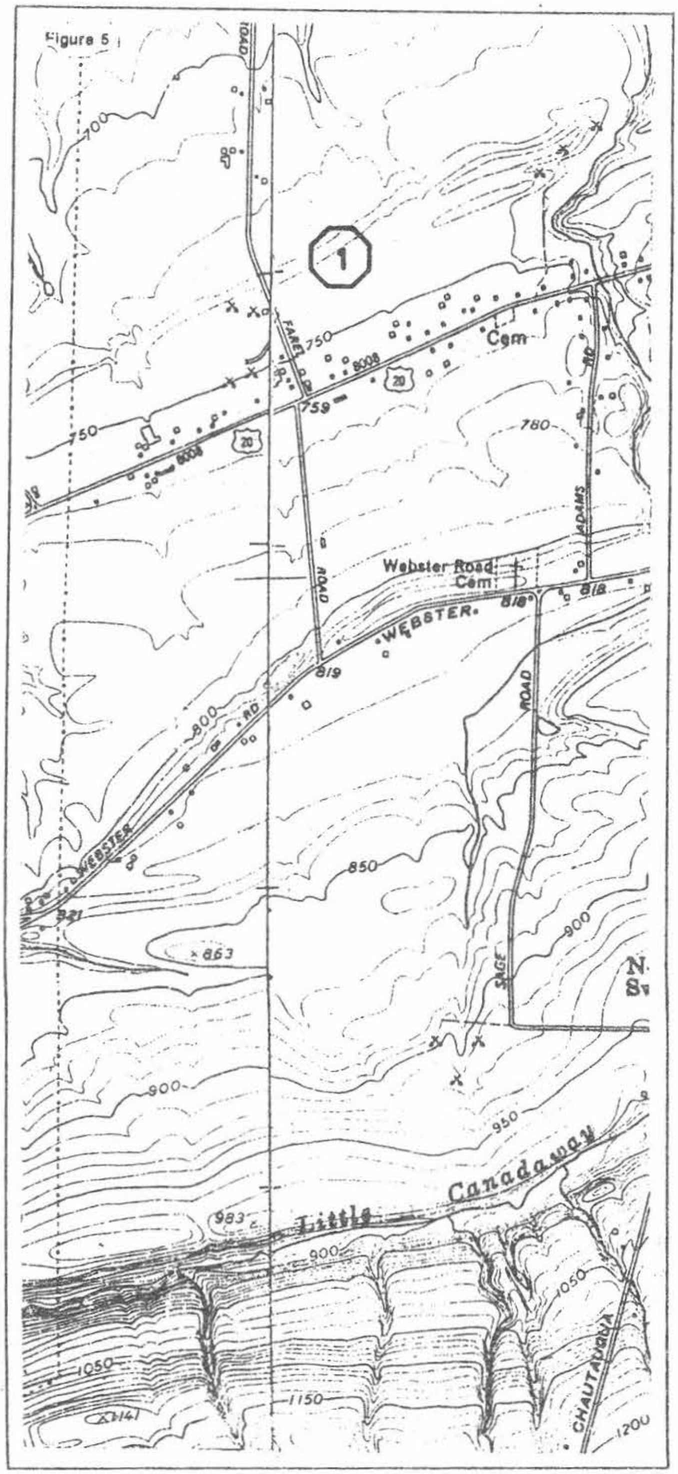


Figure 4

THE CHANNEL

During the waning stages of the Late Wisconsinan, glaciers in the Erie basin passed over the Chautauqua County area several times (see the diagrams at left). With ice over the Buffalo area blocking easterly drainage of the ancestral Great Lakes, water level rose to spill westerly into tributaries of the Mississippi drainage system. It was during one or more of these westerly-draining lake phases that the ice margin for the Erie lobe glacier stood near the continental divide several miles south of Fredonia, and remained there long enough for glacial meltwater to incise a sub-marginal channel flowing westerly into the ancestral lake(s).

The chart above shows this meltwater channel and where we will stop to have a look at it. During which glacial phase(s) the channel was formed and used is a focus of work in progress. It may correlate with the Lake Arkona period; and it may correlate with a westerly projection of the Gowanda end moraine, or part of the Lake Escarpment end moraine complex. The subject is open for discussion.



STOP 1

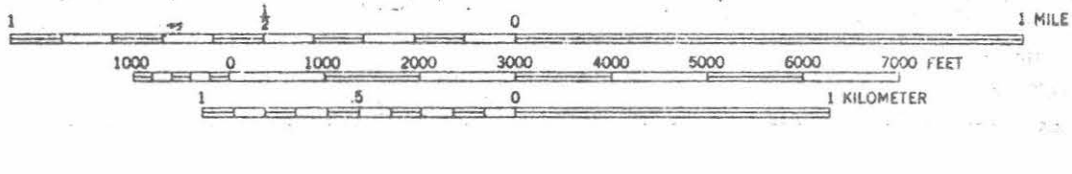
This gravel pit displays typical deposits of a Lake Warren shoreline. U. S. Route 20 is constructed primarily of this material and follows the beach ridge for several hundred miles from northern Ohio to Batavia, New York. Lake Warren deposits are typically cross-bedded, well washed sands and gravels. They were deposited within the high energy beach environment associated with each Lake Warren shoreline. The northward dipping beach face is well exposed on the western and eastern pit walls.

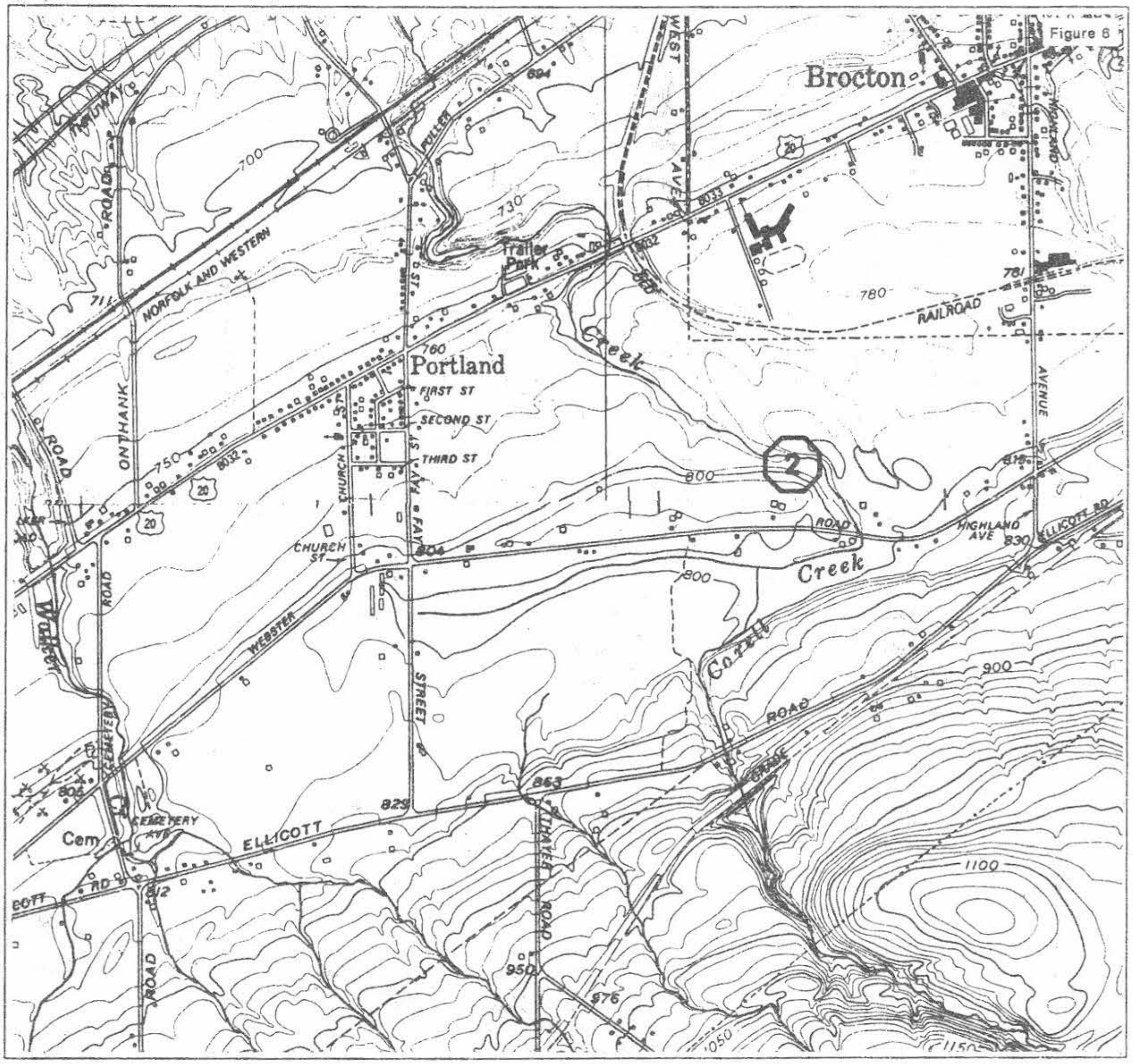
Lake Warren deposits differ from the corresponding Lake Whittlesey deposits which lie further to the south and about 70 feet higher in elevation on the lake plain (found along the ridge through Webster Road Cemetery on the map at left). Lake Warren deposits are generally finer grained and better washed than typical Lake Whittlesey deposits. This may have been a function of higher wave energy and/or a longer occupation of the Warren beach.

Notice the swallow nests bored into the firm silt deposits near the upper wall of the pit.

The beach of Lake Warren is the youngest glacial feature we will visit today, with Lake Whittlesey about 500 years older, and our subject meltwater channel (seen as Little Canadaway Creek on the map) older still.

SCALE FOR TOPOGRAPHIC MAPS, 1:24,000, NORTH IS UP FOR ALL MAPS





STOP 2

This is one of two gravel pits near Portland that were recently opened for construction of a State prison in Brocton. While the elevation and nature of the gravel correspond to a Lake Whittlesey shoreline, the original deposition may have been into an earlier lake. Notice on the map that the 800-foot contour outlines a possible relic spit, formed by wave currents moving sediment easterly from the area near Cemetery Road at Webster Road.

The gravel here is widespread, forming a wide plateau between Webster Road and Ellicott Road. Much of the sediment is locally-derived, with large clasts of siltstone, apparently from Devonian strata higher in the escarpment to the south and east. The feature we are in has been recognized as a delta by Fairchild (1907), who considered that the sediment might be primarily rock excavated from the meltwater channel to the east and northeast. The southerly wall of this channel can be seen on the map as the ridge where Ellicott Road meets Highland Avenue.

EVIDENCE FOR ICE MARGINAL FLUCTUATION IN AND NEAR A RELICT MELTWATER CHANNEL IN CHAUTAUQUA COUNTY, NEW YORK Nº 57772

WOODBURY, Randy J., and LASH, Gary G., Dept. of Geology, State University College, Fredonia, NY 14063

The escarpment southeast of Lake Erie in western New York, with relief of about 300 meters, is drained by stream networks generally trending northward. Anomalous flow directions (along with other supporting evidence) may indicate a glacio-fluvial origin for the stream valley; e.g., Little Canadaway Creek which flows to the west and is recognized as a relict sub-marginal meltwater channel and outlet of a pro-glacial lake. Little Canadaway Creek is roughly V-shaped, with about 30 m of relief and is parallel with and about 3 km north of the Lake Escarpment terminal moraine complex of late Winsconsin age. The channel, holding an under-fitting stream, was apparently incised in the flat-lying Devonian shales and siltstones during one or more glacial stands. Although much of the valley floor is exposed bedrock with a westerly inclined gradient, subsurface borings where Chautauqua Road crosses the channel through a topographic gap in the north valley wall reveal anomalously deep bedrock covered by about eight m of sediment. Fluvio-lacustrine clay and silt deposits are among the sediments recovered here. Peat and twigs collected from 5 m below present creek level yield a radio-carbon date of 8,520± 200 years BP. Palynological analysis of this horizon indicates a predominanceof hemlock (Tsuga). The anomalously low bedrock elevations, the fluvio-lacustrine sediments, and refraction seismic data suggest that the natural topographic gap used by Chautauqua Road may be the junction point for a deeper, drift-filled branch or former outlet of the upper channel. This gap separates two different geometries in the channel: sinuous westward meanders upstream and southwestward linear downstream. It appears to be a nodal point around which ice margin conditions pivoted, burying one section of the meltwater channel and cutting or reopening another. The radio-carbon and pollen data indicate post-glacial filling of the exposed junction to the buried channel.

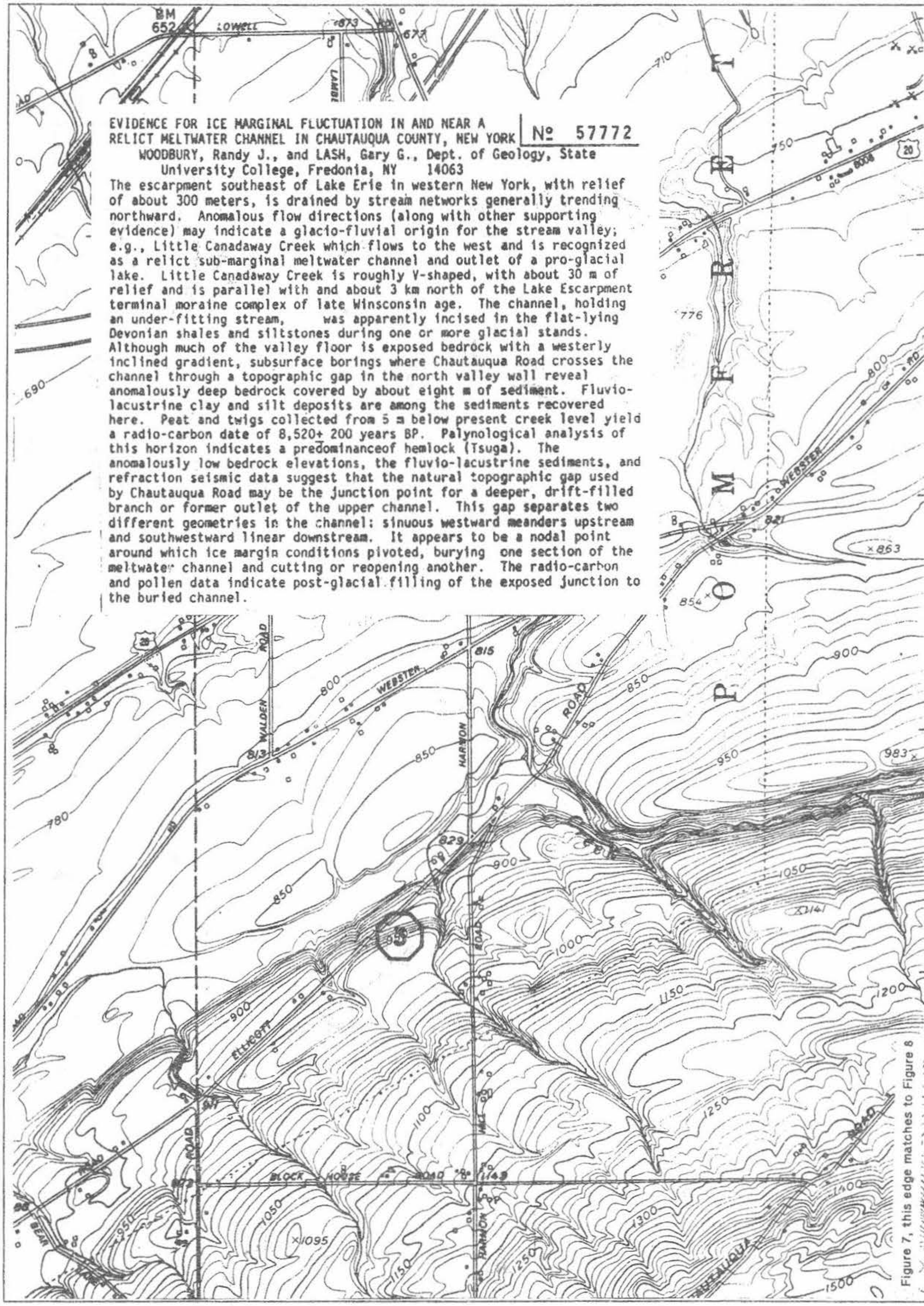
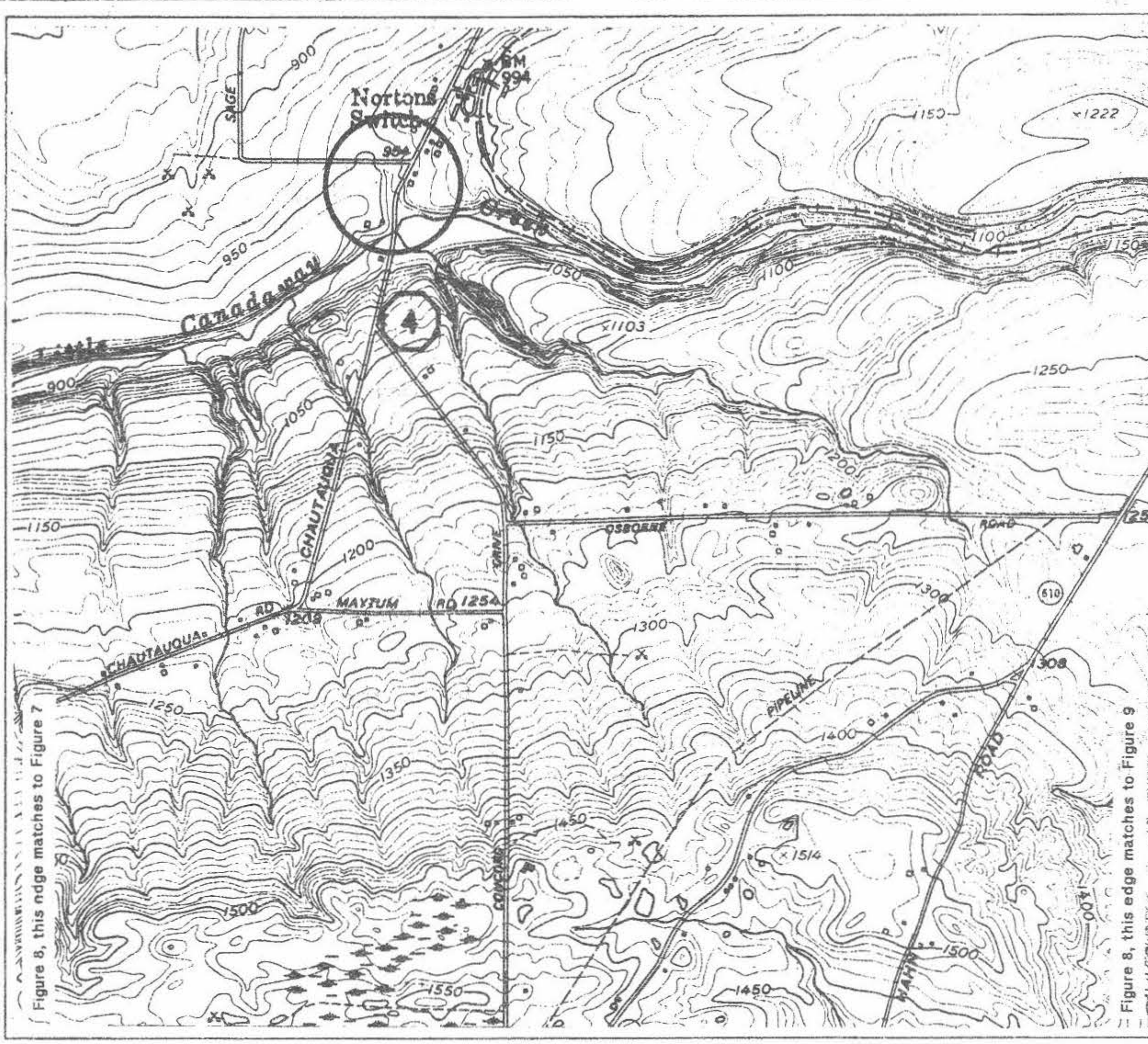


Figure 7, this edge matches to Figure 8

STOPS 3 & 4

Stop 3 is an overview of channel morphology; envision a river of meltwater raging through the now-dry valley. Stop 4 takes us to the channel's intersection with Chautauqua Road, where a drift-filled valley running northwesterly has been proposed in the area circled (Woodbury and Lash, 1985, abstract is inset on facing page). The natural gentle slope over the buried valley provided the least treacherous route for early settlers to cross the deep meltwater channel; Chautauqua Road is one of the county's oldest roads, commissioned by the Holland Land Company in the very early 1800s to connect Fredonia (then known as Canadaway) with the county seat in Mayville. At stop 4 we will visit one of several 40-foot waterfalls that are characteristic of the streams running northerly into the relic channel. The waterfalls are recessed about 1000 feet southerly from the meltwater channel, apparently marking the extent of post-glacial headward erosion away from the "hanging" positions they would have had along the channel wall when the glacier melted away.

The swampy knob and kettle terrain to the south represents an end moraine near the crest of Concord Drive.



STOP 5

This is the outlet col for pro-glacial Lake Shumla (Fairchild, 1907), formed as ice dammed the northerly flowing Canadaway Creek. The outlet elevation, controlling the lake surface, is 1105 feet. Visibility permitting, we will use a surveying instrument to view an inlet channel that brought a flood of meltwater into Lake Shumla; the inlet can be seen as a shallow topographic dent along the hilltop located about 3 miles to the northeast.

For a wider perspective of the outlet channel, fold the map on the preceding page into this page and match it to the map at right.

Also on the map, note the notch in the hillside northwest of our stop. This could have been an outlet for a lower level of Lake Shumla.

STOP 6

We are 130 feet under the water surface of Lake Shumla where fine clay particles settled to the bottom. Some investigators have classified the clays as varved, interpreting the rhythmic laminations as seasonal variations in depositional environment. The roadway is "Old Route 60," while present Route 60 is located over the knoll to the east and is also cut through lake clays. The slump-prone sediments have been a problem for highway engineers for nearly 30 years, and we will drive through a clay cut that has yet to be worked into a stable angle of repose.

Lake Shumla was formed several times and at several water levels during the Pleistocene glacial oscillations, so the clays we see are not likely the result of just the most recent glaciation.

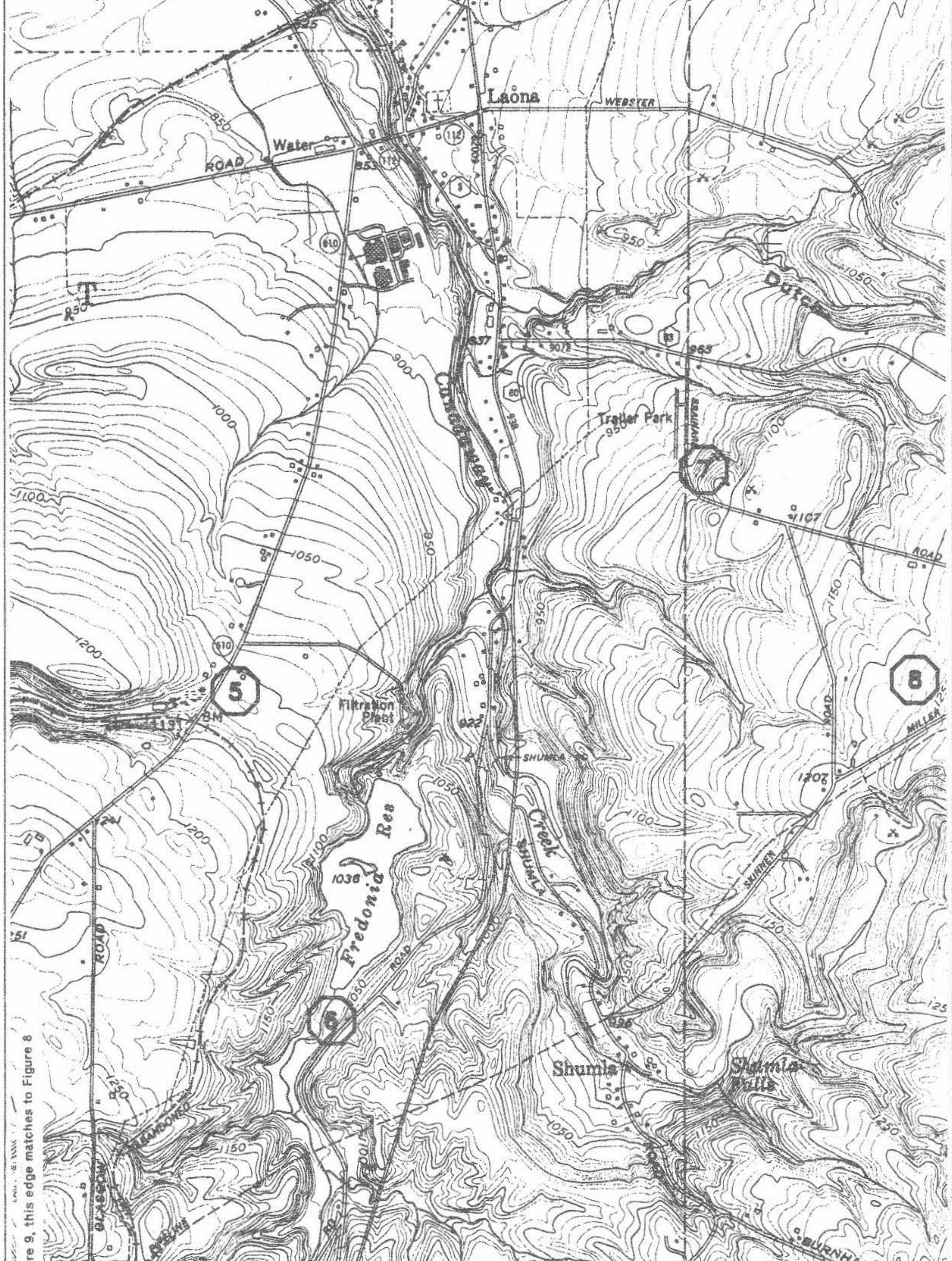
STOP 7

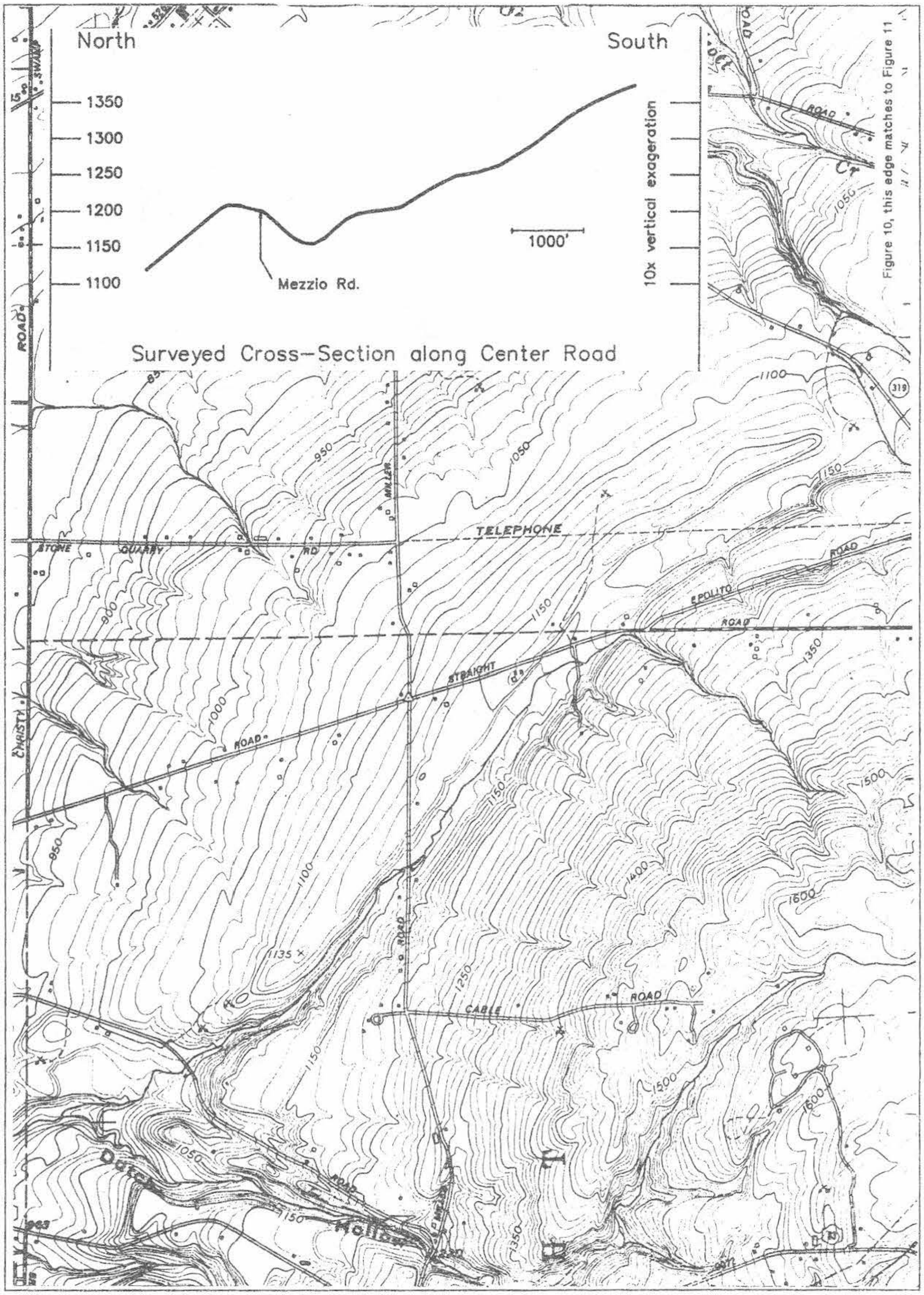
This gravel pit is near Lake Shumla's inlet. We appear to be in a delta deposited and worked by meltwater flowing from the channel and upper lake located northeast of here. Lake Shumla provided a quiet "sink" for sediments in the meltwater chain: the coarse-grain sediments settled at the delta, and the fine-grained clays deposited at mid-lake, while little or no sediment was carried by the meltwater just as it spilled out the channel to the southwest.

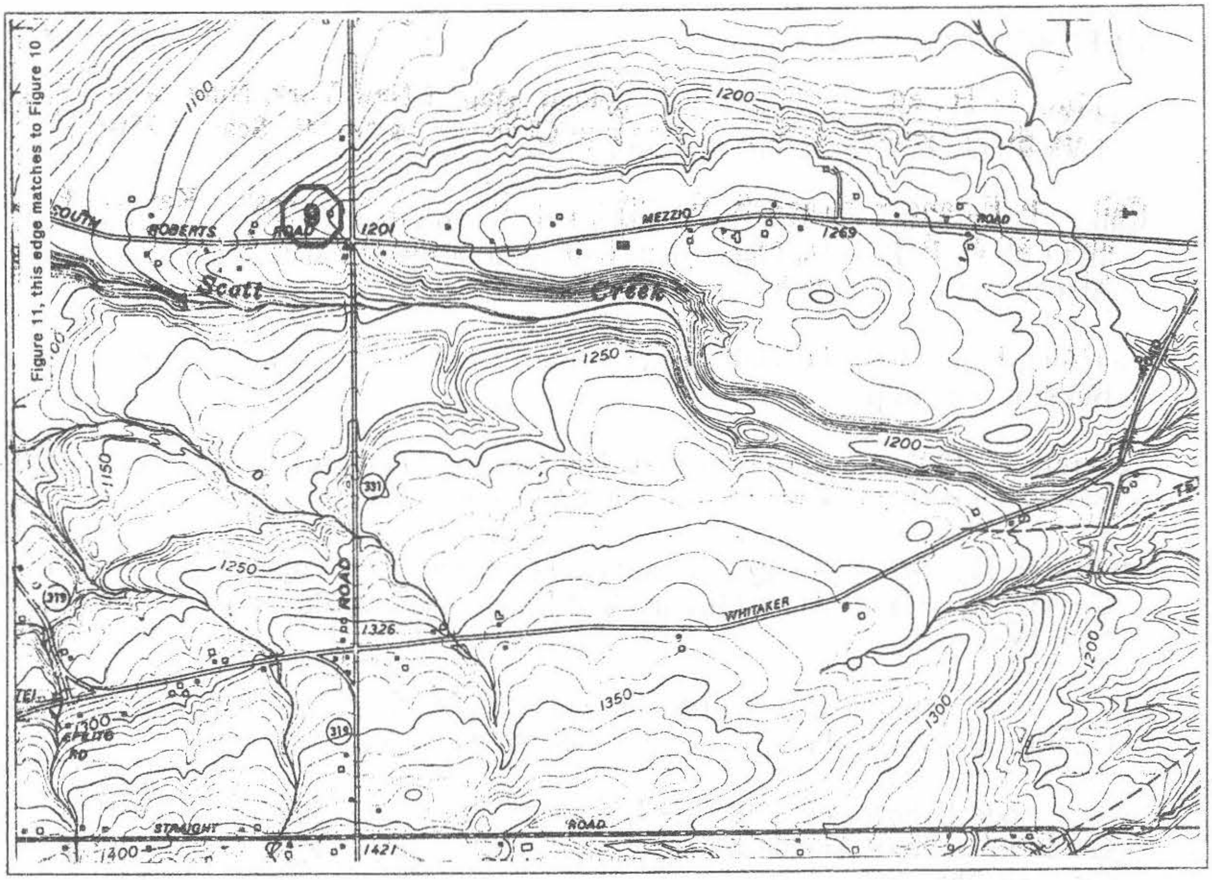
STOP 8

This stop presents a rather spectacular overview of the outlet channel of Lake Shumla. It appears as a dramatic V-notch in the hilltop about 2 miles to the west.

As you view the channel, mentally re-construct a mountain of slushy ice overhead, thickening northeasterly, toward Buffalo, and thinning to an end moraine gushing with meltwater within 2 miles to the south. We are about 150 feet above the lake, looking down on the 2.5 square-mile surface of frigid water. Can you spot any icebergs?


Laona
WEBSTER
Water
ROAD
850
900
950
1000
1050
1100
1200
1150
1250
Dutch
965
Trailer Park
7
1107
ROAD
5
8
BM
Filtration Plant
922
SHUMLA RD
1207
MILLER
Creek
Fredonia Res
1038
SKINNER
6
Shumla
Shumla Falls
BURNHAM
GLASGOW
ROAD
Figure 9, this edge matches to Figure 8


North
South
1350
1300
1250
1200
1150
1100
Mezzio Rd.
1000'
10x vertical exageration
Surveyed Cross-Section along Center Road
Figure 10, this edge matches to Figure 11
TELEPHONE
STONE QUARRY RD
EPOLITO ROAD
STRAIGHT ROAD
CHRISTY ROAD
CABLE ROAD
MILLER
ROAD
950
1050
1150
900
1000
1100
1135
1250
1350
1400
1500
1600



STOP 9

We have just driven through the meltwater channel and are parked near its northerly crest. Our route is depicted by the cross-section inset at left. Compare the section to your view looking southerly, down into and across the channel. Note the terraces on the southerly wall, which were apparently cut by ice-marginal streams walled by the glacier on their northerly sides.

On the map, note the deep, dry channel easterly and southeasterly from our stop. This was the spillway for a pro-glacial lake, with a surface elevation of 1205 feet, that formed over the Forestville area as ice dammed the northerly flowing Walnut Creek. Our field trip ends here, so we'll save Lake Forestville and its inlets for a new day and a fresh paddle.

Before we depart, examine the brass disk set in concrete near the southwestern corner of the road intersection. It is a horizontal ground control station for the photogrammetrically-produced 7.5-minute-series topographic sheets. USGS designates it "TT 38 WX 1952," with position determined at 42°27'07.544" north latitude and 79°14'12.489" west longitude.

REFERENCES CITED

Cadwell, D. H., ed., 1988, Surficial Geologic Map of New York, Niagara Sheet: New York State Geological Survey Map and Chart Series No. 40, Scale 1:250,000

Calkin, P. E., and Feenstra, B. H., 1985, Erie-Basin Great Lakes: in Karrow, P. F., and Calkin, P. E., eds., Quaternary Evolution of the Great Lakes: Geological Association of Canada Special Paper 30, p. 156-170.

Fairchild, H. L., 1907, Glacial Waters in the Erie Basin: New York State Museum Bulletin 106, 86 p.

Leverett, F., 1902, Glacial Formations and Drainage Features of the Erie and Ohio Basins: United States Geological Survey Monograph 41, 802 p.

Muller, E. H., 1963, Geology of Chautauqua County, New York, Part 2, Pleistocene Geology: New York State Museum and Science Service Bulletin 392, Albany, New York.

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

______, and Fahnestock R. K., 1974, Looks at the present and recent past: in Guide Book, 46th Annual Meeting, New York State Geological Association, Fredonia New York, p.D-1 - D-37.

Schooler, E. E., 1974, Pleistocene beach ridges of Northwestern Pennsylvania: Pennsylvania Geological Survey, General Geology Report 64, 38 p.

Woodbury, R. J., and Lash, G. G., 1985, Evidence for ice marginal fluctuation in and near a relict meltwater channel in Chautauqua County, New York: Northeast section, Geol. Soc. Am., Abstracts with Programs.

AUTHORS' NOTE

This trip has been a tour through graduate work in progress by Woodbury under advisement by M. P. Wilson. The general glacial history and the lake deposits are the forte of Jensen who has studied under P. E. Calkin at the University of Buffalo. Woodbury is primarily responsible for the graphics and text from the introduction through the stop descriptions, while Jensen is responsible for the road log narrative and the description of Stop 1.

ROAD LOG FOR PADDLING UP A MELTWATER CHANNEL

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
-1.1	0.0	Assembly area at the large limestone erratic near Houghton Hall.
-0.8	0.3	Turn left (west onto Temple Street Entrance Road)
-0.75	0.05	Turn left (south onto Temple Street)
0.0	0.75	Intersection of Temple Street and US Route 20, turn right (proceed west on US Route 20).
0.2	0.2	Passing over Canadaway Creek. This valley formed embayments for proglacial Lake Whittlesey and Lakes Warren.
0.5	0.3	Intersection US Route 20 and Chestnut Street. Continue west on US Route 20. US Route 20 follows the Lakes Warren strandline from Central Ohio into Western New York. Gravel pits close to the highway provide an abundant source of aggregate for the highway.
1.0	0.5	View of the Escarpment to the left of the highway. There are no resistant units to account for the formation of this cuesta. Abundant grape vineyards are growing in the well drained sands and gravels of Lakes Warren and Whittlesey beach deposits to the south.
2.1	1.1	A cemetery to the left. There are many cemeteries located in the well drained deposits because they can be dug in almost year round.
2.6	0.5	Turn right (north onto Farel Road)
2.9	0.3	Turn right (east into Ron Morse Gravel Pit)

STOP 1. GRAVEL PIT IN LAKES WARREN BEACH DEPOSITS.

2.9	0.0	Turn left (south onto Farel Road from the entrance to the gravel pit)
3.1	0.2	Turn right (west back onto US Route 20)
3.2	0.1	To the south the apparent wave cut terrace just north of Webster Road is the southern wall of an abandoned gravel pit.
6.7	3.5	Village of Brocton, New York.
7.25	0.55	The Brocton Arch. Prior to World War 2 several of the villages in Chautauqua County had similar arches in their villages. However, all of the villages but Brocton removed theirs to provide scrap metal for the war effort.

8.8	1.55	Turn left (south onto Fay Street)
9.3	0.5	Turn left (east onto Webster Road). Webster road follows the Lake Whittlesey Strandline for several miles from the Town of Portland east through the town of Pomfret).
9.8	0.5	Turn left (north into gravel pit (PLAN A)).
10.2	0.4	Turn left (north into gravel pit (PLAN B)).

STOP 2. GRAVEL PIT IN LAKE WHITTLESEY BEACH DEPOSITS.

10.4	0.2	Stop for a beer at the Castle.
10.7	0.3	Turn right (south onto Highland Road)
10.8	0.1	Turn left (east onto Ellicott Road)
10.9	0.1	To the north notice the fine surficial expression of the meltwater channel.
11.3	0.4	Proceed straight (west) through the left side of the fork towards the stop sign.
11.5	0.2	Proceed through the right side of the fork and continue to bear west on Ellicott Road.
11.7	0.2	Notice the gas well just off to the north of Ellicott Road. Gas wells are very common throughout Western New York. They are drilled primarily into the highly productive sandstones of the Silurian Medina Group which dip gently towards the southwest at approximately 50 ft/mi.
13.2	1.5	Pull off the road onto the right shoulder out of away from traffic.

STOP 3. OVERVIEW OF THE MELTWATER CHANNEL.

13.5	0.3	Turn left (north onto Harmon Hill Road)
13.6	0.1	Passing through the channel towards the Lake Whittlesey strandline at Webster Road.
14.0	0.4	Turn right (east onto Webster Road)
14.8	0.8	Bear left at the fork in the road at the stop sign. Webster road is still following the ancestral shoreline of Lake Whittlesey.
15.6	0.8	The southern wall of the abandoned gravel pit mentioned previously within this roadlog is just left (north) of the road.
16.1	0.5	Turn right (south onto Vine Road.) NOTE: This intersection is not marked but is just east of the sign for the Webster Cemetery.
17.0	0.9	Bear left (east at "L" intersection)
17.4	0.4	Passing over a buried channel indicated by seismic profiles and core samples.
17.4	0.0	Turn right (south onto Chautauqua Road)

17.5	0.1	Dropping into the channel and passing over the misfit Little Canadaway Creek which currenly occupies the meltwater channel.
17.6	0.1	Passing up the south wall of the meltwater channel.
17.7	0.1	Turn left (east onto Concord Drive)
17.8	0.1	Pull off the road onto the right shoulder away from traffic.

STOP 4. A WALK INTO THE ANCESTRAL MELTWATER CHANNEL PLUS A VISIT TO ONE OF ITS "HANGING" TRIBUTARIES.

18.2	0.4	Turn left (west onto Osborn Road)
19.6	1.4	Turn left (north onto Fredonia-Stockton Road)
20.5	0.9	Pull off the road onto the right shoulder away from traffic.

STOP 5. A GLIMPSE OF THE DISTANT INLET CHANNEL AND A PANORAMIC VIEW OF PROGLACIAL LAKE SHUMLA.

21.9	1.4	Turn right (east onto Webster Road)
22.0	0.1	Turn right (south onto Portage Street)
22.4	0.4	Turn right (south onto NY Route 60)
23.3	0.9	Turn right (west onto Spoden Road)
24.7	1.4	Pull off the road onto the right shoulder away from traffic.

STOP 6. DITCH EXPOSURE OF LAMINATED (VARVED?) CLAYS WITH DROPSTONES AS DEPOSITED INTO PROGLACIAL LAKE SHUMLA.

24.7	0.0	Continue west on Spoden road.
25.7	1.0	Turn left (north onto NY Route 60)
26.5	0.8	OPTIONAL STOP: This is an optional stop due to the safety constraints put on us by a large group. This is a Fine exposure of laminated clays which were deposited into Lake Shumla. When NY Route 60 was initially cut through this area the engineers placed a 90 degree roadcut through this materiel. Over the following years the materiel slumped every time it was water laden until it reached the much stabler shallower exposure that appears today. As you climb up the exposure you will see beautiful clean clay littered with abundant dropstones.
28.6	2.1	Turn right (east onto NY Route 83)
29.2	0.6	Turn right (south onto Brainard Road)

29.5	0.3	Turn left (east into Conti's Gravel Pit)

STOP 7. GRAVEL PIT IN THE DEPOSITS OF A PROBABLE DELTA BUILT INTO LAKE SHUMLA.

29.5	0.0	Turn left (south on Brainard Road)
29.8	0.3	Turn right (west onto Skinner Road)
30.5	0.7	Turn left (north onto Miller Road)
30.7	0.2	Pull off the road onto the right shoulder away from traffic.

STOP 8. SPECTACULAR VIEW OF THE COL WHICH WAS THE OUTLET OF PROGLACIAL LAKE SHUMLA.

31.6	0.9	Proceed straight through the intersection of Miller Road and NY Route 83.
32.6	1.0	Proceeding down the south wall of the channel.
32.9	0.3	Passing through the botton of the channel.
33.0	0.1	Passing up the north wall of the channel.
33.3	0.3	Turn right (east onto Straight Road)
33.7	0.4	Passing down the south wall of the channel.
33.8	0.1	Passing through the bottom of the channel.
34.0	0.2	Passing up the north wall of the channel.
34.5	0.5	Proceeding up a long slope onto the Escarpment which is the northern edge of the Allegheny Plateau.
35.4	0.9	Turn Left (north onto Center Road)
36.1	0.7	Passing over an asymetrical terrace above the south wall of the channel.
36.3	0.2	Passing over a terrace which is the crest of the southern wall of the channel.
36.5	0.2	Passing through the bottom of the channel.
36.55	0.05	Passing up the north wall of the channel.
36.6	0.1	Turn left (west onto S. Roberts Road)
36.65	0.05	Pull off the road onto the right shoulder away from traffic.

STOP 9. CROSS-SECTIONAL VIEW OF CHANNEL AND TERRACES.

36.65	0.0	Continue (west on S. Roberts Road)
38.4	1.75	Optional stop and picnic at the Woodbury Winery.
38.5	0.1	Proceed straight (north) through the stop sign.
39.4	0.9	Turn left (west onto US Route 20)
40.7	1.3	Turn right (north onto NY Route 60)
41.3	0.6	Turn right onto New York State Thruway (Interstate 90) Interchange Number 59.