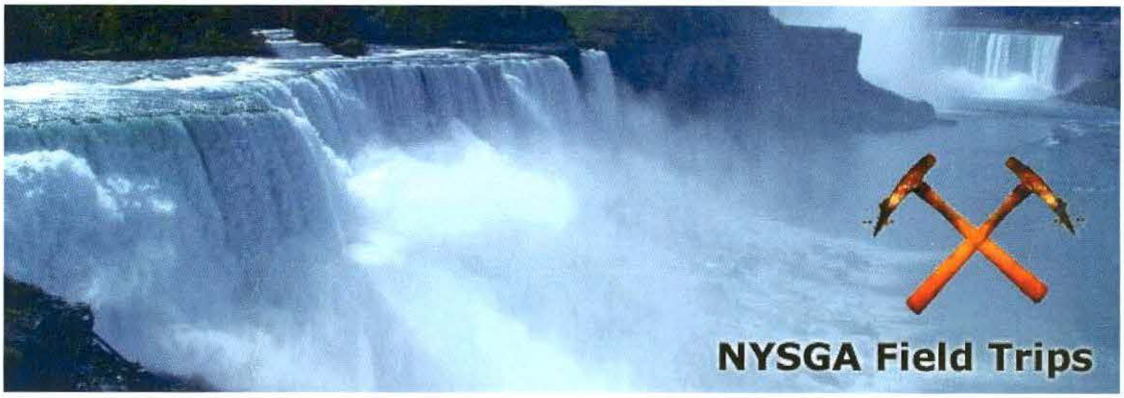

New York State Geological Association

78th Annual Meeting

October 6-8, 2006

Hosted by:
University at Buffalo Rock Fracture Group
Department of Geology, 876 NSC
University at Buffalo
Buffalo, NY 14260

Robert Jacobi, General Chair and Compiler

Conference Headquarters:
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Saturday, October 7, 2006

Sunday, October 8, 2006

Saturday and Sunday, October 7 & 8, 2006 (TWO DAY TRIP)

Faulting and Mineralization in the Cambro-Ordovician Section of the Mohawk Valley

Saturday Field Trip A1

Guidebook for the field trip held October 7th, 2006
in conjunction with the 35th Eastern Section AAPG Meeting
and 78th NYSGA Field Trips
held in Buffalo, NY

Paul Agle[1], Robert Jacobi[1], Charles Mitchell[1],
Rich Nyahay[2], Brian Slater[2], Langhorne Smith[2]

Field Trip Leaders and Guidebook Authors*
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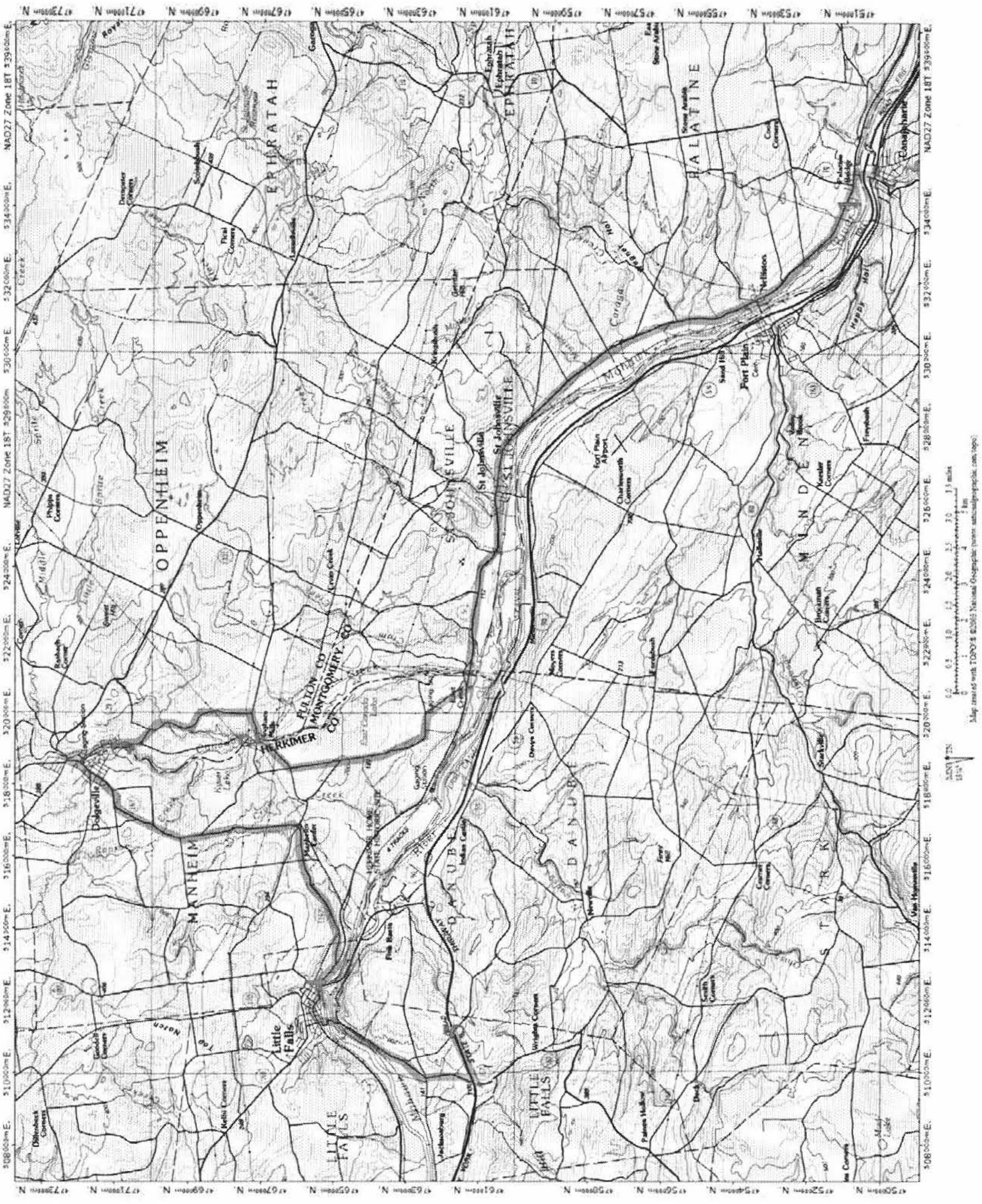
Sponsored in part by NYSERDA (New York State Energy Research and Development Authority)

October 7th, 2006

We thank STATEMAP and NYSERDA for funding research reported herein.



STOP 1: Palatine Bridge Quarry
STOP 2: Ingham Mills
STOP 3: Dolgeville Dam
STOP 4: Outcrop at Rts 5&169
STOP 5: Thruway Cut
Optional Stops also included

INTRODUCTION

The Cambro-Ordovician section in the Mohawk Valley region is one of the most well known and longest studied Cambro-Ordovician sections in the world. From Amos Eaton in the 1820s (e.g., 1824) to Vanuxem (1834, 1838, 1843) to Emmons (1842), Schuchert (e.g., Clarke and Schuchert, 1899) to Ruedemann (1897, 1925) to Cushing (1905) to Kay (1937, 1968) to the more recent geologists (e.g., Bird and Dewey, 1970; Fisher, 1977, 1979; Fisher et al., 1970; Landing et al., 1996; Jacobi 1981, 2002; Bradley and Kidd 1991; Bradley and Kusky, 1986; Cisne and Rabe, 1978; Cisne et al, 1982; Mitchell et al., 1994; Goldman et al, 1994; Lehmann et al, 1995; Brett and Baird 2002; Baird and Brett 2002), and many, many more (see review in Jacobi and Mitchell, 2002), the Cambro-Ordovician geology in the Mohawk Valley (Fig. 1) has provided critical data for understanding Cambro-Ordovician stratigraphy (Fig. 2) and the inferences that can be drawn from the stratigraphy and structure. For example, the original application of plate tectonics to land geology utilized the Mohawk Valley geology (Bird and Dewey, 1970; Fig. 3A), as have revisions to the original model (e.g., Jacobi, 1981; Fig. 3B; Bradley and Kidd, 1991; Fig. 3C). More recently, the Mohawk Valley section has functioned as the test case for detailed volcanic ash geochronology (tephrachronology) and graptolite correlations (e.g., Mitchell et al., 1994), and for gauging the relative importance of tectonic versus eustatic sea level signals on the apparent sequence stratigraphy in the basin (e.g., Mitchell et al., 1994; Lehmann, 1994; Joy et al., 2000; Jacobi and Mitchell, 2002; Brett and Baird, 2002; Baird and Brett, 2002).

During the past decade, the Mohawk Valley section has been used as an analogue for subsurface hydrocarbon reservoirs in the Cambro-Ordovician section. The major gas discoveries in the Trenton/Black River in central NYS (and elsewhere) have driven a renewed interest in the exposed Mohawk Valley Cambro-Ordovician section. The present attention involves determining and understanding the factors that controlled the development of the Cambro-Ordovician gas reservoirs.

The present field trip examines a spectacular quarry floor that is a probable analog for the dolomitized Trenton/Black River reservoirs in central NYS. Trenches, cores, GPR, and waterblasting of the outcrop led to an integrated, very comprehensive picture of how secondary porosity and dolomitization develop. The field trip also visits faults that may have been the pathways for the fluids that dissolved the limestone, resulting in vuggy porosity, and that precipitated hydrothermal dolomite and other minerals in the void space. The trip includes several optional outcrops that display the stratigraphy and the characteristic porosities in various units. Finally, the trip stops at the well-known NYS Thruway locality where the Thruway Disconformity is exposed, separating ribbon limestones of the Dolgeville Formation from the overlying black shale of the Utica (Indian Castle Member). This last stop is particularly interesting for those who have an interest in the black shales of the Utica. Because of severe time constrains, the trip cannot stop at all the localities, but we have included the optional stops for geologists to visit at a later date. Here we can examine in outcrop the stratigraphy, faults, fluid migration along these faults, and tectonics in the Mohawk Valley.

This field trip is the outgrowth of the work of two research groups, the NYS Museum Reservoir Characterization Group (e.g., Smith and Nyahay, 2005), and the UB Rock Fracture Group (e.g., Jacobi and Mitchell, 2002). Each group has worked for some time in the Mohawk Valley; the

NYS Museum Reservoir Characterization Group has investigated a hydrothermal dolomite reservoir analog in the Palatine Bridge Quarry (STOP 1); their integrated technologies yielded a significant understanding of the hydrothermal systems. The Reservoir Characterization Group led a workshop and field trip to this site, (Smith and Nyahay, 2005). The UB Rock Fracture Group and iTAS (Integrated Tectonics and Stratigraphy Group at UB) have been working on stratigraphy and structure farther west in the western part of the Mohawk Valley sequence for 25 years (albeit sporadically; e.g., Ritter, 1983; Mitchell et al., 1994; Jacobi and Mitchell, 2002).

Below we first review the Mohawk Valley Cambro-Ordovician stratigraphy that incorporates the units and structures we will inspect on this field trip. Overall, the Cambro-Ordovician sequence in the Mohawk Valley records a Laurentian passive margin carbonate bank that is overstepped by black shales and later coarser clastics. The overstep resulted primarily from a continent/island arc collision wherein the Laurentian continental passive margin first passed over the peripheral bulge and then dropped down into an east-dipping subduction zone (e.g., Jacobi, 1981; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985). A eustatic sea level rise may have been superimposed upon the tectonic signal (e.g., Brett and Baird, 2002). We also review the structure of the region, and the concepts involved in the dolomitization and development of the porosity typical of the Cambro-Ordovician carbonate section. A key component of the trip will be the link between faulting, fluid flow and reservoir development.

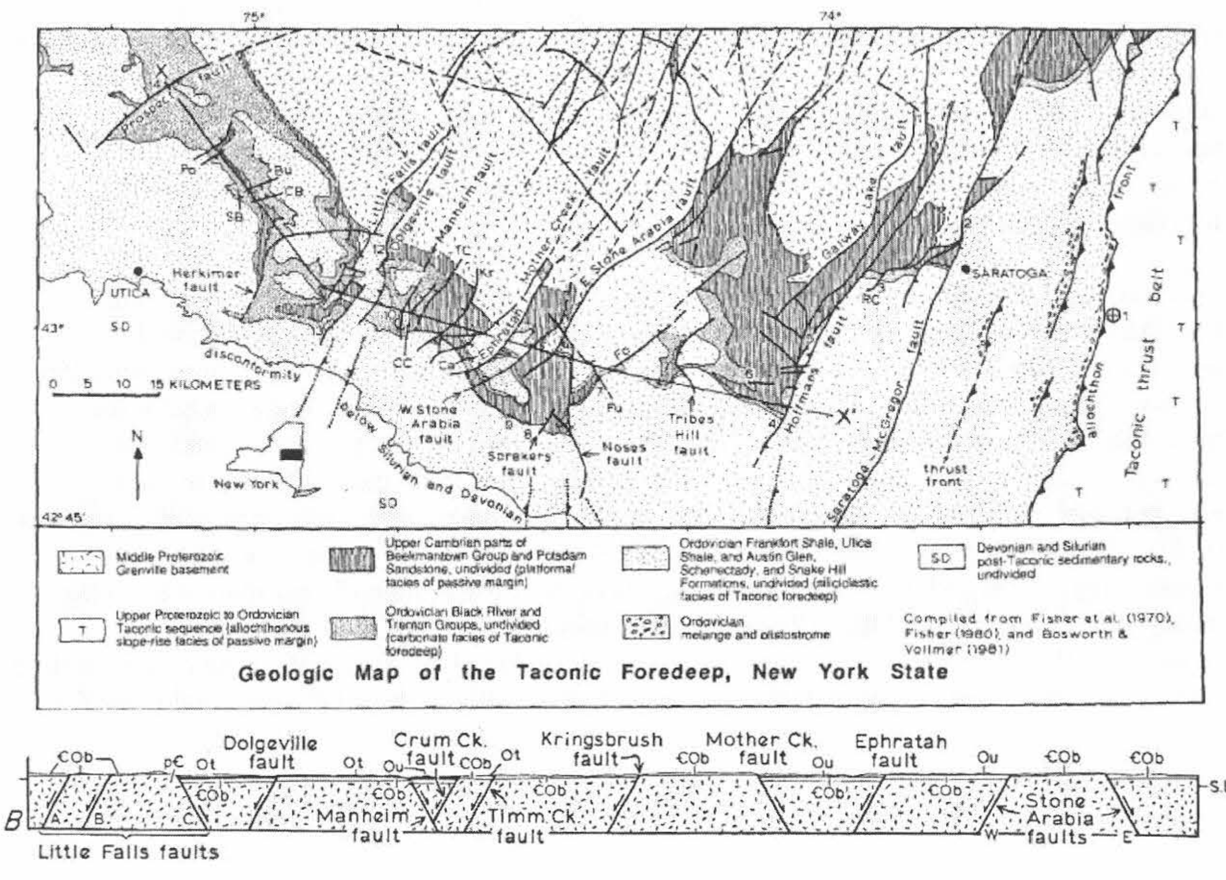


Figure 1. Geologic Map of the Mohawk Valley Region, New York State and Cross-Section across Field Trip Area. From Bradley and Kidd (1991).

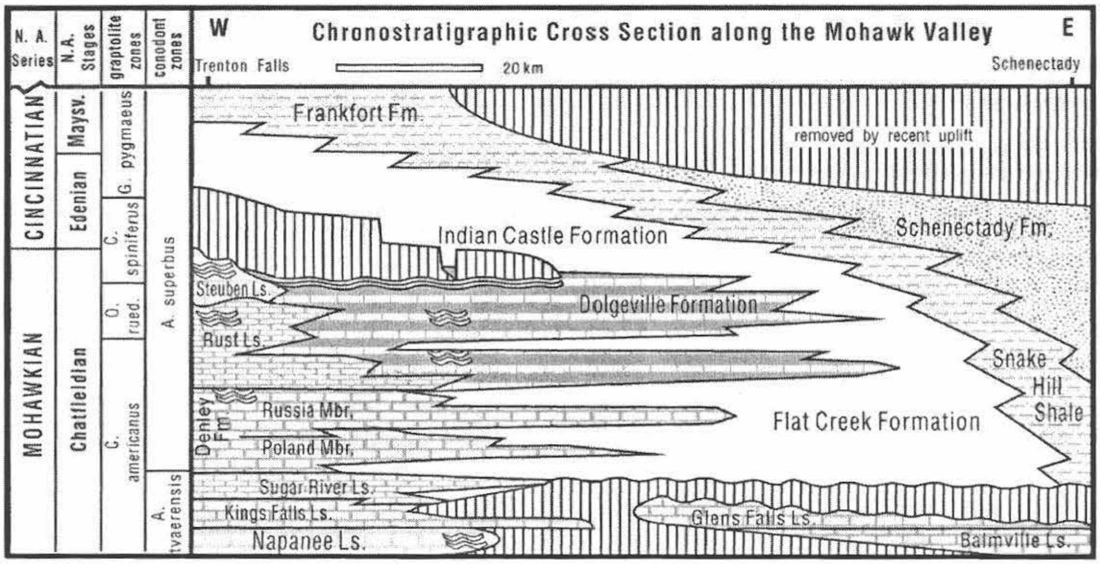


Figure 2. Chronostratigraphy of the Western Mohawk Valley, NYS. Modified from Jacobi and Mitchell (2002).

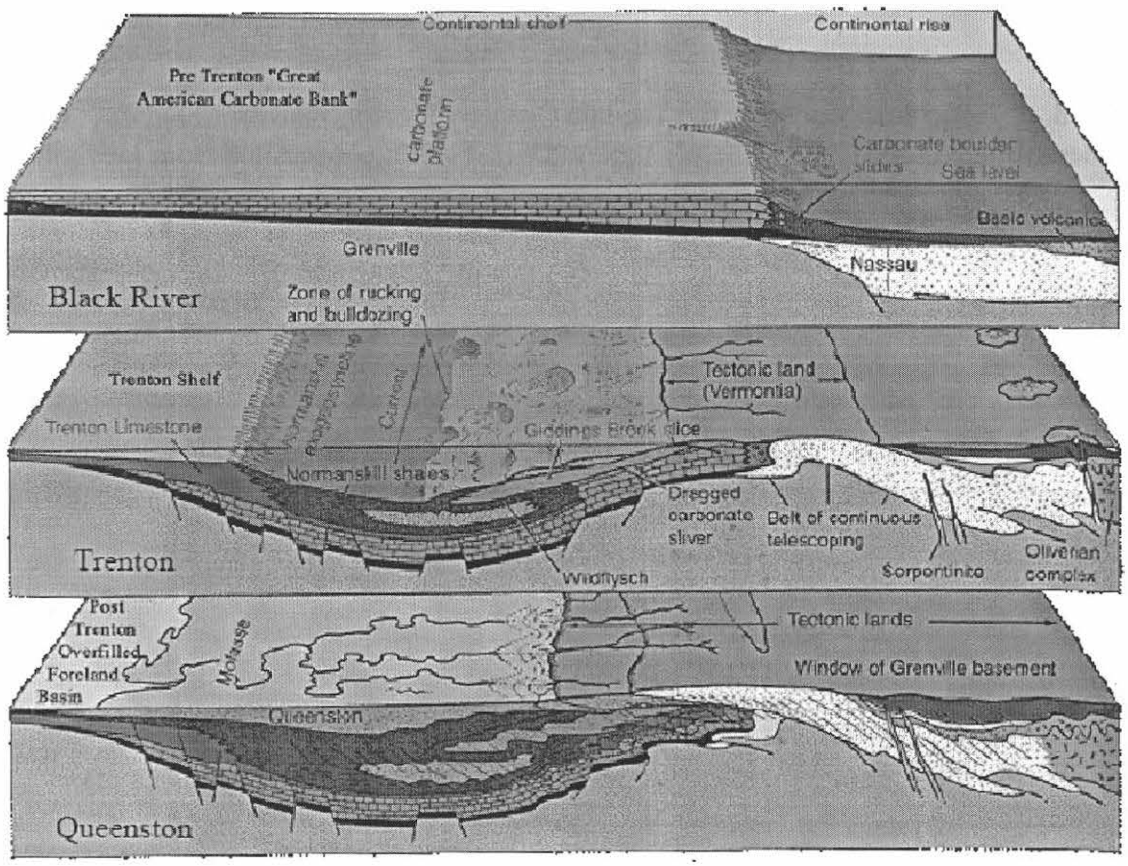


Figure 3. (A) Tectonic Model for the Taconic Orogeny that Involved Westward Subduction Located East of the Model. Modified from Bird and Dewey (1970).

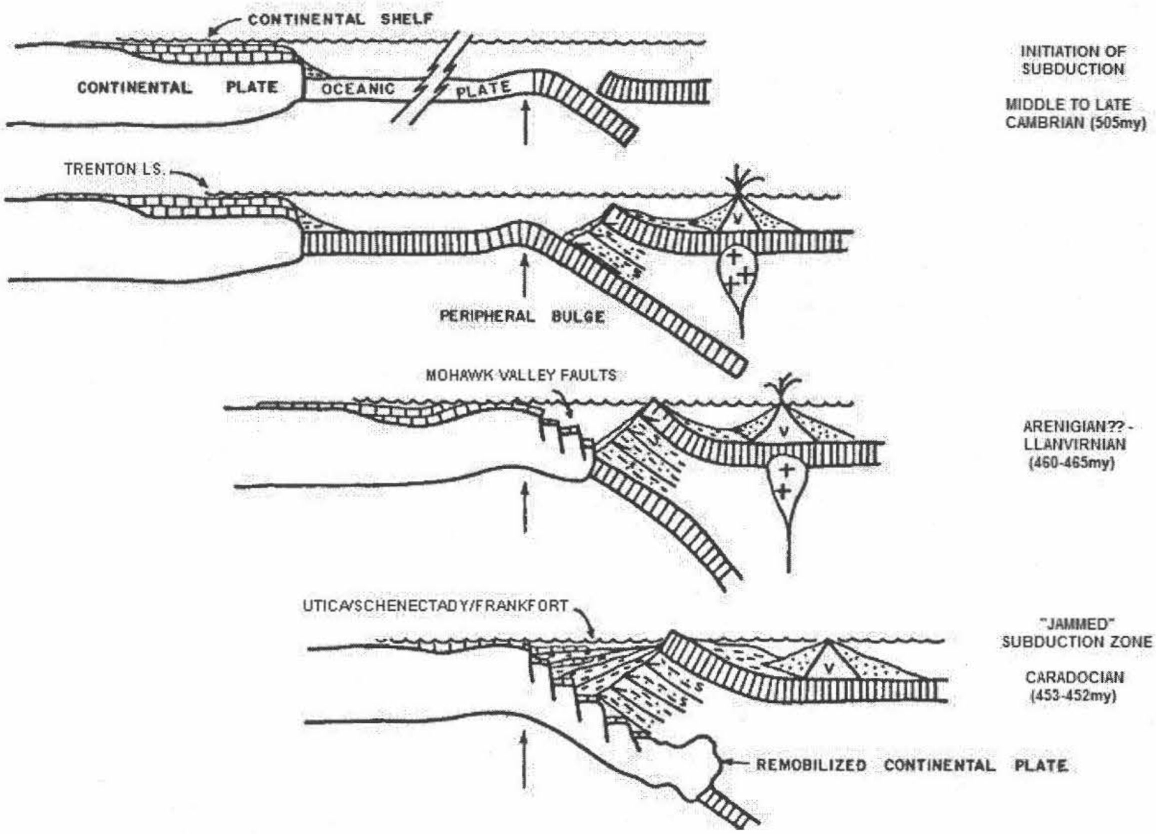


Figure 3. (B) Tectonic Model for the Taconic Orogeny Showing Eastward Subduction, Peripheral Bulge Effects, and Mohawk Valley Block-Faulting. Modified from Jacobi (1981).

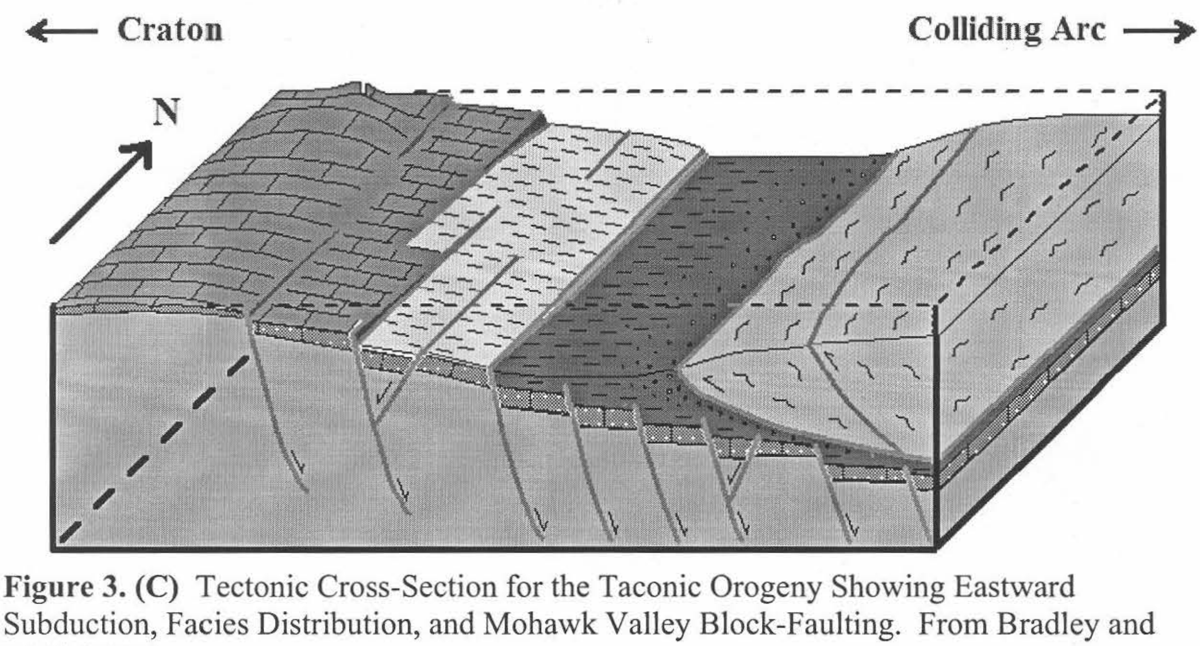


Figure 3. (C) Tectonic Cross-Section for the Taconic Orogeny Showing Eastward Subduction, Facies Distribution, and Mohawk Valley Block-Faulting. From Bradley and Kidd (1991).

CAMBRO-ORDOVICIAN STRATIGRAPHY
IN THE LITTLE FALLS AREA, MOHAWK VALLEY

Little Falls Formation (Beekmantown Group)

The Cambro-Ordovician section of the Mohawk Valley (Fig. 4) has been studied for over 180 years (Fig. 5). The first formal study of the rocks now comprising the Little Falls Formation was conducted by Amos Eaton in 1824. He assigned the rocks the name "Calciferous Sandrock". The unit held this informal lithologic designation until Clarke (1899) assigned the name Beekmantown for the exposures in Clinton County, NY. Clarke (1903) was later the first to apply the informal moniker "Little Falls Dolomite" to the rocks comprising the cliff face at Little Falls, NY. The first detailed geologic map of the Little Falls, NY area was produced by Cushing (1905), where he mapped the extent of the Beekmantown Formation (including, but not limited to the rocks now comprising the Little Falls Formation). Five years later, Ulrich and Cushing (1910) were the first to observe the age disparity between the upper and lower units of the Beekmantown Formation in the Mohawk Valley. They split the lower unit and named it the Little Falls Dolomite, while the upper unit retained the Beekmantown designation. The name Little Falls Dolomite remained in informal use until the formal designation of a type section by Zenger (1976). Since no single locality exposes the Little Falls Formation in its entirety, he described a composite type section consisting of several of the better exposures within the Little Falls 7.5' Quadrangle. Zenger (1981) provided an in-depth examination of the stratigraphy and petrology of the Little Falls Formation in its type area, in addition to a comprehensive review of previous work.

The unit consists largely of medium to dark gray, thick-bedded, medium to coarse-grained dolostone. The unit weathers to a buff color and often exhibits a "sucrosic" texture. Some beds, especially in the lower parts of the formation, contain abundant quartz grains and even feldspars. Beds near the top of the formation often have a reddish hue. In several outcrops on the fault blocks west of Little Falls Fault B (north of the city of Little Falls) the upper Little Falls Formation contains a few horizons of sediment slide (debrite/slumped) material, which average 2m in thickness. Fossils are exceedingly rare within the Little Falls Formation. Near faults, circulating hydrothermal fluids have created open vuggy porosity in susceptible layers. These vugs are variably filled with calcite, dolomite, anthraxolite, and quartz ("Herkimer Diamonds"). In the field trip area, the Little Falls Formation non-conformably overlies the Grenvillian basement and disconformably underlies the Tribes Hill Formation. The lower contact is abrupt and the upper contact, although it represents a significant time gap, can be difficult to recognize and locally appears gradational, such as at the type locality, the "Roll Away" at Little Falls (east of Little Falls Fault B). Our working definition of the "gradational contact", following Zenger (1976), is the presence of reddish staining and a distinct upsection increase in fucoidal texture (burrowing).

Tribes Hill Formation (Beekmantown Group)

The earliest studies of this unit lumped it in with the Little Falls Formation as "Calciferous Sandrock" and later Beekmantown formation, as described above. The name Tribes Hill Formation was first used by Ulrich and Cushing (1910) when they split this unit from the older

Little Falls Dolomite. The unit is named for exposures in and around Tribes Hill, NY. Fisher (1954) formally divided the formation into six distinct members. These member divisions were redefined by Landing and others (1996) in order to more easily recognize the divisions outside of the type area, although these new divisions are still difficult to recognize in the Little Falls region.

The formation consists largely of light to medium gray, thin to medium-bedded, fine to medium-grained mottled dolostones and limestones with rare shale laminations. In contrast to the underlying Little Falls Formation, fossils and bioturbation are quite common within this unit. In the field area, the Tribes Hill Formation disconformably overlies the Little Falls Formation and disconformably underlies the Lowville Formation. The upper contact is an abrupt change from limestone and dolostone to the easily recognizable limestones above.

In the eastern Mohawk Valley, the Tribes Hill Formation is variably dolomitized. At some locations it is completely dolomitized from top to bottom whereas in other locations it has very little dolomite (Landing, 1996). In some locations, it is possible to link the dolomitization to faults and fractures and this dolomitization is thought to be of a hydrothermal origin. The Tribes Hill Formation is the host of the hydrothermal dolomite reservoir analog at the quarry in Palatine Bridge (STOP1).

Lowville Formation (Black River Group)

This unit was informally referred to as the "Birdseye limestone" be early workers. Clarke and Schuchert (1899) were the first to refer to the unit as Lowville Limestone for the exposures at the railroad bridge in Lowville, NY. This formation is the only representative of the Black River Group consistently present in the field trip area. The Pamelia Formation, which underlies the Lowville Formation to the northeast is entirely absent, whereas the Watertown Formation, which overlies the Lowville Formation, appears only in a thin horizon at Ingham Mills. The Lowville Formation is perhaps the easiest to recognize in the area. It is an excellent building stone and is used in many foundations, fences, and bridges of the region, as well as many historic structures including Ingham Mills Church and Palatine Church.

The unit is a dove gray, thin to thick-bedded, very fine-grained limestone commonly exhibiting "birdseye" texture. The unit's diagnostic feature, aside from its characteristic color, is the near ubiquitous presence of the sub-vertical burrows of *Phytopsis tubulosum.* In the field area, the Lowville Formation disconformably overlies the Tribes Hill Formation and disconformably underlies the Sugar River Formation. The upper contact is readily identified by an abrupt switch in the color and texture of the limestones.

The Black River thickens to the southwest where it hosts prolific natural gas reservoirs in structurally-controlled hydrothermal dolomite. The tight shallow marine limestone facies form lateral seals around the vuggy, brecciated dolomite. A paper is included for field trip attendees by Langhorne Smith on the Black River hydrothermal dolomite reservoirs of New York.

Kings Falls and Sugar River Formations (Trenton Group)

Vanuxem (1838) was the first to formally work on the Trenton limestone at the type section of Trenton Falls. Kay (1937) was the first to divide the Trenton into smaller stratigraphic units. The rocks now comprising the Kings Falls Formation were named the Kirkfield Limestone, while the rocks of the Sugar River Formation were named the Shoreham Limestone. Later, Kay (1968) revised his stratigraphy and adopted the terms Kings Falls and Sugar River limestones to denote this part of the Trenton Group. He established the type section of the Kings Falls Formation at Kings Falls along the Deer River near Copenhagen, NY. The type section of the Sugar River was established at the excellent exposures along the Sugar River in Lewis County, NY. Both the Kings Falls Formation and Sugar River Formation outcrop in the field area, but their lithology is so similar, the outcrops so "patchy", and the section so thin that it is impractical to map at the formation level within the Little Falls area. The other limestone formations of the Trenton Group are wholly absent from the field trip area, with the exception of the section at Ingham Mills, which includes the underlying Napanee Formation.

The Kings Falls Formation consists of gray to dark gray, medium to thick-bedded, very coarsely fossiliferous limestone with calcareous shaly interbeds. This formation also commonly displays surfaces marked by large-scale ripples. The Sugar River Formation is a very similar lithology, although the beds tend to be thinner and more irregular and the grain size tends to be less coarse. Additionally, the high-energy sedimentary structures typical of the Kings Falls are largely absent in the Sugar River. In the field area, these formations lie disconformably on the Lowville Formation and conformably underlie the Flat Creek Member of the Utica Formation/Group. The upper contact is a rather abrupt switch from coarse-grained limestone to black shale.

The Trenton Limestone has produced gas from numerous shallow fields around Lake Ontario for more than a century. This gas play is very different than the hydrothermal dolomite play. Wells commonly encounter extremely high pressure gas that flows at a very high rate for a few hours or days and then decreases to almost nothing. It is the opinion of co-leaders Smith and Nyahay that this overpressured gas comes from bedding planes in the interbedded shale and limestone of the Trenton Group. The high pressure of the gas may be enough to open the bedding plane until it starts producing at which point the bedding planes close, cutting off the flow of gas.

Dolgeville Formation (Trenton Group)

The Dolgeville Formation was known as the "Trenton-Utica passage beds" by early workers. Cushing (1909) was the first to suggest that these beds be mapped as a distinct unit, not merely as the gradation between the Trenton and the Utica Groups. He established the type section as the well-exposed beds near Dolgeville, NY, in East Canada Creek. Goldman and others (1994) noted that this unit splits the Utica Group into two "tongues", a lower (Flat Creek Member) and an upper (Indian Castle Member). They stated that the Dolgeville is age equivalent to the Denley Formation (upper Trenton), younger than was previously believed. The Dolgeville is now believed to correlate with the Rust (which overlies the Denley; Jacobi and Mitchell, 2002; Brett and Baird, 2002; Baird and Brett, 2002). As its early moniker suggests, the Dolgeville Formation is intermediate between the limestones of the upper Trenton Group and the black shales of the Utica Group. A typical outcrop consists of "ribbon" limestones with rhythmic interbeds of black shale.

The ribbon limestones consist of dark gray to black, thin to medium-bedded, fine-grained limestone. Some of the limestones display graded bedding, planar laminae, climbing ripples, and striations on the sole of the beds, all consistent with a turbiditic origin for these ribbon limestones (e.g., Mehrtens, 1989; Jacobi and Mitchell, 2002). At several localities, including Nowadaga Creek, Crum Creek, and the Thruway road cut near Little Falls, the upper Dolgeville is dramatically folded and truncated by a surface termed the Thruway Discontinuity by Baird and others (1992). The significance of the Thruway Disconformity has been debated (Fischer, 1979; Baird and Brett, 2002; Brett and Baird, 2002; see review in Jacobi and Mitchell, 2002), but it is probable that the sharp erosive contact between the undisturbed black shale of the Indian Castle Member of the Utica Group and the underlying folded Dolgeville shales and ribbon limestones marks a slide scar. In the field area, the Dolgeville Formation conformably overlies the Flat Creek Member. The disconformable upper contact approaches a gradational conformity to the east and in the Dolgeville graben at Dolgeville, where a lack of sediment sliding apparently preserved the original gradational contact.

Flat Creek Member (Utica Group)

The rocks now comprising the Utica Group were first properly named by Emmons (1842). Attempts to subdivide the thick mass of black shale proved difficult because of the uniform lithology. Kay (1937) proposed divisions based on graptolite biostratigraphy, but these divisions were later abandoned by Fisher (1977) on the basis that they were not recognizable lithostratigraphic divisions. The unit now known as the Flat Creek Member is the stratigraphically lowest "tongue" of black shale that extends westward below the Dolgeville Formation. Fisher (1979) proposed a simple correlation whereby the carbonates of the Trenton Group grade eastward into the mixed carbonates and shales of the Dolgeville Formation and that this unit then grades into the black shale of the Utica Group. Using k-bentonite and biostratigraphic correlation, Goldman and others (1994) demonstrated that the Utica shale was both younger and older than the Dolgeville Formation. They proposed the name Flat Creek Member for the lower tongue of black shale below the Dolgeville Formation and established the type section at Flat Creek. This unit consists of calcareous black shale, indistinguishable lithologically from that of the Indian Castle Member, which lies above the Dolgeville Formation. In places, such as at Wintergreen Park, the unit consists of flaggy limestones and interbedded shales. In the field area, this unit conformably overlies the Sugar River Formation and conformably underlies the Dolgeville Formation. The upper contact is gradational and was defined by Goldman and others (1994) as the point where ribbon limestones comprise less than 20% of the unit.

Indian Castle Member (Utica Group)

When Goldman and others (1994) noted that the Dolgeville Formation divided the Utica into two tongues in the Little Falls area, the lower tongue was named the Flat Creek Member and the upper tongue was named the Indian Castle Member. Previously, the rocks of the Indian Castle Member were known simply as the Utica shale. The type section of this unit was established at

Nowadaga Creek (Goldman et al, 1994). The Indian Castle Member is lithologically indistinguishable from the Flat Creek Member, though it is younger and lies above the Dolgeville Formation. This unit consists of calcareous black shale with rare thin limestone beds. In the field area, this unit disconformably overlies the Dolgeville Formation and conformably underlies the Frankfort Formation. The upper contact is gradational and defined by the increasing silt content.

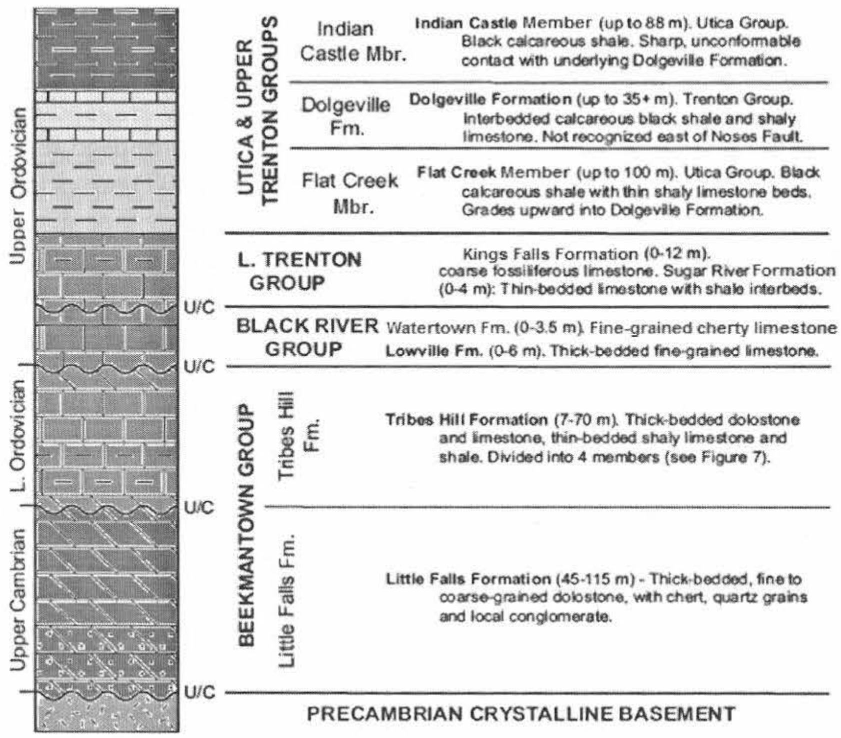


Figure 4. Stratigraphic Section of the Western Mohawk Valley and Unit Descriptions. Modified from Cross (2004).

Eaton 1824	Vanuxem 1842	Cushing 1905	Ulrich and Cushing 1910	Kay 1937	Fisher 1954	Kay 1968	Zenger 1981	Goldman and others 1994	Landing and others 1996
Greywacke	Utica Slate	Utica Shale	Utica Shale Dolgeville Utica Shale	Utica Sh Canajoharie Sh	Not Studied	Utica Sh Canajoharie Sh	Not Studied	Indian Castle Dolgeville Flat Creek	Not Studied
Metalliferous Limerock	Trenton LS	Trenton LS	Trenton LS	Shoreham LS Kirkfield LS Rockland LS		Sugar River LS Kings Falls LS Napanee LS		Sugar River LS Kings Falls LS Napanee LS	
Sparry Limerock	Birdseye LS	Black River LS Lowville LS	Black River LS Lowville LS	Chaumont LS Lowville LS		Watertown LS Gull River LS		Not Studied	
Calciferous Sandrock	Fucoidal Layers	Beekmantown Formation	Tribes Hill LS and DT	Tribes Hill LS and DT	Chuctununda Fonda Mbr Wolf Hollow Cranesville Palatine Bridge FortJohnson	Not Studied			Canyon Road Wolf Hollow Van Wie Mbr Sprakers Mbr
	Calciferous Sandrock		Little Falls Dolomite	Little Falls Dolomite	Not Studied		Unit A Unit B Unit C Unit D		Not Studied
Primary	Primary	Precambrian	Precambrian	Precambrian			Precambrian		

Figure 5. Stratigraphic Nomenclature of the Rocks of the Mohawk Valley

FAULTING IN THE MOHAWK VALLEY

Introduction

Faults in the Mohawk Valley, were recognized by the earliest geologists in the region; the faults with the largest stratigraphic offset were mentioned in Vanuxem's report of the third geological district (1842). Later workers examined these faults in greater detail (Darton, 1895; Cushing, 1905; Megathlin, 1938; Dunn, 1954, Fisher, 1954; Fisher et al, 1970; Fisher, 1980; Bradley and Kidd, 1991, see review by Jacobi and Mitchell, 2002). The apparent stratigraphic offsets and dips on exposed faults suggested to the early geologists that the faults in the Mohawk Valley are, to the first approximation, normal faults (although kinematic indicators were, and continue to be, elusive). However, using primarily fault map patterns, Shaw (1993) proposed that similar faults to the north in Canada were strike slip faults. Jacobi and Mitchell (2002) provided evidence for dip-slip, as well as oblique slip, motion on faults in the Mohawk Valley (see discussion below). Further, the en echelon nature of the dolomitization "pods" (and other considerations) at the first stop of this trip, Palatine Bridge, suggested to Smith and Nyahay (2004, 2005) that WNW-striking faults were strike slip.

In the field trip area, the faults strike in three distinct orientations: NNE, N, and WNW. The NNE-striking faults parallel gravity anomalies and probably formed during Eocambrian Iapetan opening and/or Cambrian Rome Trough development (Jacobi, 2002; Jacobi et al, 2004, 2005). These faults were apparently reactivated during Taconic times, when the major motion on the faults occurred. Offset Ordovician contacts, Ordovician growth fault geometries, and stratigraphically-bound breccias from fault scarps indicate Taconic motion (see reviews in Bradley and Kidd, 1991; Jacobi and Mitchell, 2002). The NNE-striking faults experienced oblique slip, based on bedding dip, slickenlines (STOP 4), and tectonic model considerations (Jacobi and Mitchell, 2002; Jacobi et al., 2004, 2005). The WNW-trending faults are postulated to have formed as transfer zones between the NNE Iapetan faults (Jacobi, 2002), and follow older Precambrian trends (Jacobi and Smith, 2000). The WNW-striking faults were also reactivated during the Taconic. Jacobi (2002) showed that the N-trending faults cross gravity anomalies and are assumed to have initiated during Taconic time as a result of trench/peripheral bulge tectonics. The northerly-trending Dolgeville Fault (STOP 3) displays down-dip slickenlines and drag folds with horizontal plunges, all of which support a down-dip slip on this fault (Jacobi and Mitchell, 2002).

The Taconic extensional faulting resulted from continental plate flexure over the peripheral bulge and into the trench/subduction zone (e.g., Jacobi, 1981; Bradley and Kidd, 1991; Jacobi and Mitchell, 2002; Baird and Brett, 2002). As the Taconic collision continued, it is possible that some of these faults reversed motion (Jacobi et al., 2004, 2005).

History of Investigations on the Fault Timing

The debate on timing of initiation of faulting in the region has a long history (see reviews in Bradley and Kidd, 1991; and Jacobi and Mitchell, 2002). Cushing (1905) first suggested that the faults may be as old as the Taconic "disturbance". Megathlin (1938) stated that the normal offsets on the faults suggest a tensional stress regime and thus the faults cannot be associated with the compressive stages of any "revolution"; rather, they must have occurred during a subsequent relaxation phase making them post-Taconic. This apparent quandary is resolved by the peripheral bulge flexure models that have been advanced by more recent workers (Jacobi, 1981; Bradley and Kidd, 1991) that allow for the seemingly contradictory co-existence of regional extension (and normal faulting) in the upper crust and far-field tectonic compression.

A variety of evidence indicates that faulting was active during deposition of the Ordovician units. Fisher (1954) suggested that the "disturbance" may have initiated in early Tribes Hill time, citing folding in the Tribes Hill Formation and thinning of the Palatine Bridge Member (now called Sprakers Member) near major faults. Cisne et al. (1982) examined the distribution of graptolite biozones and facies in the fault blocked region and concluded that local tectonics had a major influence on local facies patterns and basin bathymetry. Fisher (1979) noted the westward vergence of folds within the Dolgeville Formation, rather than the expected eastward direction down the inferred paleoslope into the foreland basin. Jacobi and Mitchell (2002) also observed anomalous fold directions and paleoflow directions at several locations in the Mohawk Valley, and attributed the inferred anomalous paleoslopes to local fault block rotations and bathymetric reorganization. They also cited changes in stratigraphic thickness as evidence of the local significance of these faults during deposition. In contrast, Brett and Baird (2002) believed that the faults exerted only minor control, if any, on deposition. They proposed that the faults had no surface expression at the time of deposition, based on the persistence of marker beds and facies gradations. They suggested that broad tectonic warping and eustatic sea-level changes were the dominant influence on deposition.

There is some evidence for an earlier episode of faulting in the Late Cambrian/earliest Ordovician including the debrites at the top of the Little Falls Formation, the distinct chemistry of the vein and vug fills in the Little Falls Formation (Cross et al, 2004), and growth faults in the Little Falls Formation that die out up section (Smith and Nyahay, 2005). Jacobi et al. (2006) suggested that these deformation and fluid circulation events were related to the initiation of subduction.

Jacobi (2002) suggested that the NNE-striking faults initiated during Iapetan opening time (and/or Rome trough development time), based on map patterns of aeromagnetic, gravity, and satellite image lineaments. He traced NNE-striking Iapetan opening faults from Pennsylvania where they arc through the Pennsylvania Salient to the ENE-tending basement faults in central NYS. From there eastward, it is possible that the ENE-striking faults swing once again, this time more northerly into the NNE-tends of the Mohawk Valley faults. No Iapetan-opening sediments have been recognized in the fault basins of the Mohawk Valley. However, following Burke and Dewey (1973), Jacobi et al. (2005) proposed that an Iapetan-opening mantle plume was located southeast of the New York Promontory. Such a plume could account for the major bend in the Iapetan trends, and for the lack of Iapetan-opening fault basin sediments along the Mohawk Valley faults, since the Mohawk Valley region would have been near the center of the mantle plume dome, and so would experience primarily erosion, not deposition.

The youngest age of motion on the faults was assumed to be of Utica age. All the geological investigations recognized that the faults clearly offset the carbonate/black shale contact, but the lack of markers within the Utica, the lack of critical outcrop in the stratigraphically (and topographically) higher units to the south, and the lack of observed significant offset of younger units led to the assumption that motion on the faults ceased in Utica time (e.g., Fisher et al., 1970; Fisher, 1980; Bradley and Kidd, 1991). Indeed, on geology maps, the faults were portrayed as not offsetting the Ordovician/Silurian contact south of the Mohawk Valley; rather, the faults end in the Utica (e.g., Fisher et al., 1970; Fisher, 1980; Bradley and Kidd, 1991). However, to the south, Jacobi and Smith (2000) proposed that the NNE-trending faults were reactivated in Silurian time, and controlled the depositional limits of the Clinton Group, Vernon Formation, and Cobleskill Formation. Each of these units pinches out at a location that is on strike with particular Mohawk Valley faults to the north. Further, well logs south of the Mohawk Valley display anomalous thickness changes across the southward projection of the Mohawk Valley faults (Jacobi and Smith, 2000). Jacobi and Smith (2000) also found that the Devonian Onondaga Formation to the south was disposed in monoclines, and faulted, on strike with the NNE-striking Mohawk Valley faults. The conclusion is that these faults were reactivated in the Silurian and the Devonian (and perhaps in the Alleghanian) times. The minimum age of motion along the Mohawk Valley faults is provided by the age of an undeformed peridotite dike within the Manheim Fault zone (130 ±10 my; Fisher, 1980).

It is apparent that the faults of the Mohawk Valley have a long and complex history in which these zones of weakness were reactivated at many different intervals throughout the history of the basin. It is probable that their sense of motion would shift with changes in paleostress orientations.

MINERALIZATION OF THE MOHAWK VALLEY

Mineralization in and around fractures and faults is of great scientific and economic interest. Many carbonate hosted ore deposits occur around faults interpreted to be conduits for upward flowing hydrothermal fluids. Hydrothermal dolomite formed around faults and fractures hosts prolific oil and gas reservoirs in the Ordovician and Devonian carbonates of the Appalachian and Michigan Basins.

In the past decade, more than twenty new natural gas fields have been discovered in laterally discontinuous hydrothermal dolomites of the Upper Ordovician Black River Group in south-central New York. The dolomites form around basement-rooted wrench faults that are detectable on seismic data. Most fields occur in and around elongate fault-bounded structural lows interpreted to be negative flower structures (Smith, 2006). Away from these faults, the formation is composed of impermeable limestone and forms the lateral seal for the reservoirs. In most cases the faults die out within the overlying Trenton Limestone and Utica Shale. Most porosity occurs in saddle dolomite coated vugs, breccias and fractured zones. Matrix porosity is rare in the Black River cores in the New York State Museum collection.

The patchy distribution around basement-rooted faults and geochemical and fluid inclusion analyses support a fault-related hydrothermal origin for the saddle and matrix dolomites. This play went for many years without detection because of its unconventional structural setting (i.e. structural lows versus highs).

Hydrothermal fluid flow is thought to be most common while faults are active and much less common during periods of tectonic quiescence (Sibson, 1990, 2000; Davies, 2001; Knipe, 1993; Muir-Wood, 1994; Davies and Smith, 2006). The timing of hydrothermal alteration is closely linked to the timing of fault movement. In the case of the Black River hydrothermal dolomite reservoirs it appears that the faults primarily moved during the Late Ordovician Taconic Orogeny when the Trenton and Black River Groups were still buried to very shallow depths (less than a kilometer). This fits with the depth at which many other hydrothermal dolomite reservoirs and ore deposits are thought to have formed (Davies and Smith, 2006).

The exposure of en echelon dolomitized bodies in the Palatine Bridge Quarry (STOP 1) is interpreted to be a scaled analog for the Trenton Black River reservoirs. Most of the faults that have associated dolomitization in the Mohawk Valley also appear to have been active during Trenton and Utica time (during the Late Ordovician Taconic Orogeny) and largely inactive after that time. On seismic data, most dolomitized wrench faults in New York die out in the Trenton and Utica and sags are commonly filled in during Trenton or Utica time. This suggests that most of these faults were active during the Taconic Orogeny but were not reactivated during subsequent mountain building events.

Smith (2006) has presented the following model for the development of the Black River hydrothermal dolomite reservoirs (Fig. 6). This model is speculative, but is supported by all of the known facts at this time. Black River Group carbonates were deposited on relatively stable craton (Fig. 6A). Major collision between North America and Volcanic Island Arc begins during earliest Trenton time and continues through the Late Ordovician and into the Silurian (Ettensohn and Brett, 2002). This collision led to reactivation of appropriately oriented older faults or activation of new faults. Near the thrust front, thrust loading led to normal faulting oriented subparallel to orogenic belt. In the more distal parts of the craton, strike-slip faulting is initiated along appropriately oriented faults.

High-pressure, high-temperature fluids flowed up active basement-rooted strike-slip and transtensional faults (particularly in dilational parts of fault zones) during the time of Trenton and Utica deposition, hit low permeability beds at the base of the Trenton and flowed out laterally into the more permeable limestones of the Black River. Cooling hydrothermal fluids leached the limestone and produced vugs in a migrating front moving away from the fault zone (Fig. 6B). As permeability was enhanced by fracturing and leaching, warmer dolomite-supersaturated fluids migrated farther from fault zone precipitating dolomite (Fig. 6C). These fluids first produced a halo of matrix dolomite, particularly on the downthrown sides of faults in negative flower structures. Because the fluids flowed up from greater depths where pressures are higher, the elevated pressure of the fluids may have led to hydro-fracturing (*sensu* Phillips, 1972), enlargement of existing fractures, and further brecciation. Some dissolution vugs may have formed prior to and during matrix dolomitization. Matrix dolomitization was followed by further fracturing, brecciation and vug development as tectonic activity continued (Fig. 6D). Fractures and vugs were lined or filled with saddle dolomite soon after their formation. This later mineralization occurred during active fracturing as is demonstrated by episodic filling of fractures as they opened.

As time passed, fluids evolved and precipitated a range of other minerals including quartz, bitumen, sulfides and calcite. Bitumen may have formed when kerogen within the altered formation and near the faults was heated by the hydrothermal fluids and small quantities of oil formed that coated some pores and fractures ("forced maturation" of Davies, 2001). If the faulting was over by Late Ordovician or Early Silurian time (as it appears to be on many seismic lines), that would make most or all of the diagenesis Late Ordovician to Early Silurian in age. If the faulting continued or recurred during the Devonian Acadian or Pennsylvanian Alleghanian Orogenies some of the later stages of mineralization may have occurred during those times. Some calcite cementation may have occurred during later pressure solution of the adjacent limestones under normal burial conditions.

A similar sequence of events is interpreted for much of the mineralization of faults and fractures in the Mohawk Valley. Faults with at least a component of strike-slip are much more likely to be mineralized by hydrothermal fluids, especially at dilational jogs along the faults.

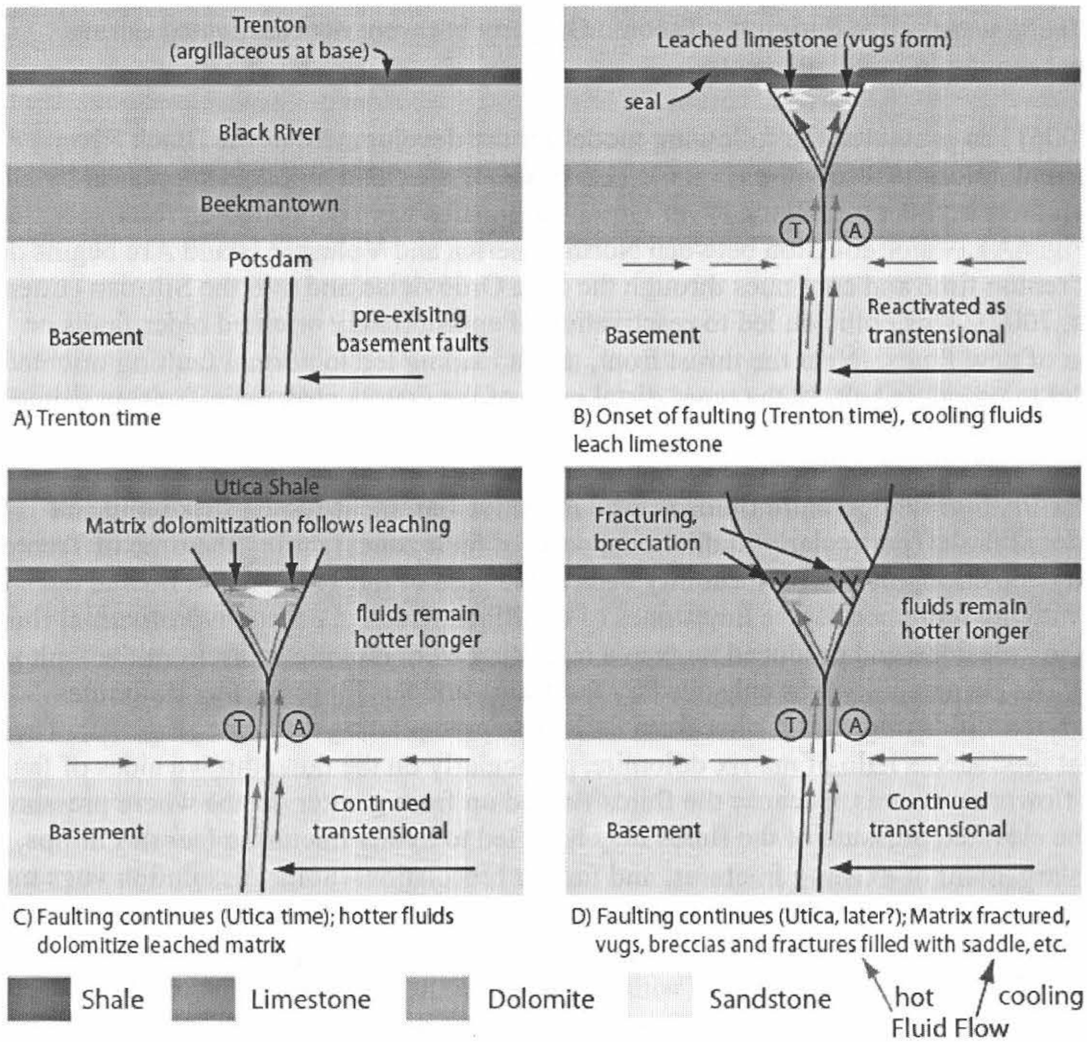


Figure 6. Schematic Model for Development of Black River Hydrothermal Dolomite Reservoirs. From Smith (2006)

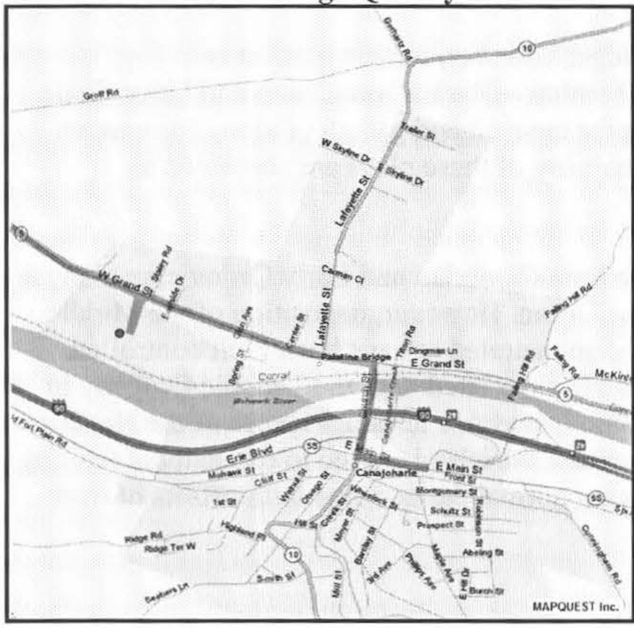
FIELD TRIP STOPS

Field Trip begins at Canajoharie Exit (#29) of New York State Thruway (I-90)

	Leg	Cum.
Right onto E. Main St. (NY-5S)	**0.2 mi**	**0.2mi**
Right onto Church St. (NY-10)	**0.3mi**	**0.5mi**
Left onto W. Grand St. (NY-5/NY-10)	**1.1mi**	**1.6mi**
Left onto unnamed access road near Palatine Motel	**<0.1mi**	**1.6mi**

Park at end of access road if gate is open (permission is needed)

STOP 1: Palatine Bridge Quarry **GPS (NAD27): 533685E 4750885N**



The Palatine Bridge study site is located in an inactive quarry along the Mohawk Valley, approximately 50 miles west of Albany, NY. The outcrop occurs as a highly localized dolomite alteration in the Tribes Hill limestone that makes up the floor of the quarry. Although there are at least three areas of the quarry that have been dolomitized, research has been focused on the one region most accessible. The site consists of two dolomite bodies (Figs 7&8). The eastern most body (body 1) is a long, linear feature approximately 55 feet long, 5 feet wide, and has a strike of 305°. The second body (body 2) is located immediately west of the first, and is also a linear feature with similar strike, however this body is substantially longer than the first, measuring 110 ft. Body 2 is broken into two parts which are separated by a southerly bend, referred to as the jog. The dolomite is easily distinguished from the surrounding limestone by its distinct orange color and massive texture.

Analog Comparison

Although it is much smaller than the hydrothermal dolomite fields in production today, all other characteristics of the quarry are virtually identical to those of its larger counterparts (Fig. 9). Hydrothermal dolomite fields such as the Rochester field of Ontario, CAN, the Albion-Scipio fields of Michigan, and the numerous gas fields of South Central NY all bear striking resemblance to the Palatine Bridge Quarry outcrop.

Setting

The Palatine Bridge Quarry lies within the Tribes Hill unit of the Upper Beekmantown Group. It is separated from the Precambrian basement by the Little Falls dolostone (lower Beekmantown Group) and the Potsdam sandstone which may or may not be present as it thins to the east. The Tribes Hill is a thin to medium-bedded limestone which was deposited in a shallow, peritidal to subtidal environment. It is underlain by the Late Cambrian Little Falls dolostone which is described as a thick-bedded dolostone and is also believed to have formed by deposition in a shallow, peritidal to subtidal environment. The Little Falls is approximately 400 ft. thick in this portion of the Mohawk Valley.

The Tribes Hill Limestone outcrops in Palatine Bridge; however, further south it is unconformably overlain by the Upper Ordovician Trenton - Black River Group and Utica Shale. The Trenton and Black River units are well known for their oil and gas plays along the eastern portion of the US and into southern Ontario. The majority of these plays are contained in hydrothermal dolomite reservoirs.

The Beekmantown Group was deposited during the Late Cambrian and Early Ordovician along a passive margin following the opening of the Iapetus Ocean. However, deposition of the Middle Ordovician Trenton - Black River Group has been demonstrated to have been fault controlled. Active tectonism during the Middle-Late Ordovician is associated with the Taconic Orogeny in which Proto-North America collided with the Taconic Island Arc terrain. Flexure of the North American plate during its attempt to be sub ducted under the island arc led to extensional faulting and produced a series of north east-south west striking normal faults, some with offsets of over 1000ft.

Field Relations

There are many characteristics of the outcrop that indicate a fault related hydrothermal origin. The central sag flanked by monoclines is a typical feature of many hydrothermal structures (Fig. 10). Fault breccia and vuggy porosity can be observed in several places (Fig. 11). These areas are commonly located near, but not limited to, the tips of each body. An intense fault and fracture zone surrounds the entire outcrop increasing the width of the affected area to approximately 15 ft. (Fig. 12). Many faults have vertical slickenlines which indicate a stress regime that was predominantly extensional. Groups of faults commonly join to form a relay ramp structure as seen in Figure 13.

Ground Penetrating Radar

In the spring of 2004 a shallow ground penetrating radar survey was run over body 2. The purpose of this survey was to aid in the excavation process and to get a profile view of the outcrop. Although the radar was only able to penetrate a couple meters, it showed that the outcrop has geometry very similar to that of the Trenton / Black River gas plays (Fig. 14). Each body consists of a central syncline, or sag, that is flanked by smaller anticlines on either side.

Cores

A series of six cores 40 - 60 ft. deep were taken from strategic points on the outcrop so that cross sections could be constructed both along strike and perpendicular to the bodies. The perpendicular cross section (Fig. 15) demonstrates that the dolomitization is highly localized as Hole Six, less than 20 ft. outside the body, is almost completely undolomitized. The parallel cross section (Fig. 16) shows that the limestone gap between bodies 1 and 2 is only 3 feet thick. The two bodies join at depth.

Trenches

Six trenches measuring 3'wide by 15'long by 18"deep have been cut across the outcrop (Fig. 17) to expose the folds of the body as well as the dip and offset of the fractures (Fig. 18).

Paragenic Sequence

Examination of thin sections made from outcrop and core samples indicate a fairly complicated diagenetic history including multiple episodes of faulting and mineral precipitation (Fig. 19).

Geochemistry

Strontium isotope analyses done on dolomite samples yield values significantly higher than that of seawater during the time of Tribes Hill deposition (Fig. 20&21). Limestone and calcite samples plot on the curve as expected. The elevated strontium content is interpreted to be caused by interaction of the parent fluid with silici-clastic rocks, such as the Grenville basement, for an extended period of time.

Stable Isotope analyses show that dolostone crystals formed from a fluid much lighter than seawater dolomites (Fig. 22). It is believed that these lighter isotope ratios are an indication of higher temperature fluids traveling upward from deeper, hotter depths.

Fluid inclusion analyses conducted on 40 samples indicate that both dolomite and calcite formed at temperatures greater than 100°C (Fig. 23). Although thermal chronology studies indicate that the Tribes Hill has been buried to a geothermal temperature of over 100°C, core analysis and paragenic sequencing suggest that dolomitization occurred at a shallow depth before significant burial.

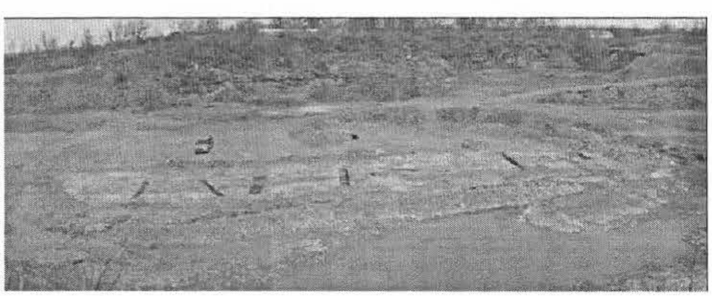


Figure 7. View of the Entire Outcrop (looking north).

Figure 8. View of the Entire Outcrop (looking west).

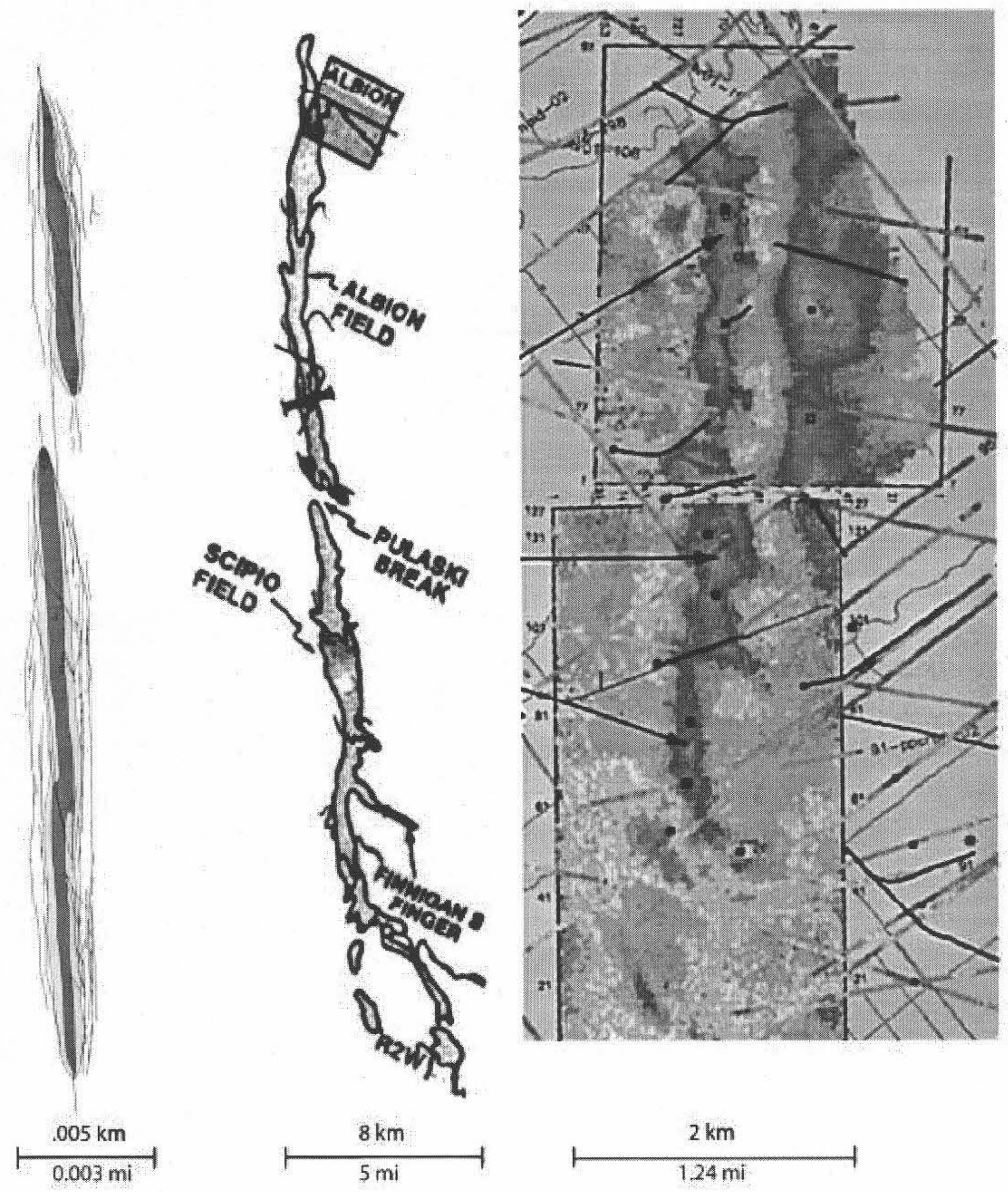


Figure 9. Comparison of Palatine Quarry Dolomite (left) with Albion-Scipio Play in Michigan (center) and 3-D Seismic Line of Rochester Field in Ontario. Not to Scale

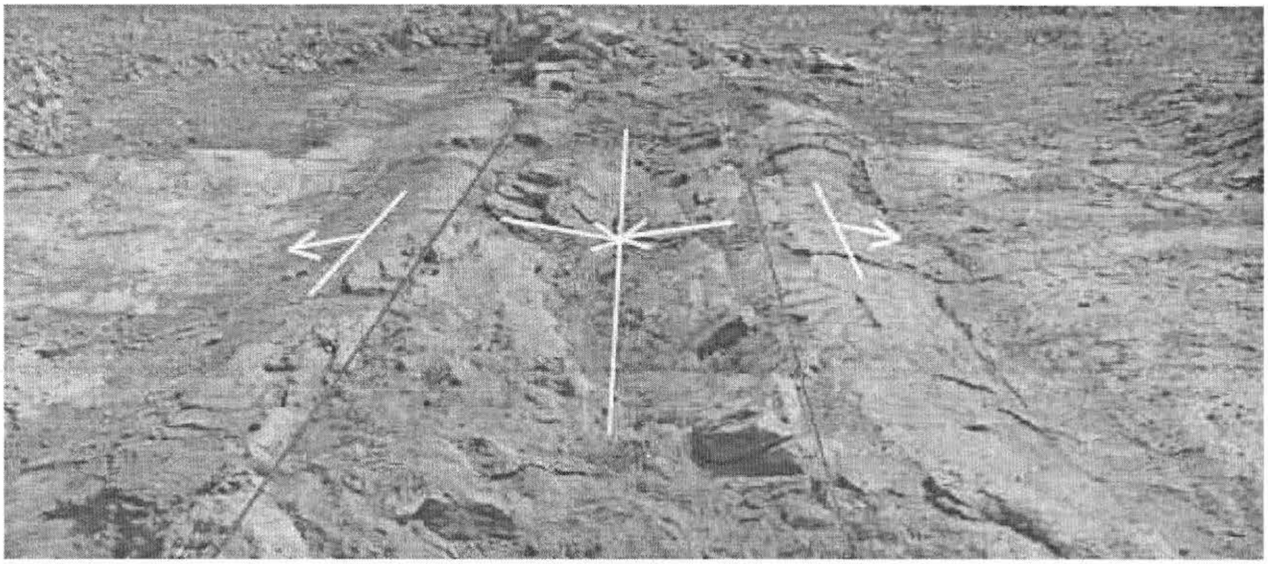


Figure 10. Central Sag Flanked by Monoclines

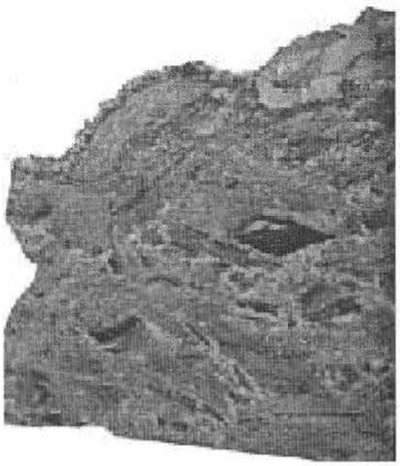


Figure 11. Fault Breccia (left) and Dolomitized and Calcite-filled Vug (right)

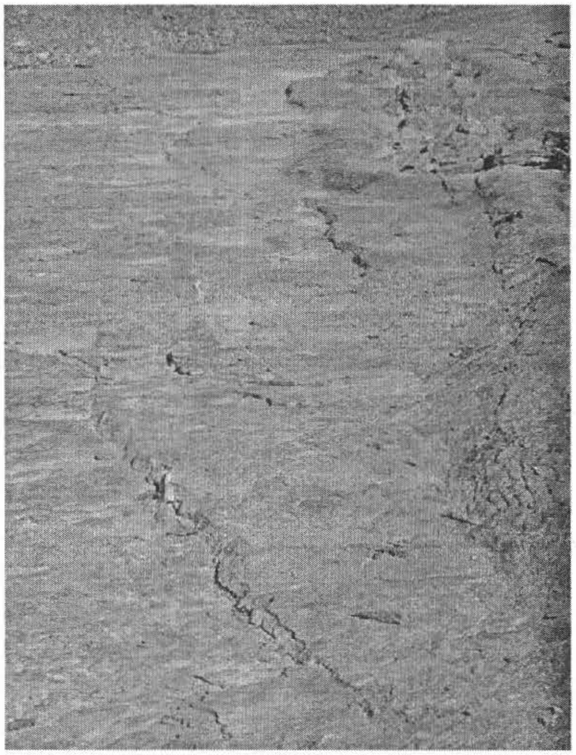


Figure 12. Fault and Fracture Zone Surrounding the Dolomite Exposure

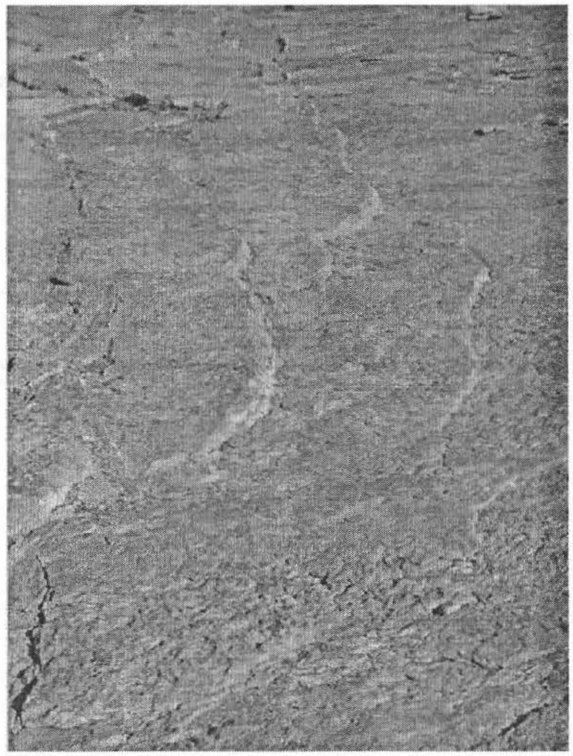


Figure 13. Relay Ramp Structures

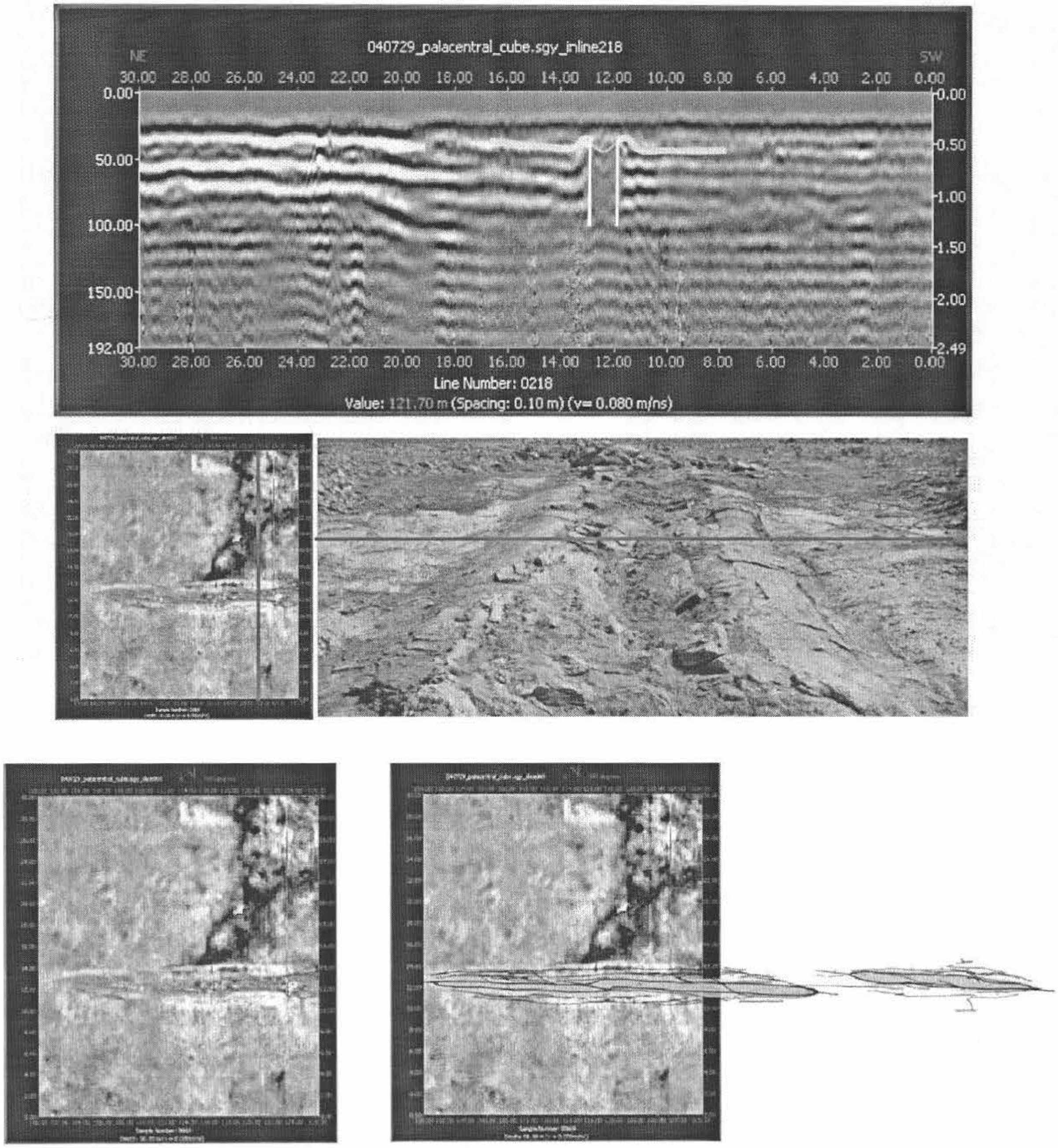


Figure 14. A Compilation of GPR Images

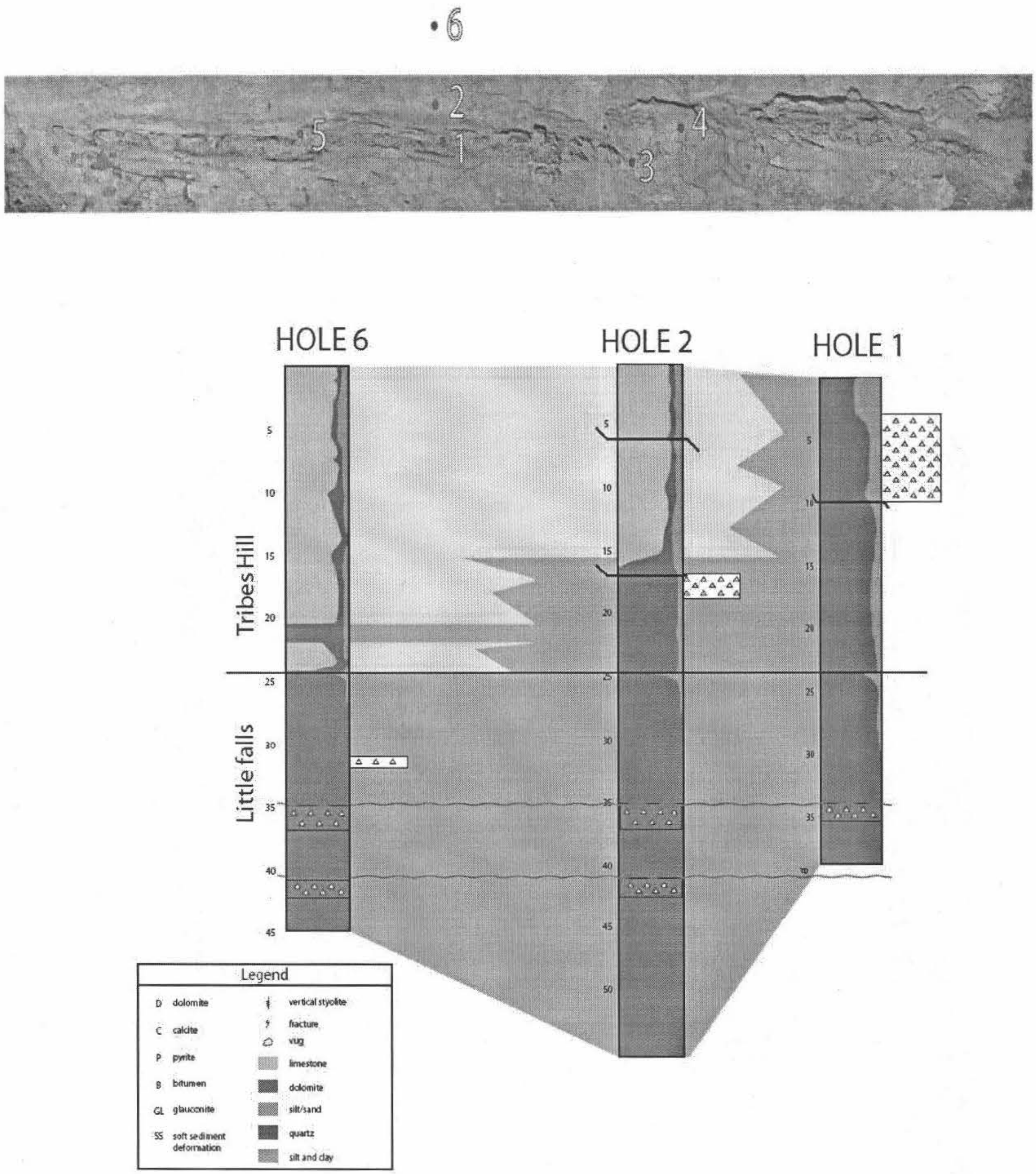


Figure 15. Cross-Section Perpendicular to Body Across Holes 1, 2, and 6

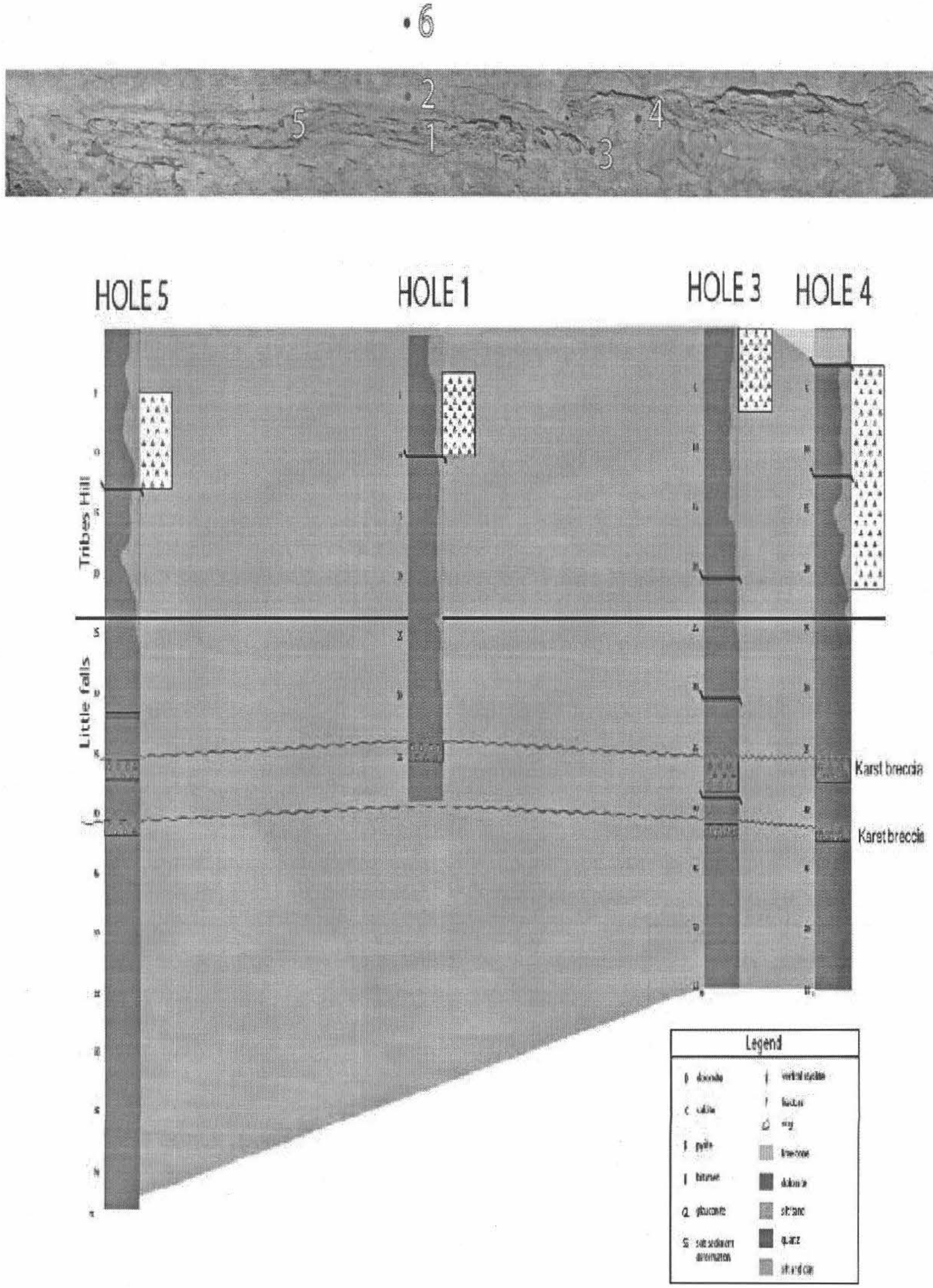


Figure 16. Cross-Section Parallel to the Bodies Across Holes 1, 3, 4, and 5

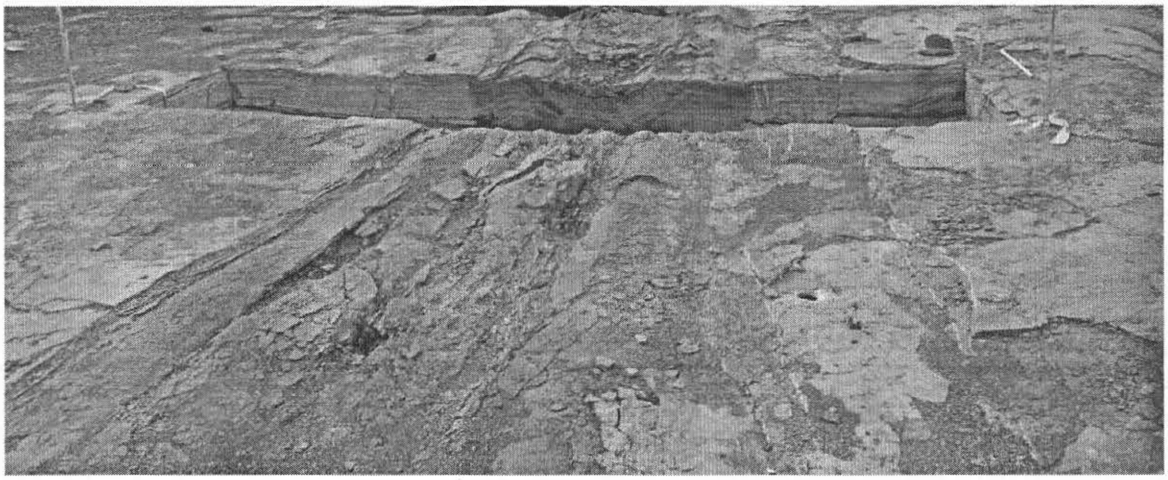


Figure 17. The Western Wall of Trench 4

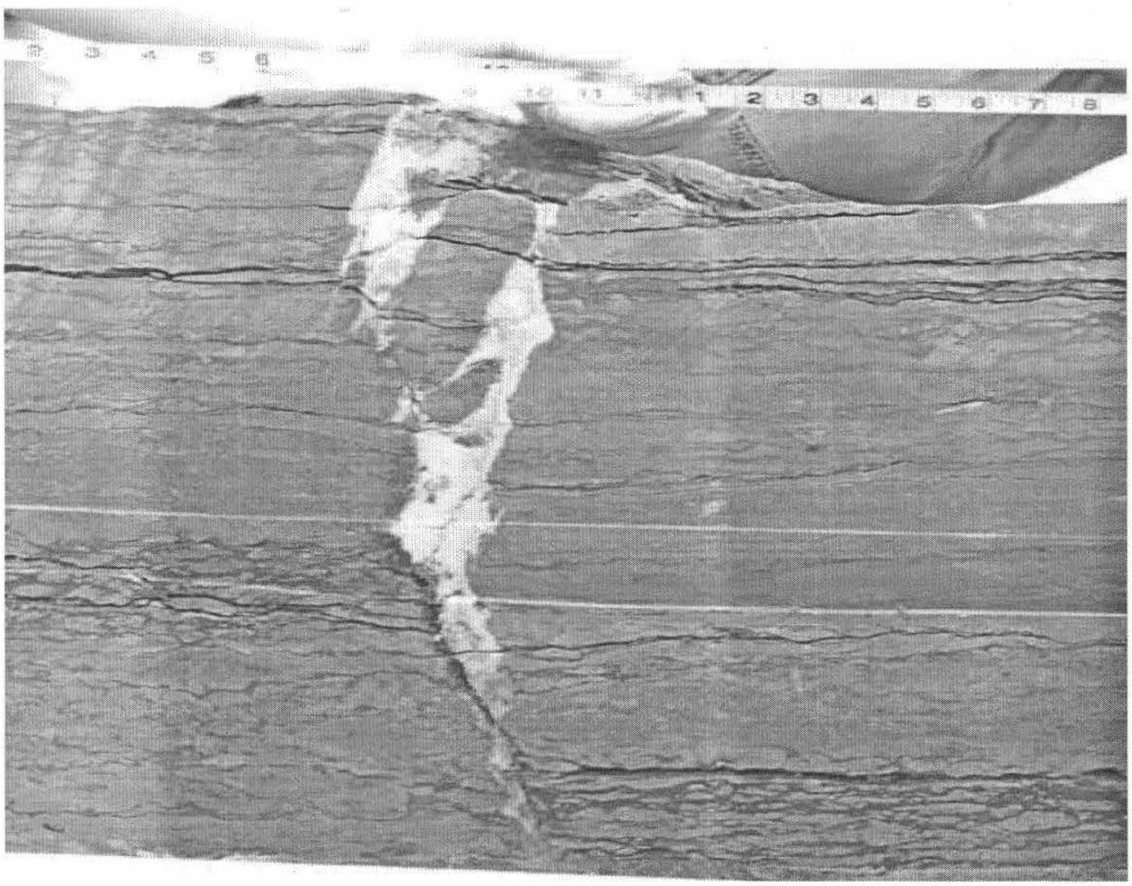


Figure 18. A Bifurcating Calcite-filled Fracture Appearing in one of the Trench Walls. Note the offset along the dark shaly layer.

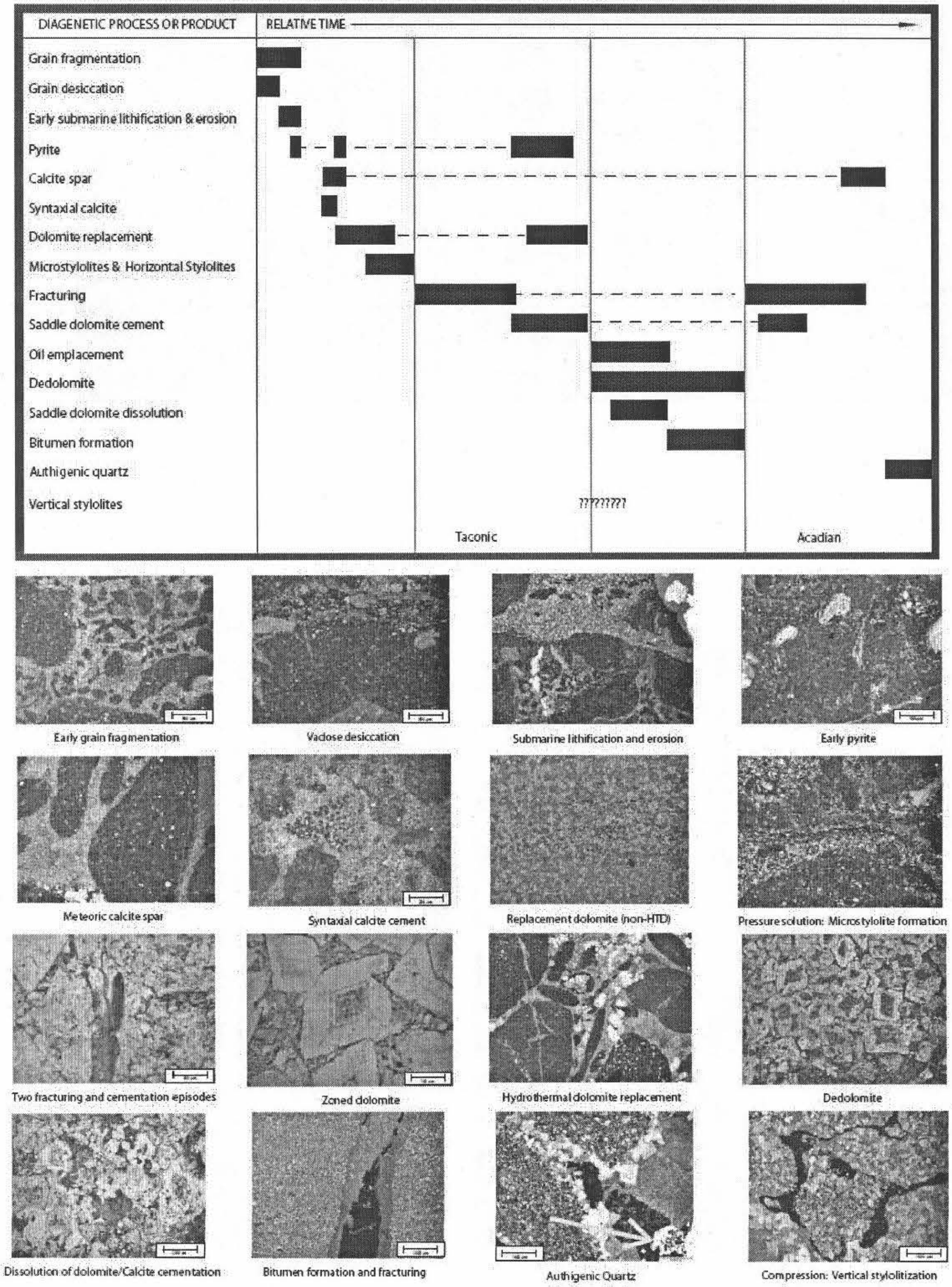


Figure 19. Paragenic Sequence and Photomicrograph Examples of each Type

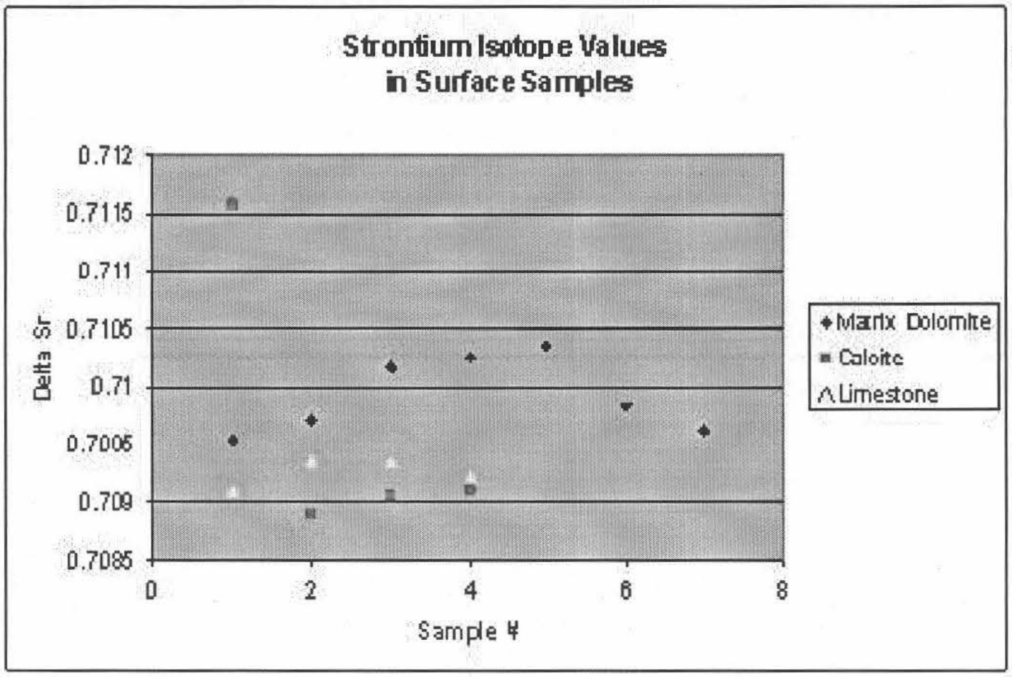


Figure 20. Graph of Strontium Isotope Values for Quarry Samples

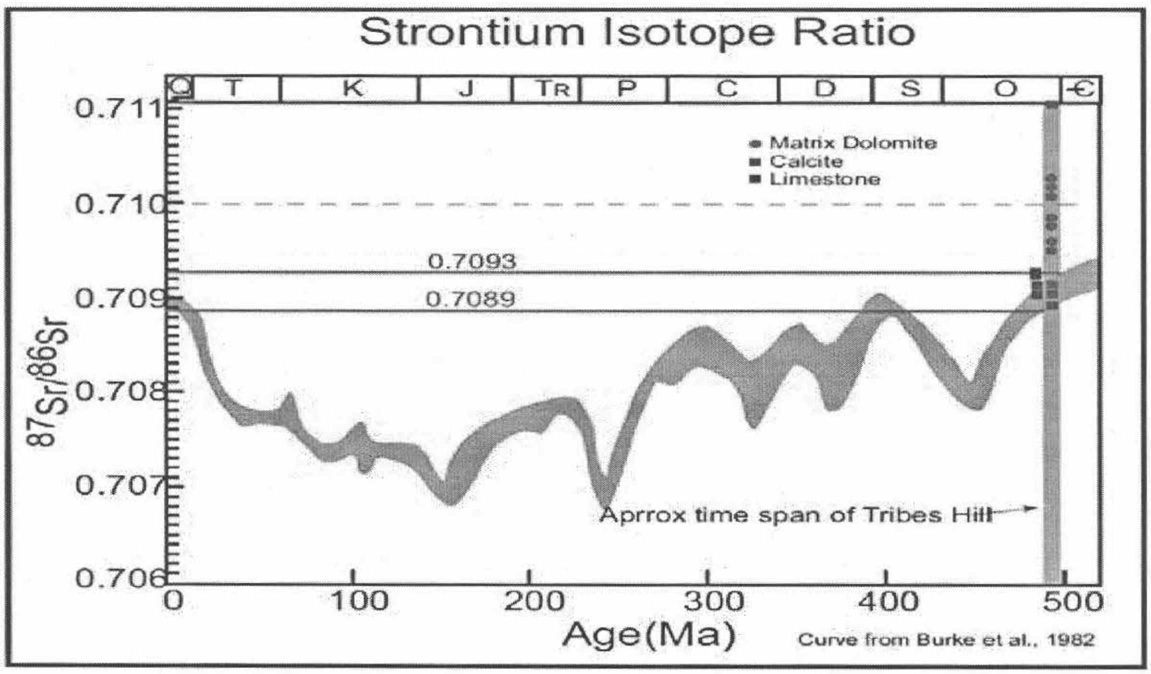


Figure 21. Strontium Isotope Content of Quarry Samples Plotted on Prehistoric Ocean Content Curve

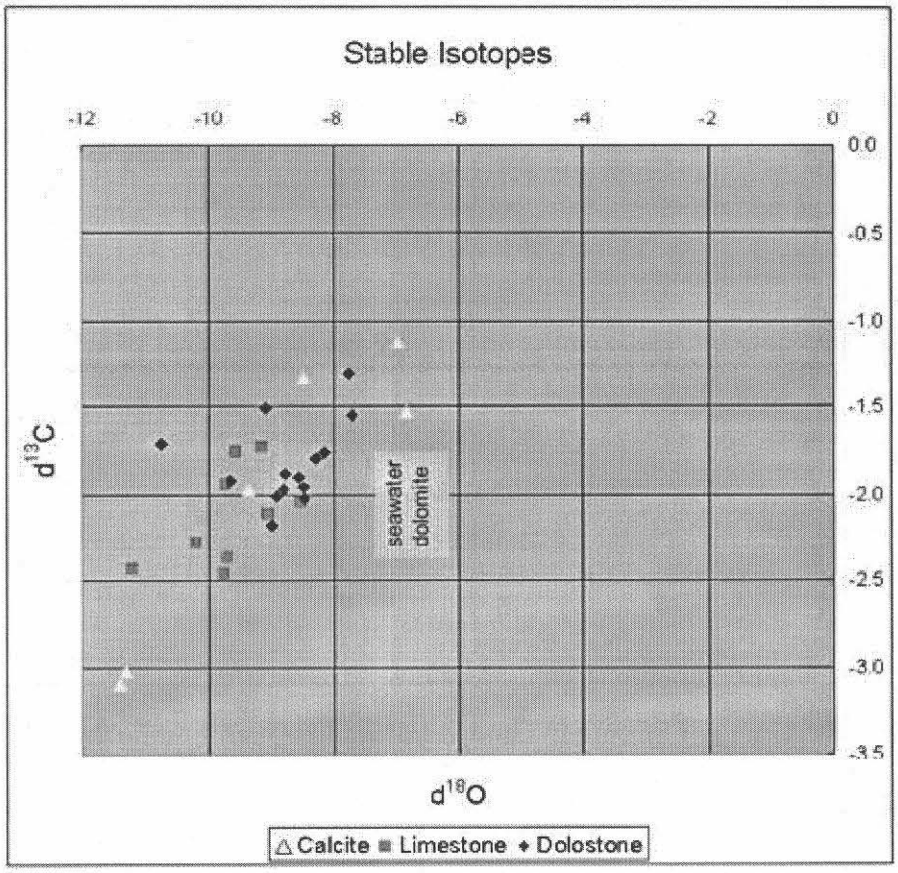


Figure 22. Plot δ^{18}O vs. δ^{13}C for Samples of Calcite, Limestone, and Dolostone from the Quarry

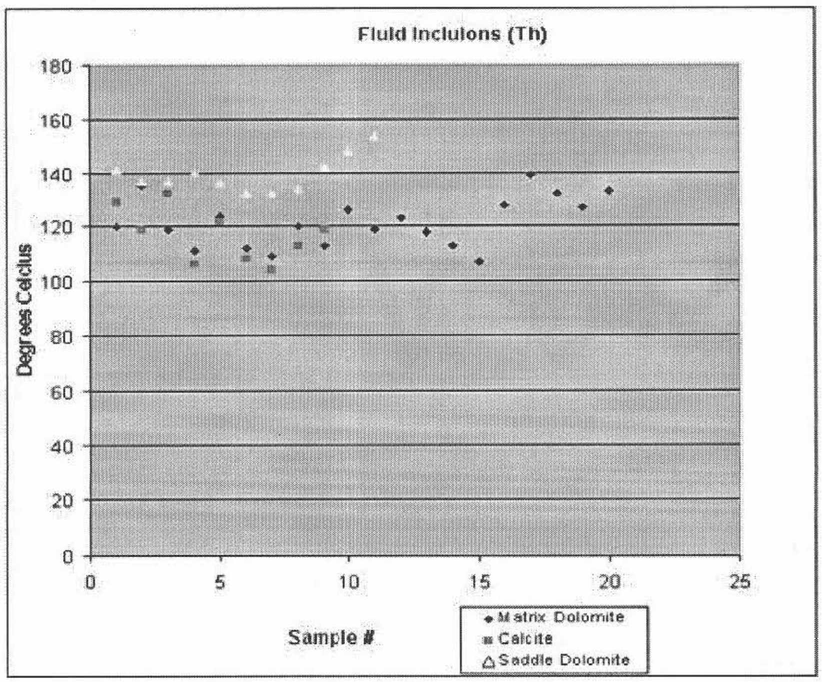
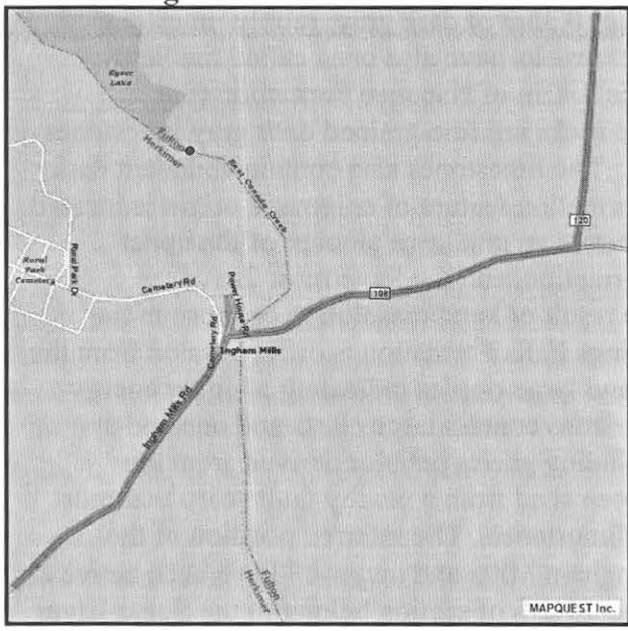


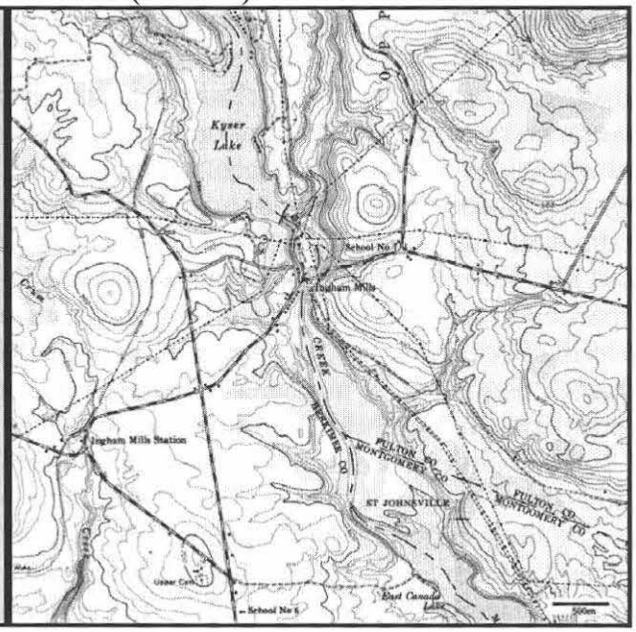
Figure 23. Homogenization Temperatures for Fluid Inclusions from Quarry Samples

Reverse Direction on unnamed quarry access road	**<0.1mi**	**1.6mi**
Left on NY-5	**11.1mi**	**12.7mi**
Right on Snell's Bush Rd. (CR-23)	**3.8mi**	**16.5mi**
Right on Ingham Mills Rd. (CR-127)	**0.8mi**	**17.3mi**
Left on Powerhouse Rd.	**<0.1mi**	**17.3mi**

Park in front of Powerhouse and walk to dam spillway

STOP 2: Ingham Mills **GPS (NAD27): 519170E 4767490N**





Ingham Mills is a well known paleontological and stratigraphic locality due to the complete Trenton/Black River section at this location (Fig. 24). The section here is over 23m thick and is the farthest eastward extent of the Watertown and Napanee Formations. The section is exposed on the spillway beneath the hydro dam that holds back the waters of Keyser Lake. The lowermost outcrop at the base of the spillway is a buff-weathering, fine-grained dolostone. This dolostone has previously been classified as the uppermost surface of the Little Falls Formation (Cameron et al., 1972; Cameron and Kamal, 1977; Mitchell and Marsh, unpublished data 1998). The hummocky nature of the surface does resemble thrombolitic horizons within the Little Falls Formation, though the fine-grained texture of the dolostone is very rare within this formation. However, new observations reveal that this is not a simple sedimentary contact. The dolomite observed at the section is actually an aureole of dolomitization hosted within the Lowville limestone along a fracture trending 290° (Fig. 25). On the north side of this fracture, the limestone layers clearly "onlap" over the dolomite body, but on the south side of the fracture, the dolostone grades rather abruptly into limestone within the same bedding plane. This gradation indicates that this dolostone was formed by migration of a dolomitizing fluid through the limestones along this fracture trend. The geometry of this dolomitization aureole and the surrounding limestones are strikingly similar to those observed at the Palatine Bridge Quarry (STOP 1).

Above the dolomite body another 8.66m of Lowville Formation is exposed. The rocks here consist of blue-gray to dove-gray, very fine-grained limestone. Many horizons are riddled with vertical burrows of *Phytopsis tubulosum* and vertical fingers of anthraxolite (filling in voids of dissolved burrows?). One layer in the upper Lowville contains large colonies of *Tetradium* both over-turned and in life position. Two horizons within the Lowville show distinct tidal channel structures. The lowermost of these horizons show a rubbly lag on the lower surface of the channel. The rocks herein assigned to the Lowville Formation have been previously designated as Gull River by Cameron and others (1972), whereas Cornell (2003) classified the lowermost 3m of this section as the Pamelia Formation and assigned the remainder to the Lowville Formation. At the top of the Lowville Formation lies 0.35m of dark gray, rubbly limestone which we assign to the Watertown Formation. These rocks have also been called the Selby Formation by Cornell (2003). Above this horizon is 3.42m of Napanee Formation (the lowermost formation of the Trenton Group). These rocks are fine-grained dark gray limestones that weather to a distinctive chocolate brown color. The limestones also contain abundant dark gray, calcareous shale interbeds. A prominent deformation feature of enigmatic origin is located on the west side of the outcrop; this feature incorporates an unknown amount of the upper Lowville, the entire thickness of the Watertown Formation, and the lowermost 2m of the Napanee Formation. This structure is probably the result of karst/dissolution collapse in the Lowville Formation. The overlying 9.3m of the Kings Falls Formation is differentiated from the Napanee Formation by its coarser-grained texture and large ripples indicating a higher energy depositional environment. A few of the rippled horizons contain large clasts and one bed at 14.2m contains clasts of some underlying units including gneiss pebbles derived from the Precambrian basement. These clasts had to have been shed from a nearby fault scarp that must have existed nearby during the deposition of these limestones. The inferred position of the Dolgeville Fault lays a mere 400m to the west of Ingham Mills and may well have been active at this time (Jacobi and Mitchell, 2002). The uppermost 1.2m of section belong to the Sugar River Formation, which is differentiated from the underlying Kings Falls by a decreased grain size, absence of high-energy ripples, and numerous bedding planes with abundant *Prasopora* colonies.

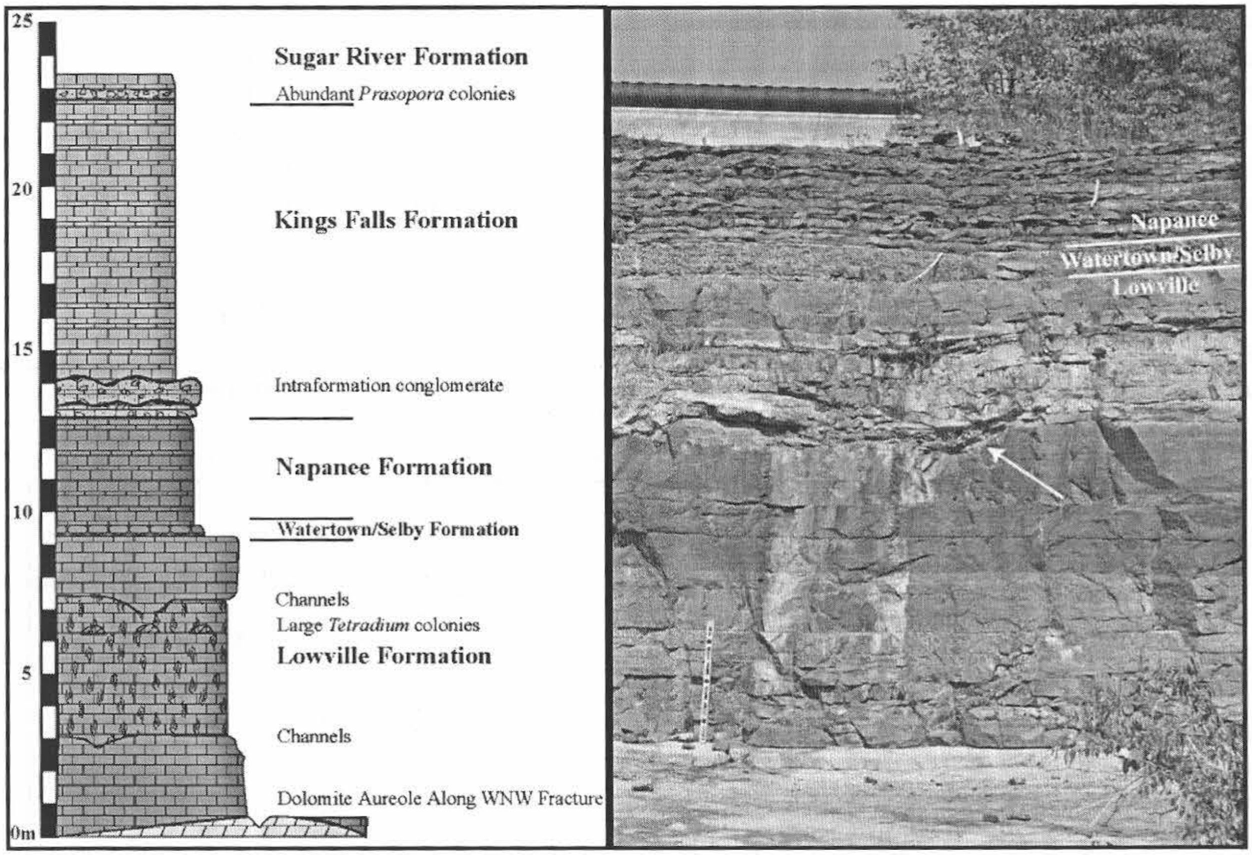


Figure 24. Stratigraphy at Ingham Mills. Stratigraphic column (left) showing 23m section present at this locality. Photo (right) showing ledgy nature of the Lowville Formation. White arrow points to a tidal channel. Note column does not correspond directly to the photo.

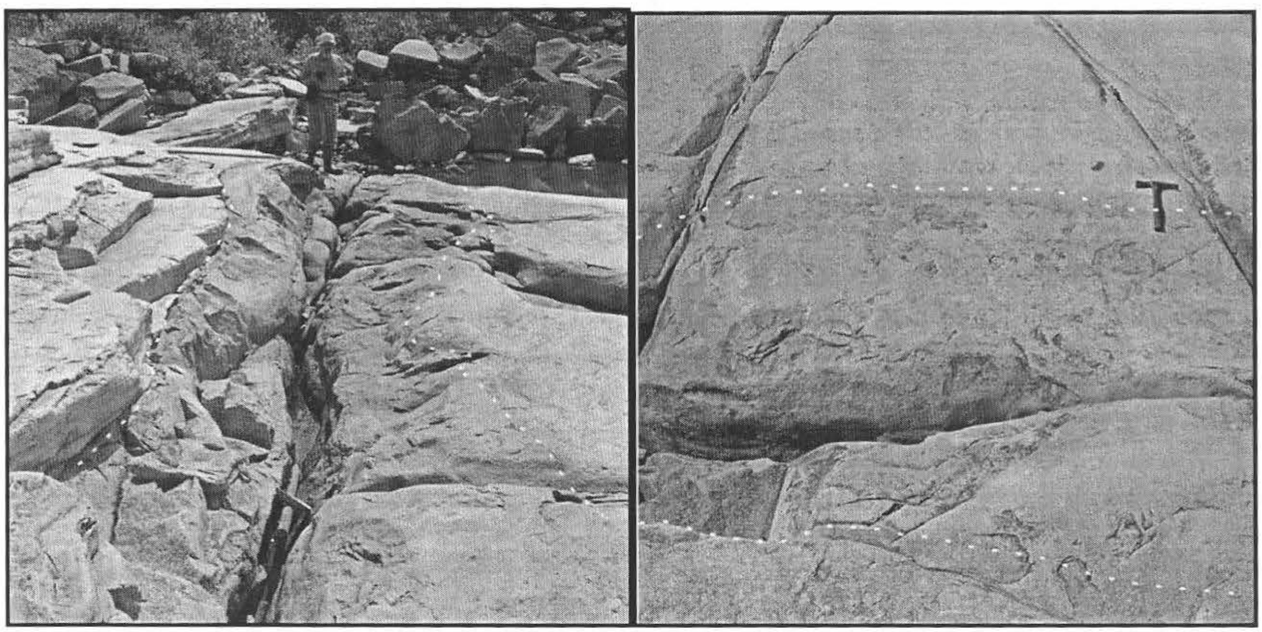
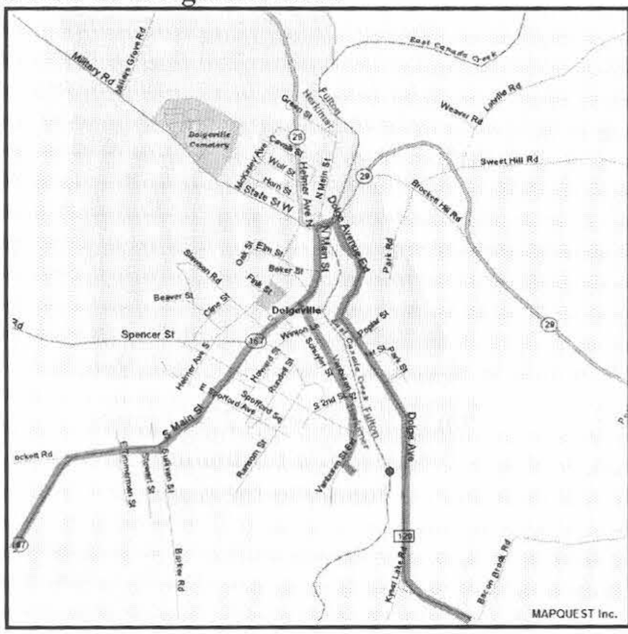


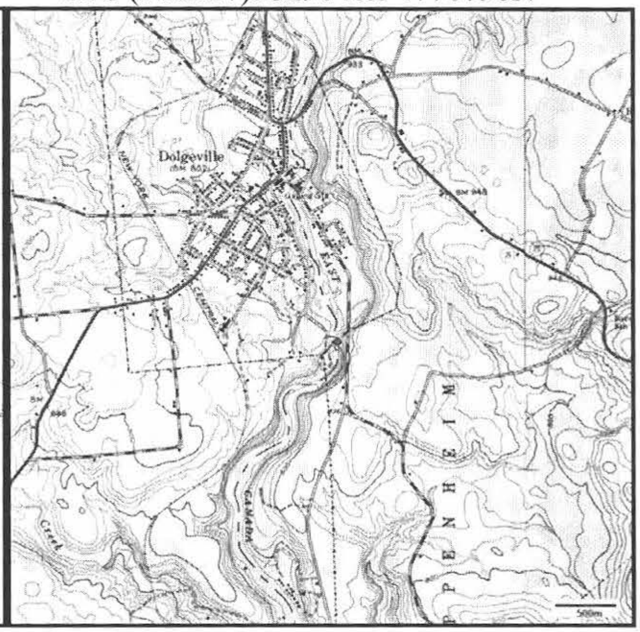
Figure 25. Photo (left) of WNW-trending Dolomitized Fracture (photo looks ESE). Note monoclinal dolomitization contact on the left side of fracture and gradational contact on the right. Photo (right) shows gradational nature of the aureole. Dotted lines show contact between limestone and dolostone.

Reverse Direction on Powerhouse Rd.	**<0.1mi**	**17.3mi**
Left on Ingham Mills Rd. (CR-108)	**0.6mi**	**17.9mi**
Left on CR-120	**2.6mi**	**20.5mi**
Name Change CR-120 to Dolge Ave	**0.6mi**	**21.1mi**
Right on Dolge Ave Ext.	**0.3mi**	**21.4mi**
Left on State St. East (NY-29)	**<0.1mi**	**21.4mi**
Left on N. Main St. (NY-167)	**0.3mi**	**21.7mi**
Left on VanBuren St.	**0.6mi**	**22.3mi**
Left on Power House Rd.	**<0.1mi**	**22.3mi**

Park at Dolgeville Hydroelectric Power Project and walk down to East Canada Creek

STOP 3: Dolgeville Dam **GPS (NAD27): 519040E 4770790N**





The power dam at Dolgeville on East Canada Creek is the most accessible exposure of the Dolgeville Fault (Fig. 26). The fault plane itself is not visible at this outcrop, though the effects of faulting are clearly and dramatically observed in the downthrown block. The fault disappears beneath thick glacial cover to the north and cannot be traced reliably for any great distance. To the south, the actual fault plane outcrops along the east bank of Kyser Lake (now accessible only by boat) and consists of a well-cemented fault breccia with iron/sulfide mineralization (Fig. 27). Cushing (1905) drew an excellent cross-section of this southern outcrop along East Canada Creek as it was before the dam at Ingham Mills raised the river level to create Kyser Lake (Fig. 28). The fault does not outcrop south of this location, but stratigraphic evidence reveals that it probably crosses the southern part of the lake and continues southward (with diminishing throw), where it passes somewhere between West Crum and East Canada Creek. If the fault crosses the Mohawk River, then the throw is severely reduced. At the Dolgeville Dam, the fault trends roughly N-S and drops down to the west (antithetic to the other major faults). Cushing (1905) estimated the throw of the fault at >300ft (91.4m), whereas Fisher (1954) estimated 450ft (137.2m). The Dolgeville Fault here exhibits dip slip kinematic indicators. Minor drag folds have nearly horizontal axes, and down-dip slickensides (Fig. 26 inset) occur on minor faults and on bedding around some of the folds (Jacobi and Mitchell, 2002). Farther south Bradley and Kidd (1991) also found downdip slickensides on a minor fault near the Dolgeville Fault (their Figure 8a). Thus, if these kinematic indicators are related to Ordovician motion, then the Dolgeville Fault was primarily a dip slip (normal) fault. We believe this fault originated in response to plate stress as it descended into the trench. The fact that this fault cross-cuts gravity anomalies and that the N-trending fractures (Dolgeville trend) usually post-date the NNE-trending fractures (Little Falls trend) indicate that slip on this fault likely initiated relatively late as a result of peripheral bulge tectonics and is younger than the Little Falls Fault. There are a few folds and faults that indicate a later reversal in motion on the Dolgeville fault.

At this site, the upthrown block consists of sub-horizontal Little Falls dolostone that outcrops along Dolge Ave. overlooking the falls. Looking northeast at the spillway outcrop, the downthrown block consists entirely of Indian Castle black shale with rare limestone beds. The beds of the falls are sub-horizontal near the powerhouse, but slowly begin to dip to the west nearing the fault. At the small line of trees, the bedding dip abruptly increases across a minor fault. The beds then continue to gradually increase in dip to angles measured as high as 59°W, although the beds are much steeper on the inaccessible shear cliff-face. Within the steeply dipping beds, the Dolgeville/Indian Castle contact is gradational, with an upsection increase in the thickness of the intervening shales. No time gap is thought to exist at this locality between the Dolgeville and the Indian Castle (Jacobi and Mitchell, 2002). The lack of a significant, observable "Thruway disconformity" (viewed at STOP 5) is consistent with the lack of a significant number of the distinct folds that characterize the NYS Thruway outcrop. No unequivocal slump folds, no time gap, and no locally angular unconformity all suggest that the "Thruway disconformity" is not a regional unconformity rather it is a slide scar, and the sediment slide did not affect this section (Jacobi and Mitchell, 2002). That scenario is consistent with the tectonic position of this outcrop—it is in the base of the Dolgeville graben, whereas the NYS Thruway outcrop is located on the horsts of the Little Falls fault system.

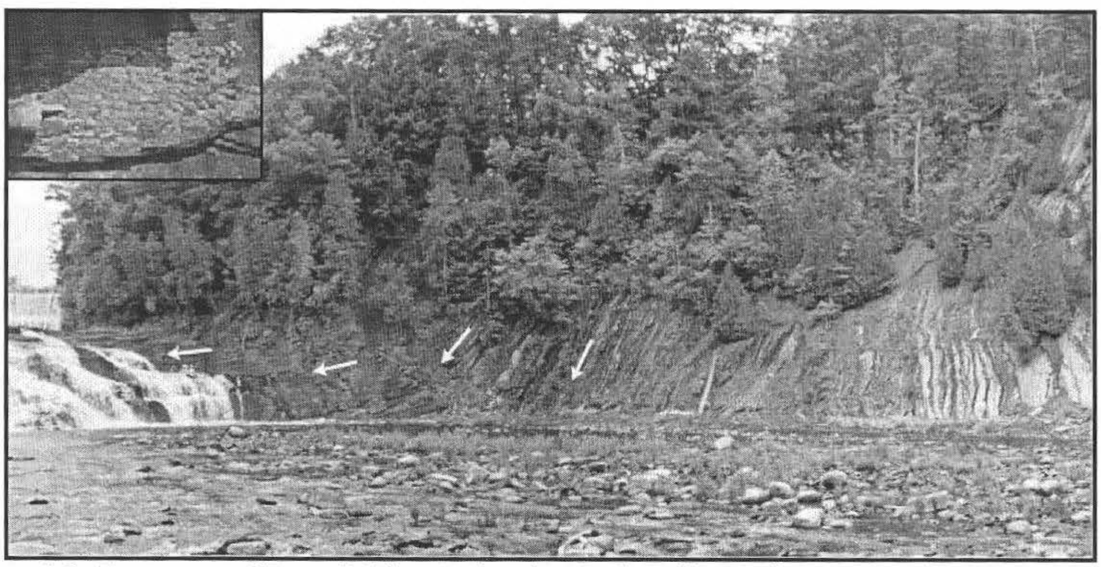


Figure 26. Exposure of Drag-folding Related to Dolgeville Fault at Dolgeville Dam. White arrows highlight changes in bedding dip associated with folding. Inset shows down-dip slickenlines observed here. The Indian Castle Member shales grade into Dolgeville Formation shales and limestones from left to right and the units are drag folded against the fault (to the left of picture). The Little Falls Formation is exposed on the upthrown block.

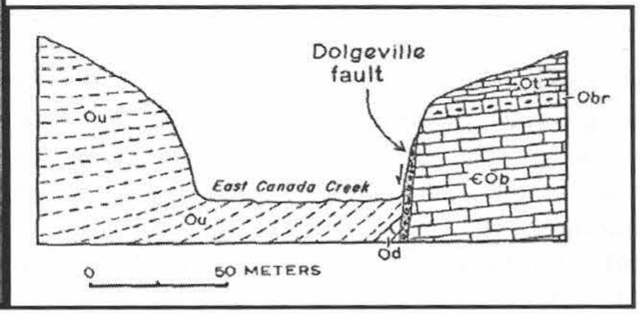


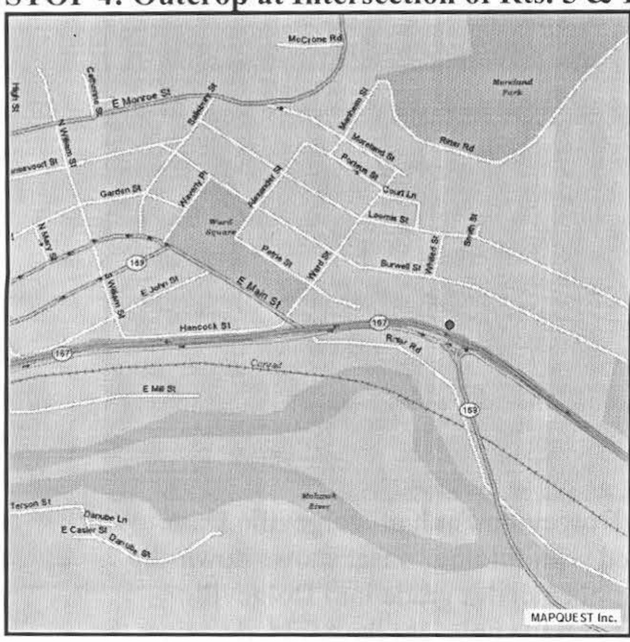
Figure 27. Fault Breccia Observed Along East Bank of Kyser Lake (south of the Dolgeville dam).

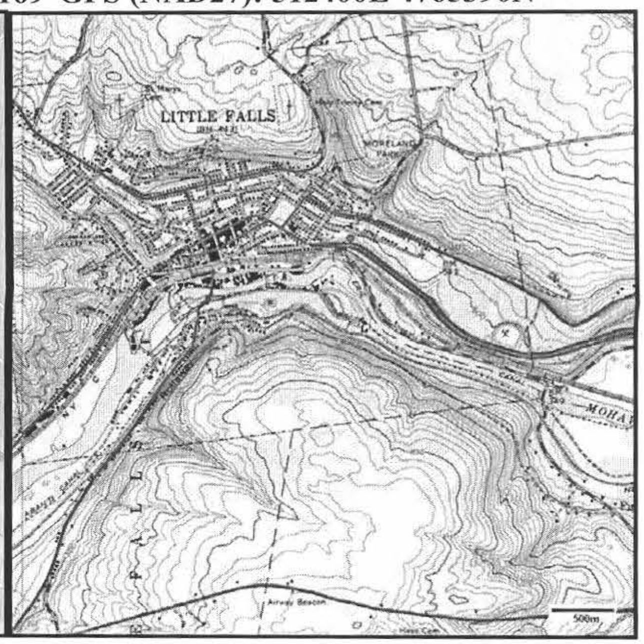
Figure 28. Schematic Cross-Section of Dolgeville Fault at Southern Exposure. Stipple pattern is fault breccia seen in Figure 27. From Cushing (1905).

Reverse Direction on Power House Rd.	**<0.1mi**	**22.3mi**
Right on VanBuren St.	**0.6mi**	**22.9mi**
Left on Main St. (NY-167)	**6.8mi**	**29.7mi**
Right on NY-5/NY-167	**0.7mi**	**30.4mi**

Park along wide shoulder near base of incline across from intersection with NY-169

STOP 4: Outcrop at Intersection of Rts. 5 & 169 GPS (NAD27): 512400E 4765390N





Cushing (1905) described several outcrops of the Little Falls Fault in the field area. Today, these outcrops no longer exist or are severely degraded. The only locality that exposes the fault plane is the Thruway cut south of the Mohawk River (STOP 5). To the north of this locality, a strong NNE-trending topographic lineament is easily recognized. In the creek to the east of Moreland Park, a NNE-trending fracture intensification domain (FID) occurs within the Little Falls Formation along with loose blocks of stock-work breccia. The fractures within this FID have an average strike of 038° with a frequency of 20 fractures/m. This lineament appears to offset both the Precambrian-Little Falls contact and the Little Falls-Tribes Hill contact with the western side down-dropped. To the south, this topographic lineament truncates Moss Island on the west side. Across the Mohawk River, this lineament is clearly visible as a small cleft in the "Rollaway". The Precambrian-Little Falls contact appears to be offset down to the east across this cleft, though no other evidence for faulting was observed at that locality. The fact that the westerly block is down-dropped on the north side of the Mohawk River, whereas the easterly block is down-dropped on the south side indicates that this splay is either a scissors fault or there is an unrecognized cross fault between the two localities. At the present location, this lineament is expressed as the small creek that we see at the stop. At this site, we observe another NNE-trending (average strike = 040°) FID (fracture frequency = 13.5 fractures/m) and also a fairly dramatic change in lithology in the Precambrian basement. The east side of the creek is well-foliated charnockitic gneiss, whereas the west side is darker, much more mafic, poorly-foliated gneiss. Slickenlines can be observed on both NNE and WNW surfaces at this outcrop (Fig. 29). Fractures of the NNE-trend increase in frequency nearing the fault (Fig. 30). The preferred growth orientations of crystals within the slickenlines indicate that the NNE trending fault experienced right lateral oblique slip with the SE fault block down-dropping (Fig. 29A). This sense of motion is opposite to the stratigraphic offset along this splay of the Little Falls Fault on the north side of the Mohawk River, thus the offset on the Little Falls Fault may be the result of several cumulative slip events with fault motion reversing at intervals throughout its history.

The slickenlines on the WNW surface are sub-horizontal (plunge = 04°) and indicate left-lateral slip with the northern block being slightly down-dropped (Fig. 29B). Leaving the creek and looking at the road cut exposures along Rt. 5, another WNW-trending surface shows well-developed slickenlines. The slickenlines at this location record left-lateral oblique slip with a slight down on the south component (Fig. 29C). This sense of offset does not agree with the offset along the inferred WNW transfer fault to the north, but it is consistent with the sense of motion required to drop the Precambrian-Little Falls contact down on the south side of the Mohawk River (on an inferred WNW fault). On the north side of the river, this contact lies at 560ft (170.7m), whereas on the south it is at 400ft (121.9m). This offset can be partially explained by regional dip on this contact, which is estimated at 03° to the south based on the topographic gradient observed on the north side of the Mohawk River. Projecting this dip across the Mohawk River would account for only 125ft (38.1m) of offset. The kinematics of all the slickenlines (Fig. 29D) at this locality are consistent with the expected motions in an eastward-directed far-field compressional regime, though these faults probably experienced several stages of slip and reversals in sense of motion.

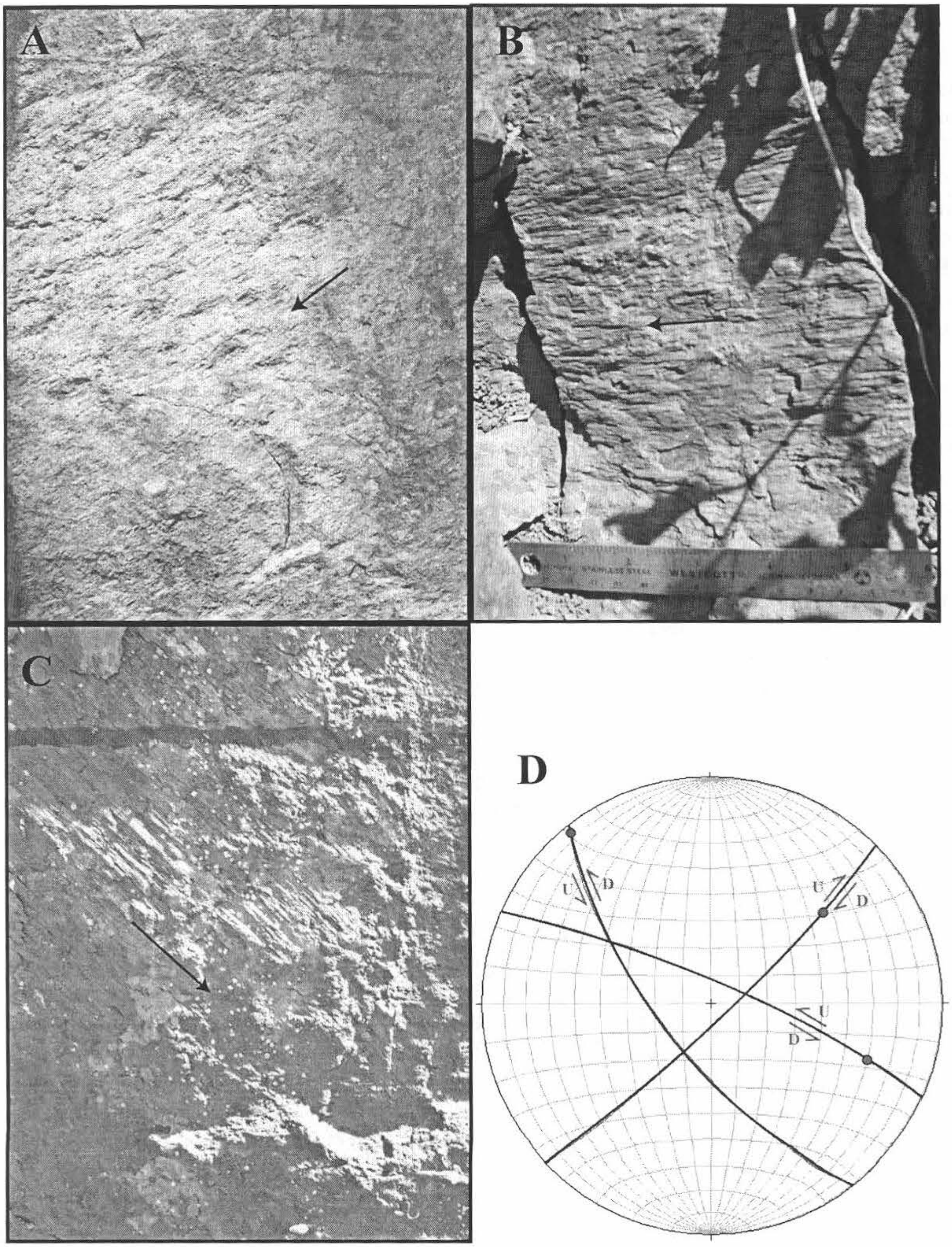


Figure 29. Slickenline Data from STOP 4. Arrows show likely relative slip of the block that the photographer is standing on (slip determined from crystal growth directions). **(A)** Slickenlines on NNE-trending fracture (photo taken looking WNW) **(B)** Slickenlines on WNW-trending fracture (photo taken looking NNE). **(C)** Slickenlines on WNW-trending fracture (photo taken looking NNE) **(D)** Stereoplot of slip planes and crystal lineations. Arrows show inferred direction of slip and U and D indicate up and downthrown side

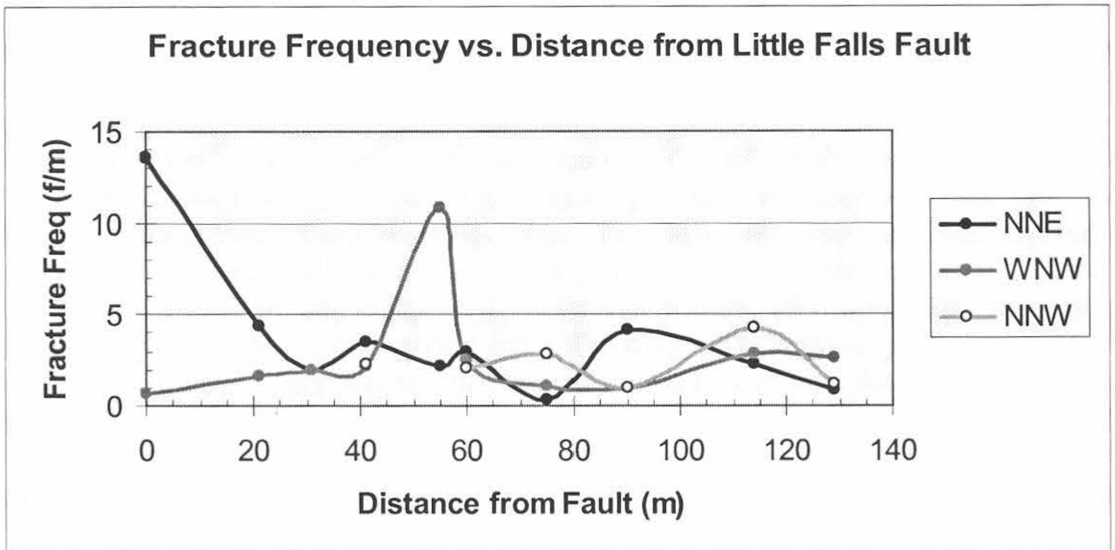


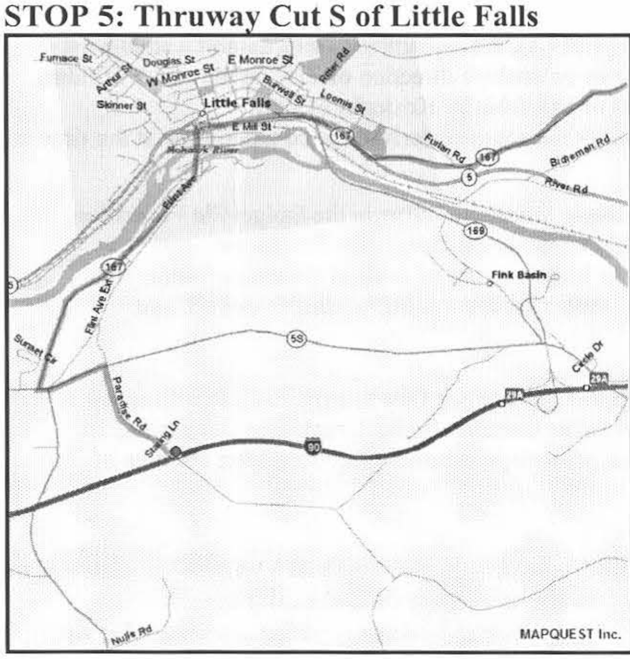
Figure 30. Plot of Fracture Frequency vs. Distance from Little Falls Fault "B" at STOP 4.

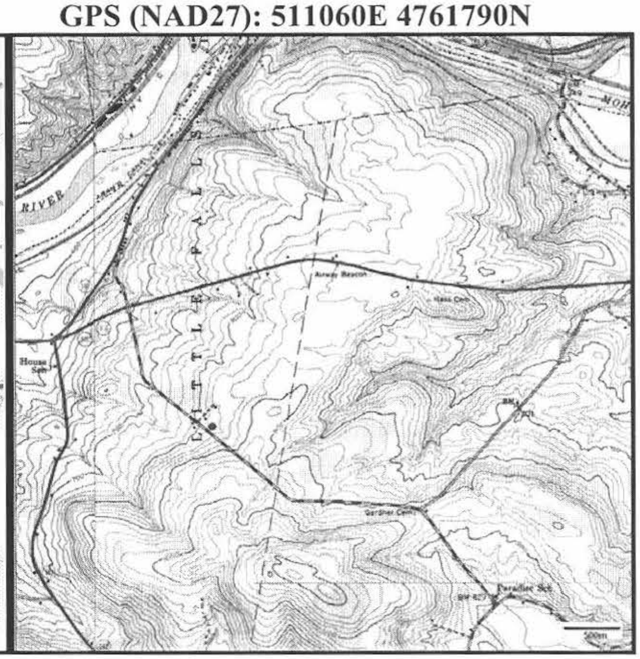
Right on NY-5/NY-167	**0.6mi**	**31.0mi**
Right on Albany St. Ramp	**0.1mi**	**31.1mi**
Right on NY-167S	**2.1mi**	**33.2mi**
Left on NY-5S	**0.5mi**	**33.7mi**
Right on Paradise Rd.	**0.8mi**	**34.5mi**

END ROAD LOG (Total Mileage: 34.5mi)

Park in construction zone at northwest corner of Paradise Rd. and Thruway intersection

STOP 5: Thruway Cut S of Little Falls **GPS (NAD27): 511060E 4761790N**





The NYS Thruway roadcut south of Little Falls provides a 1.2 mile long look at the ribbon limestones of the Dolgeville Formation, the overlying black shale of the Indian Castle Member, and the unique contact between these two units (Fig. 31). At the Thruway roadcut, the upper few meters of the Dolgeville Formation are intensely folded. Fisher (1979) first studied these folds in 1953 when the roadcut was dug. Fisher (1979) noted the dominant westward vergence of the folds, and suggested that westward slumping away from the foredeep (not toward the basin) had caused these folds. He proposed that westward gravity sliding of the Taconic Giddings Brook "Slice" into the foredeep (Magog, or Snake Hill, Basin), which was located east of the thruway roadcut along an axis between the Hudson River/Lake Champlain and the Green Mountains/Taconics) had depressed the crust below the gravity slides and caused the Adirondack Arch to rise. Since the arch was located east of the roadcut, the rising arch would have reversed the original eastward slope to a westward slope, and thus the slumps traveled west, not east, off the carbonate ramp. The overlying Indian Castle Member lies largely undisturbed above the local unconformity known as the Thruway Discontinuity (after Baird and others, 1992).

Fisher's (1979) fold vergence was supported in general by studies by two of Jacobi's students who analyzed the outcrop in the 1980s (Ritter, 1983; Koslosky, 1986). Jacobi and Mitchell (2002) further analyzed the folds and paleoflow indicators at this site. Most of the folds are disposed in "trains" and affect only a relatively few beds; i.e., they are intrastratal (or intraformational), disharmonic folds. As described in Jacobi and Mitchell (2002):

> Only kink band folds (box folds) affect more than a few beds. At the unconformity, eroded crestlines and troughs filled with breccia from the ribbon limestones, demonstrate that at least here the folds were exposed at the seafloor to erosive processes. These folds are not the result, therefore, of relatively deeply buried bedding-parallel slip. Both Fisher (1979) and Koslosky (1986) suggested that the folds were "slump folds" related to downslope slump processes, based on fold form and character. Highly contorted crestlines of some folds indicate that these folds underwent variations in strain along their axes as the sediments were transported downslope during the sliding event(s).
>
> At this stop (near milepost 213), the asymmetric folds have fairly straight crestlines and west vergence. Application of the arc separation method (Fig. 32) yielded a down-paleoslope direction of 234.5° at the time of folding (Fig. 33). At a western outcrop (at milepost 214.8), the asymmetric folds exhibit a southwest vergence and the arc separation method yielded a down-paleoslope direction of 229° at the time of folding. These values are in accord with the detailed analyses of 698 folds by Koslosky (1986).
> Other features (Fig. 34) in the eastern outcrop that suggest a southwesterly-directed paleoslope at the time of sliding include the following observations:
>
> 1. small west-directed thrusts that affect single limestone layers in the Dolgeville Formation,
> 2. westward-directed intrastratal micro-fold nappe in the Dolgeville Formation,
> 3. imbricate, east-dipping clasts in breccias in the troughs of folds at the unconformity that suggest a westerly current flow immediately after the folding (sliding event[s]), and
> 4. a small slide scar that faces west.
>
> Paleoflow data from the Dolgeville Formation at this outcrop imply a flow toward 207°; this direction is only 27° away from the paleoslope determined from the asymmetric folds. Graptolites and parting lineations above the unconformity show that the local paleoslope reversed directions after the time of slumping.

The major splay of the NNE-striking Little Falls Fault ("C") cuts the units in this stop (Fig. 35). The fact that bedding dip in the downthrown Utica (on the east side of the fault) is not perpendicular to the fault plane is evidence that this fault probably experienced oblique slip at this location (Jacobi and Mitchell, 2002). The newly recognized oblique-slip slickenlines along faults in the Precambrian basement at Little Falls (STOP 4 along Little Falls Fault "B") are consistent with the sense of rotation on the block inferred from the anomalous dip. In the past, no kinematic indicators were observed in the fault surface at the Thruway, but with the recent NYS Thruway slope abatement project, we may find kinematic indicators. Minor splays of the Little Falls Fault occupy valleys to the east and are recognized by minor changes in dip across the valleys. It is our contention that motion on these faults reversed the local paleoslope, causing the slumping activity recorded as the slump folds and the Thruway disconformity (a slide scar). This motion was related to the plate descending into the trench. (It should be noted that others maintain that the paleoslope did not reverse; Brett and Baird, 2002).

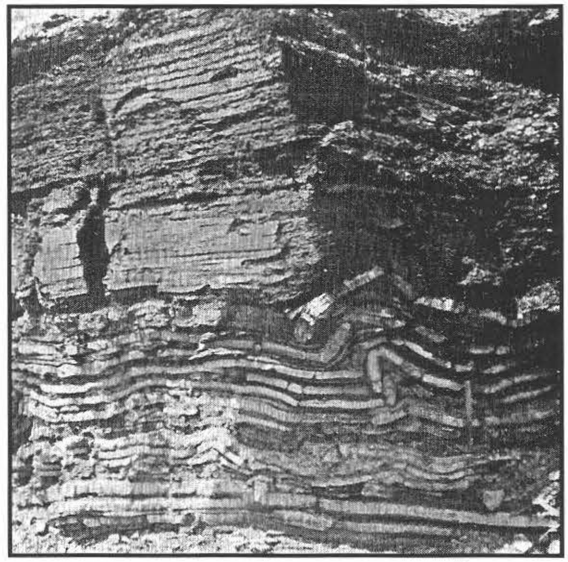


Figure 31. Photo of the Thruway Discontinuity. Undisturbed Dolgeville Formation at bottom of photo, overlain by folded layers of this formation, followed by an abrupt transition to undisturbed black shales of the Indian Castle Member. From Fisher (1980).

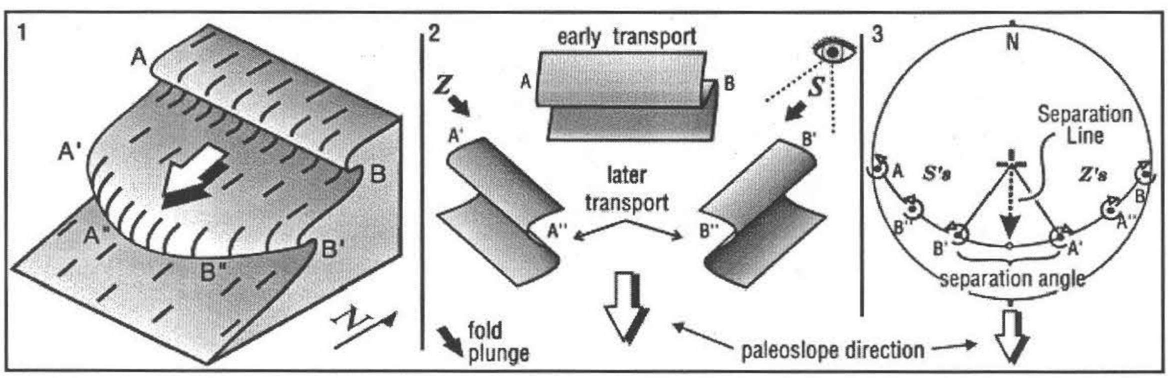


Figure 32. Rationale for the Arc-separation Method for Determining Paleoslope from Asymmetric Slump Fold Axes. Panel 1: slump folds display an evolution from straight crestlines to curvilinear crestlines as they are translated downslope. Panel 2: schematic diagram showing the evolution from straight crestlines to *Z* and *S* asymmetric folds (the asymmetry of folds, *Z* and *S*, is defined looking downplunge, opposite to the upslope view in the panel). Panel 3: stereonet plot of the axes shown in Panel 1; paleoslope is directed between the cluster of *Z* and the cluster of *S* asymmetric fold axes. From Jacobi and Mitchell (2002).

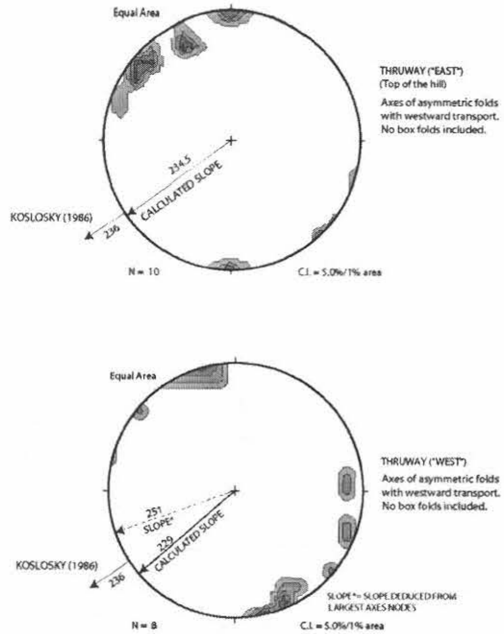


Figure 33. Stereoplots Using the Arc-Separation Method. Folds exposed at STOP 5 (Thruway East stereonet, top) yield a paleoslope of 234.5°. Folds exposed west of STOP 5 (Thruway West stereonet, bottom) yield a paleoslope of 251°. From Jacobi and Mitchell (2002).

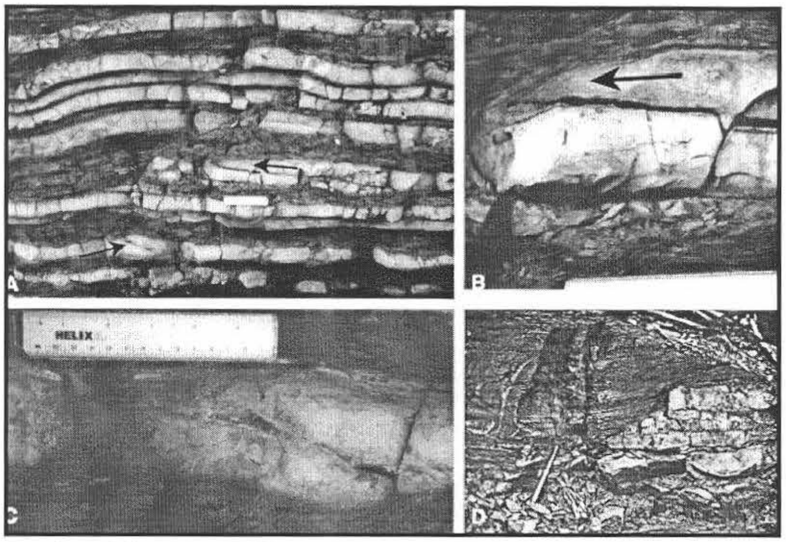


Figure 34. **(A)** Dolgeville Formation at Thruway East Below the Paradise Road Overpass. A recumbent, isoclinal fold is located immediately above the ruler at the left-directed arrow. A forced fold above the recumbent fold developed as a consequence of the asymmetric recumbent fold and associated thrust. Thrust in Fig. 34C is located below the ruler at the right-pointing arrow. Photo looks north. **(B)** Small intrastratal westward-directed, asymmetric, recumbent fold at Thruway East. Note the isoclinal folding of layering in the fold nose left of the arrow. **(C)** Small intrastratal thrust at Thruway East under the Paradise Road overpass. Photo looks north. Ramping thrust only affects one ribbon limestone, and implies west over east transport (i.e., the slump trajectory was to the west). Thrust can be seen in Fig. 34A below ruler. **(D)** Mini-slide scar at Thruway East. The abrupt termination of the top two ribbon limestones is consistent with a mini-slide scar that faces west. Photo looks north. From Jacobi and Mitchell (2002).

Figure 35. Little Falls Fault "C" at the Thruway Cut (photo looks NNE). Offset is down to the east, juxtaposing Dolgeville Formation and lowermost Indian Castle Member (west block) with Indian Castle Member. A minor splay can be seen on the right with an unknown offset within the black shales.

OPTIONAL FIELD TRIP STOPS

Manheim Power Station **GPS (NAD27): 520880E 4763220N**

At the Manheim Power Station, the Manheim Fault is exposed along both sides of East Canada Creek (Fig. 36). Access from the Power Station driveway is on the west bank and permission is needed as this is private property. From this side of the creek, the rocks are easily accessed even at high water. The rocks along this bank superficially resemble the interbedded ribbon limestones and shales of the Dolgeville Formation and have been designated as such by previous workers (Megathlin, 1938; Fisher, 1954; Bradley and Kidd, 1991). However, faunal evidence indicates that the rocks observed here are the "Allen Road beds" of the Flat Creek Member (Mitchell, 2006 personal communication). Numerous fractures and veins dissect the outcrop at this locality. The fractures are of NE, WNW, and NNW orientations. It was expected that the NE-trending (fault-parallel) fractures would increase in frequency nearing the fault plane, but actually the NNW (fault-perpendicular) fractures show the most marked increase (Cross, 2004;

Cross et al., 2004; Fig. 37). At the northern extent of the outcrop on this side of the creek is a fault breccia. This breccia contains lithologies similar to those of the adjacent Flat Creek Member consisting of large, rounded clasts of limestone that are well-cemented in a dark, calcareous mud matrix. Across the creek on the eastern side, one can see beds of the Flat Creek Member drag-folded upward on the eastern block to angles approaching 60°SE juxtaposed against gently dipping (07°SE) beds of the Little Falls Formation on the western block (Bradley and Kidd, 1991). Here the fault strikes roughly 50° (it is drawn as roughly N-trending in Fisher, 1980) and drops down to the east (Cross, 2004). According to early workers (Conrad, 1837; Vanuxem, 1838) there is a one inch thick calcite vein with associations of galena and pyrite present, which were mined for a brief period in the area of the fault. In addition, the fault zone hosts a peridotite dike. Similar dikes in the area roughly parallel the fault trend and have been dated using Rb-Sr as 130±10 million years old (Fisher, 1980). Estimates of the offset on this fault vary greatly with Darton (1895) claiming 152ft (46.3m), Megathlin (1938) 60ft (18.3m), Fisher (1954) 400 ft (121.9m), and Bradley and Kidd (1991) 131ft (39.9m).

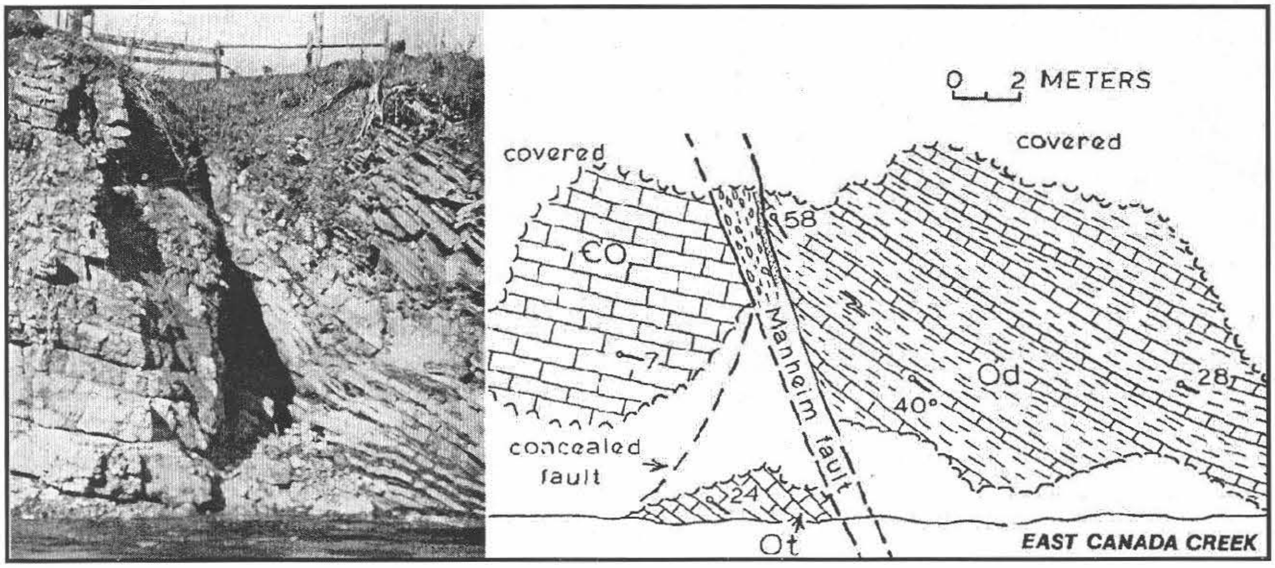


Figure 36. Manheim Fault at Manheim Power Station. Photo (left) looking northeast at the outcrop of the fault surface. From Megathlin (1938). Sketch (right) of outcrop showing bedding dips and fault breccia. From Bradley and Kidd (1991). This fault juxtaposes drag-folded Flat Creek Member (called Dolgeville Formation by previous workers at this locality) against Little Falls Formation with the east side down-dropped.

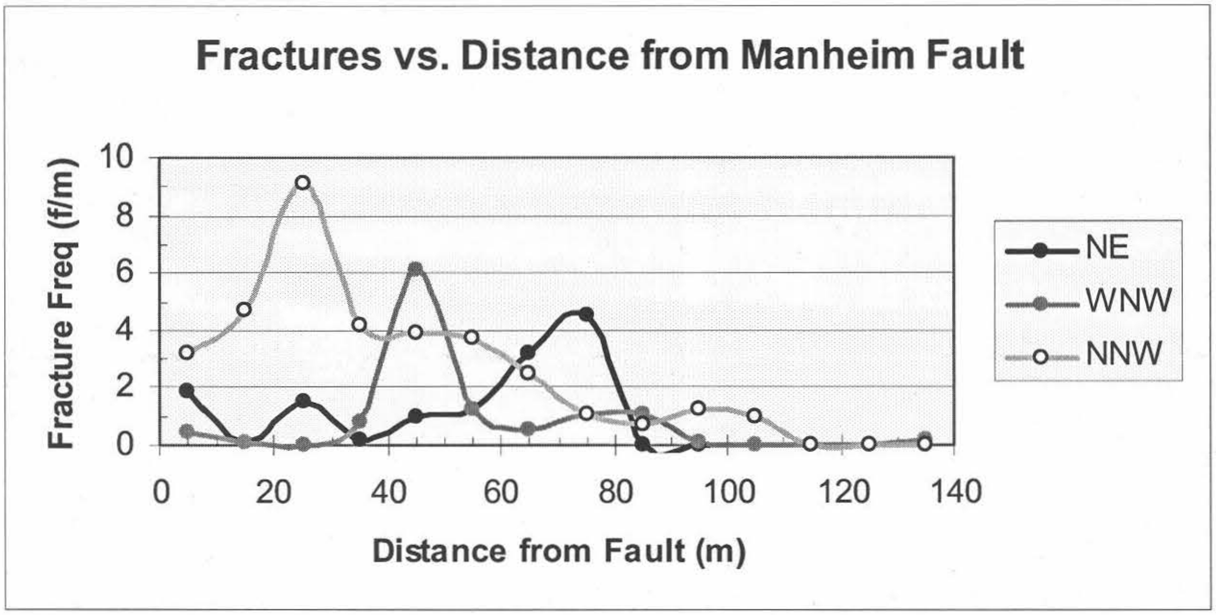


Figure 37. Fracture Frequency vs. Distance from Manheim Fault. From Cross (2004).

Miner Rd. Outcrop **GPS (NAD27): 514910E 4771185N**

In the course of widening Miner Rd., an excellent outcrop of the lowermost Tribes Hill Formation was exposed. At this locality, the rocks are a light gray, medium-bedded (11cm average), coarse-grained dolostone. The bedding at this locality trends 002/03W. Several beds show excellent herring-bone cross-stratification typical of a tidal environment. Four distinct fracture sets are present at this outcrop (N, NE, WNW, and NW). At this outcrop, the NE-trending fracture set (average strike = 53°) is master (oldest) and parallel the inferred orientation of the Little Falls Fault just to the east. Reddish staining along the NW set is interpreted as sulfide and/or iron-rich vein mineralization (Fig. 38). The outcrop also hosts many vugs within the dolostone matrix. These vugs are variably filled with white calcite (Fig. 39) and occasional buff-colored dolomite. The presence of the veining and vug development are indicative that this horizon was favorable to the migration of fluids from the nearby fault. Heading south along Miner Rd. from this outcrop, there are excellent exposures of the upper Little Falls Formation in the woods on the western side of the road. These exposures contain stromatolite colonies and also host vugs partially filled with white calcite.

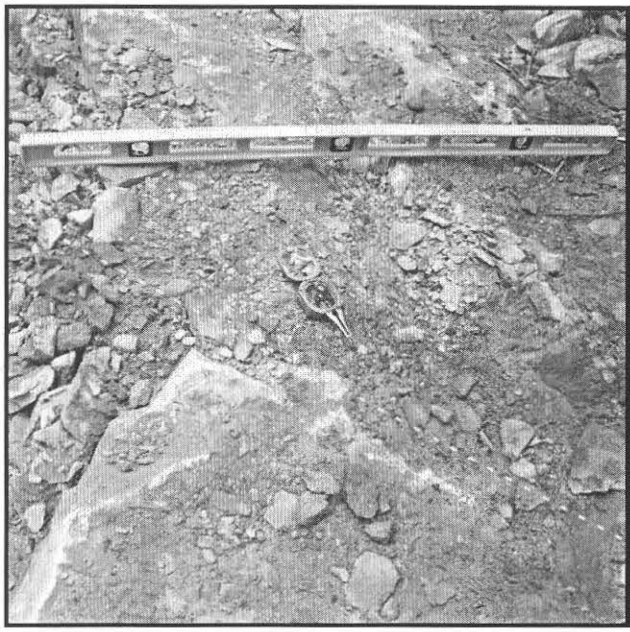


Figure 38. Sulfide Mineralization Along NW-trending Fracture. Fracture orientation is highlighted with dotted line.

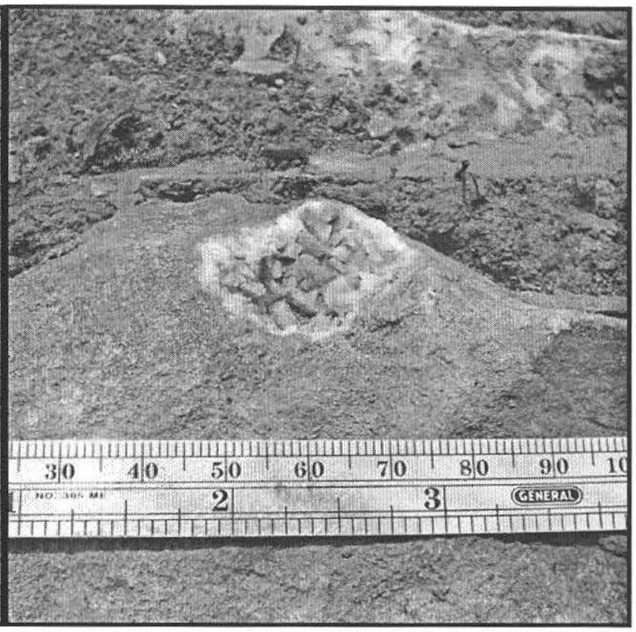


Figure 39. Calcite-filled Vug Within Tribes Hill Formation.

Rt. 167 Overlook **GPS (NAD27): 514585E 4765080N**

The scenic overlook of Rt. 167 just before entering the city of Little Falls, NY provides an excellent vantage point to appreciate the magnitude of the fault offsets in the region (Fig. 40). Though the Little Falls Fault is not directly exposed at this locality, the steep linear scarp must be a close approximation of the fault plane. If one parks in the designated area and walks to the fence, a grand overview of the Mohawk River Valley opens up before you. In contrast to the narrows at Little Falls, NY, in the resistant Precambrian basement, the valley opens broadly to the east in the much less resistant shales of the Utica Group. If the gate is open, it is possible to walk out onto a small outcrop of Precambrian basement at the edge of the scarp. The foliation here is well defined and trends 284/26NE. This foliation is fairly uniform throughout the Little Falls area. Fractures in this outcrop trend NNE and WNW and are mutually abutting. The NNE-striking fractures (average strike = 037°) of this outcrop parallel the topographic lineament and the inferred trend of the major splay ("C") of the Little Falls Fault in this area. While standing on this outcrop of Grenvillian basement, one can look down into the valley approximately 150 feet below and see the Thruway Interchange and its impressive exposures of the Upper Ordovician Indian Castle Member. Cushing (1905) estimates the throw on this fault at this location at 650-750ft (198.1-228.6m), whereas Fisher (1954) estimates 900ft (274.3m). The fault extends northward into the crystalline rocks of the Adirondack Dome and southward where it dies out in the Silurian shales. Across Rt. 167 from the scenic overlook is an old quarry. This quarry contains a 16m section of the lowermost Little Falls Formation, though the lower contact with the Precambrian basement is not exposed here. This section of Little Falls dolostone is very coarse-grained and contains large amounts of quartz and some large detrital feldspar grains. Many beds display cross-bedding and a few isolated horizons host vuggy porosity.

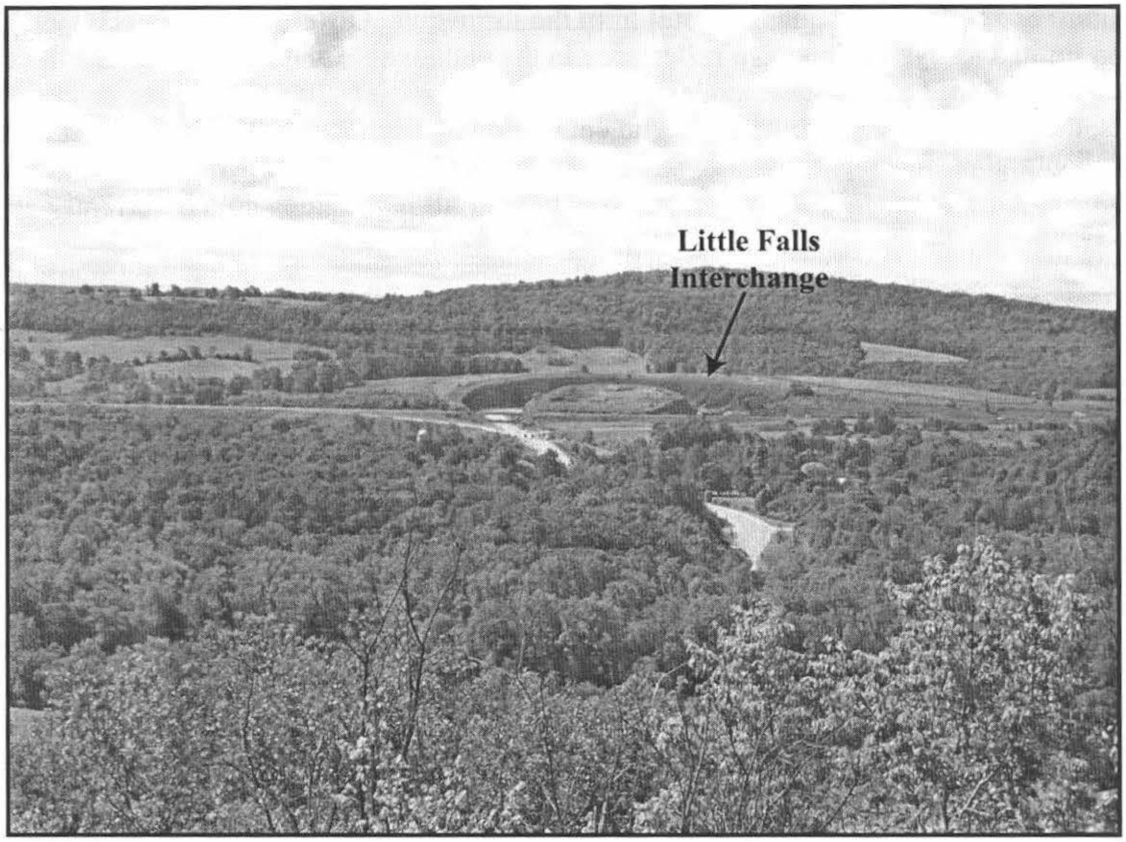


Figure 40. View from Rt. 167 Overlook. Photo is taken looking SSE. Photographer is standing on Precambrian basement looking down across the Little Falls Fault into the Mohawk Valley and the Indian Castle Member exposed in the Little Falls Interchange

Little Falls Bike Path Area **GPS (NAD27): 511580E 4764860N**

The bike path on the south side of the Mohawk River now occupies an old railroad bed. Walking eastward from the designated parking area, large exposures of Precambrian basement can be observed. Approximately halfway through the cut about 20ft from ground level on the south side, the contact between the Precambrian basement and the Little Falls Formation is exposed (Fig. 41). The contact is inaccessible without climbing gear, but clearly visible. Walking back to the parking lot and heading south, there is a footpath that winds its way up above the bike path. Along this path, there are excellent exposures of Little Falls dolostone. These outcrops are buff-weathering, thickly-bedded, coarse-grained dolostone that display cross-bedding, stromatolites, and numerous vugs. These vugs are filled with saddle dolomite, quartz, and anthraxolite (Fig. 42). Dolomite usually lines the vugs, followed by anthraxolite, though a few vugs contain anthraxolite films that pre-date the dolomite. Where quartz is present, it is always the last mineral to precipitate, although anthraxolite commonly occurs as solid inclusions within the crystals. The fact that the age relationships of dolomite and anthraxolite are inconsistent, suggests that these minerals may have significant overlaps in their time of precipitation. Dunn and Fisher (1954) analyzed the anthraxolite from three similar vugs in the area and obtained an average composition of 90.39% carbon, 7.03% water, and minor components of nitrogen, sulfur, and ash. The temperature of precipitation of these vug-filling minerals is poorly studied. Dunn and Fisher (1954) cite Dr F.G. Smith as having obtained a temperature of 51°±2°C for a quartz crystal from the formation. O'Reilly and Parnell (1991), based on limited analysis of bitumen reflectance in the anthraxolite and fluid inclusions in calcite and quartz, developed a four-stage model in which the first stage (<70°C) deposited calcite cements, the second stage precipitated calcite veins and vug fills (70 to 130°C), a third stage of anthraxolite (and presumably dolomite) and quartz (>120°C), and a fourth very high temperature stage 200±10°C. Further fluid inclusion analyses of vein fill in the study area is currently underway.

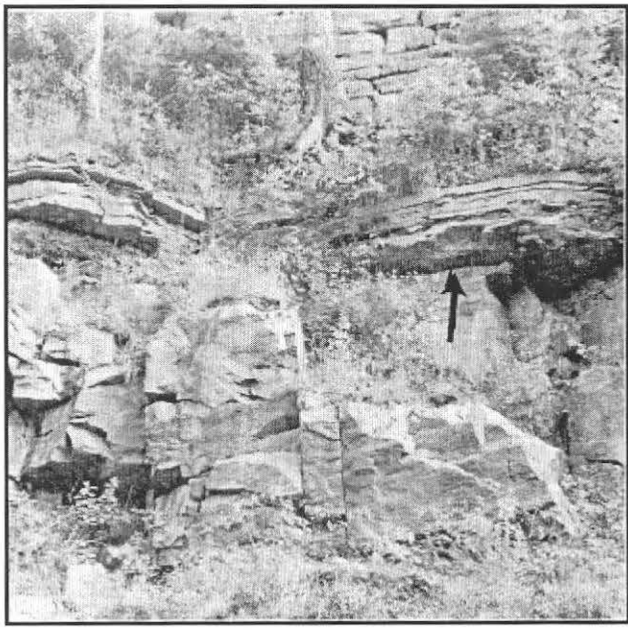


Figure 41. Precambrian-Little Falls Formation Contact (arrow). Exposed in wall of bike path. From Zenger (1981).

Figure 42. Vugs in the Little Falls Formation. Vugs filled with anthraxolite found in exposures on footpath above the bike path.

Nowadaga Creek **GPS (NAD27): 517265E 4760140N**

The exposures along Nowadaga Creek at the River Rd. bridge offer an opportunity to study the Dolgeville Formation and the contact with the overlying Indian Castle Member. The creek bed downstream of the bridge exposes the upper surface of the Dolgeville Formation and is readily observable at low water levels. The upper beds of the Dolgeville Formation are a complexly interference-folded surface (Fig. 43) and the contact with the overlying Indian Castle Member is very abrupt. For a detailed study of the folds and paleoflow directions at this locality see Jacobi and Mitchell (2002). Downstream from the contact is exposed a nearly complete section of the Dolgeville Formation. Very small (10cm wavelength) asymmetric folds with irregularly-bearing axes are the result of mini-blind thrusts involving single limestone beds (Marsh, 2000). These thrusts suggest that the entire Dolgeville section was deposited during times of instability. The fractures present within the Dolgeville trend NNE, WNW, and NW. There are also veins present in the Dolgeville Formation at this locality. These veins parallel the NNE fracture trend and range in thickness from 1-3mm of white calcite. Upstream of the bridge, the creek is floored by the Indian Castle Formation. This abrupt transition from contorted Dolgeville ribbon limestones and shales to undisturbed calcareous black shales represents a significant time gap, but not as much as that at the Thruway Discontinuity (Jacobi and Mitchell, 2002). In an outcrop on the northern bank, the lower Indian Castle Member is seen to contain a few thin carbonate beds, but overall the lithology is dominated by black calcareous shale. The fractures hosted within these shales show three prominent orientations: NNE, WNW, and NW (Fig. 44).

Figure 43. Interference-folded Upper Surface of the Dolgeville Formation. Downstream of the River Rd. bridge on Nowadaga Creek

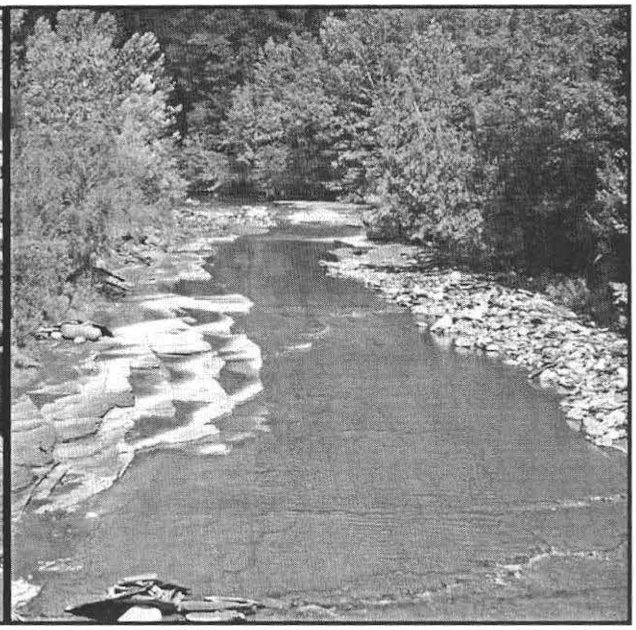


Figure 44. Fractures Within Lower Indian Castle Member. Upstream of River Rd. bridge on Nowadaga Creek. Photo taken looking NW.

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Depositional and Tectonic Models for Upper Devonian Sandstones in Western New York State

Saturday Field Trip A2

Sandstone Outcrops from Allegany, Cattaraugus and Wyoming counties.

Guidebook for the field trip held October 7th, 2006 in conjunction with the 35th Eastern Section AAPG Meeting and 78th NYSGA Field trips held in Buffalo, New York

Gerald J. Smith and Robert D. Jacobi,
Field trip leaders and guidebook authors

UB Rock Fracture Group
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Field trip sponsored in part by NYSERDA (New York State Energy Research and Development Authority)

October 7th, 2006

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INTRODUCTION

Goals and Objectives

The intention of this field trip is to visit examples of the major clastic depositional environments prevalent in the Late Devonian Appalachian foreland basin. Additionally, we will examine how syndepositional faulting influenced, and in some cases controlled, the deposition of the major petroleum reservoirs. This field trip represents a continuing evolution in our understanding of the Upper Devonian depositional controls. Our earliest sandstone field trip (Jacobi et al., 1990) proposed turbiditic and storm depositional environments in a basin setting, consistent with the views of the time that the Appalachian Basin in New York State was essentially structurally featureless (except for rare folds in the south and the Clarendon-Linden Fault), and that the shorelines had been muddy. Our next field trip (Jacobi and Smith, 1999) showed that post-depositional faulting was likely, that sandy/silty shoreface environments were possible farther east, and that syndepositional faulting affected the Rushford Formation.

In this field trip, we will visit outcrops representing the major sandstone depositional systems (turbidites, shelf-ridges, beaches, and fluvial) which correlate with the main Upper Devonian oil and gas reservoir sandstones of the Elk, Bradford and Venango groups. At each stop we will point out sedimentary structures, trace fossils and bedding relationships that assist in determining the depositional environment. We will also discuss the significance of syndepositional deformation structures that occur ubiquitously within Allegany and Cattaraugus counties of New York State. These deformation structures, as well as other data sets (e.g., paleoflow indicators, structure contour maps, isopach maps, and ichnofacies), indicate that the depositional systems were influenced by an interplay among fault block rotation, eustatic sealevel and sediment supply.

OVERVIEW

The Devonian stratigraphic section is well represented in outcrop in the Appalachian Basin, with significant continuous sections occurring within New York State. Correlations of Lower and Middle Devonian stratigraphic units can be carried across long distances throughout the basin, but Upper Devonian stratigraphic correlations become difficult not only over long distances, but also over a township or across a river valley. The difficulty arises from a combination of deposition within an energetic system and recurring seismic activity, which modified the basin topography.

The reservoir sandstones of the Upper Devonian in New York and Pennsylvania are commonly conceptualized as "discontinuous bodies within marine shales" (Woodrow, et al., 1988). A multitude of names based upon the field location, rather than the stratigraphic unit, reflects the difficulty in correlation of Upper Devonian units on the basis of lithology. The problem is the lateral variability of sandstone and conglomerate units, referred to as lenses or lentils from 1902 onward (e.g., Clarke, 1902; Glenn, 1903; Tesmer 1955). The term "lens" provides an inadequate description of the unit, since sand-filled channels, sand ridges and remnants could all imply a lensing morphology (Figure 1). It is the tacit expectation that sedimentary units maintain a constant lateral thickness and lithology that in part drives the perception of lensing sandstones and the inability to correlate over long distances. Lateral variation is inherent in sandstone deposition; whether a tidal unit possesses internal mud-drapes or a tempestite contains basal coquina lags, the unit will not be laterally homogeneous in a shallow marine environment. Monolithic sands units of constant thickness will not form in typical clastic depositional environments; instead, a sand-packet comprised of several events will generate a thick sand unit. Accompanying this problem is the extrapolation of essentially point source data (well-logs) to reflect a broader area. A well-log will provide geophysical data for a narrow window surrounding the well-bore; two wells in the same sand-packet may yield different log responses, and therefore suggest no possible correlation between the wells, whereas outcrops would readily demonstrate that the two wells were sampling the same unit (Figure 2).

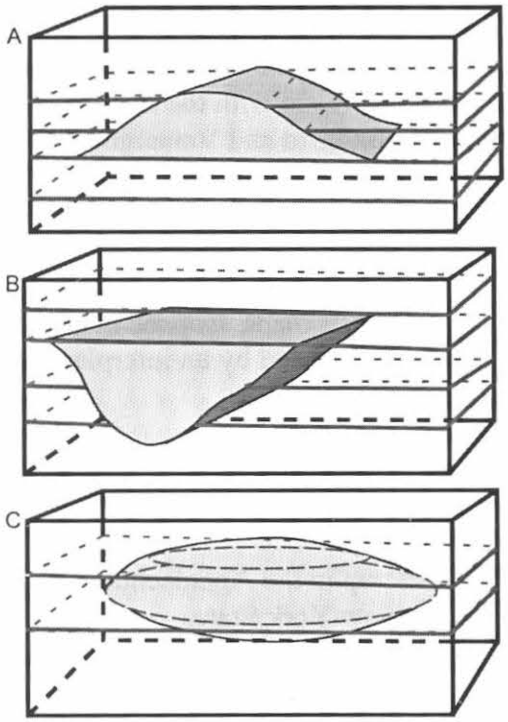
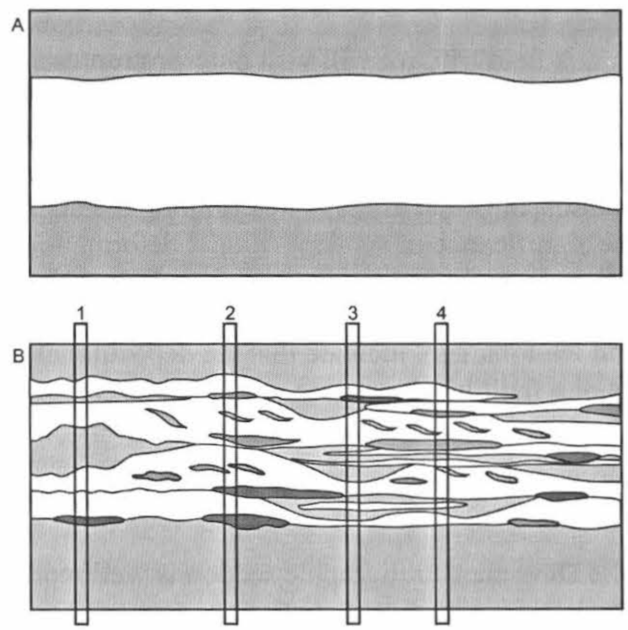


Figure 1. Variations in morphologies for lenses. A - ridge; B channel; C - erosional remnant.



Sandstone
Shale
Coquinite (shell-lag)
Coarser sand-conglomerate (lag)

Figure 2. A - Cross-section of a homogeneous sand body that rarely occurs but are assumed to represent any correlateable sandstone. B - hypothetical example of a normal sandstone packet made up of several events with a distinct heterogeneous lithology. If a wells were drilled a locations 1, 2, 3, or 4 the logs would yield a "lensing" characteristic that is due more to lateral variation than as an isolated ridge or channel.

The focus area contains the Upper Devonian groups that comprise major hydrocarbon reservoir units in New York State. The sandstones in the upper part of the Canadaway and Conneaut groups are correlatives to the Bradford sands in Pennsylvania, whereas the sandstone and conglomerates in the Conewango Group are correlative to the Venango sands in Pennsylvania (Figure 3). Examination of the boundaries of the reservoirs shows a strong relationship between faults, folds and the oil fields (Figure 4 and 5). Syndepositional faulting along the Clarendon-Linden Fault System controlled the orientation of clastic deposits in Allegany County (Smith and Jacobi, 1996, 1998, 1999, 2001). Similar reactivation along other north-south trending faults in Cattaraugus

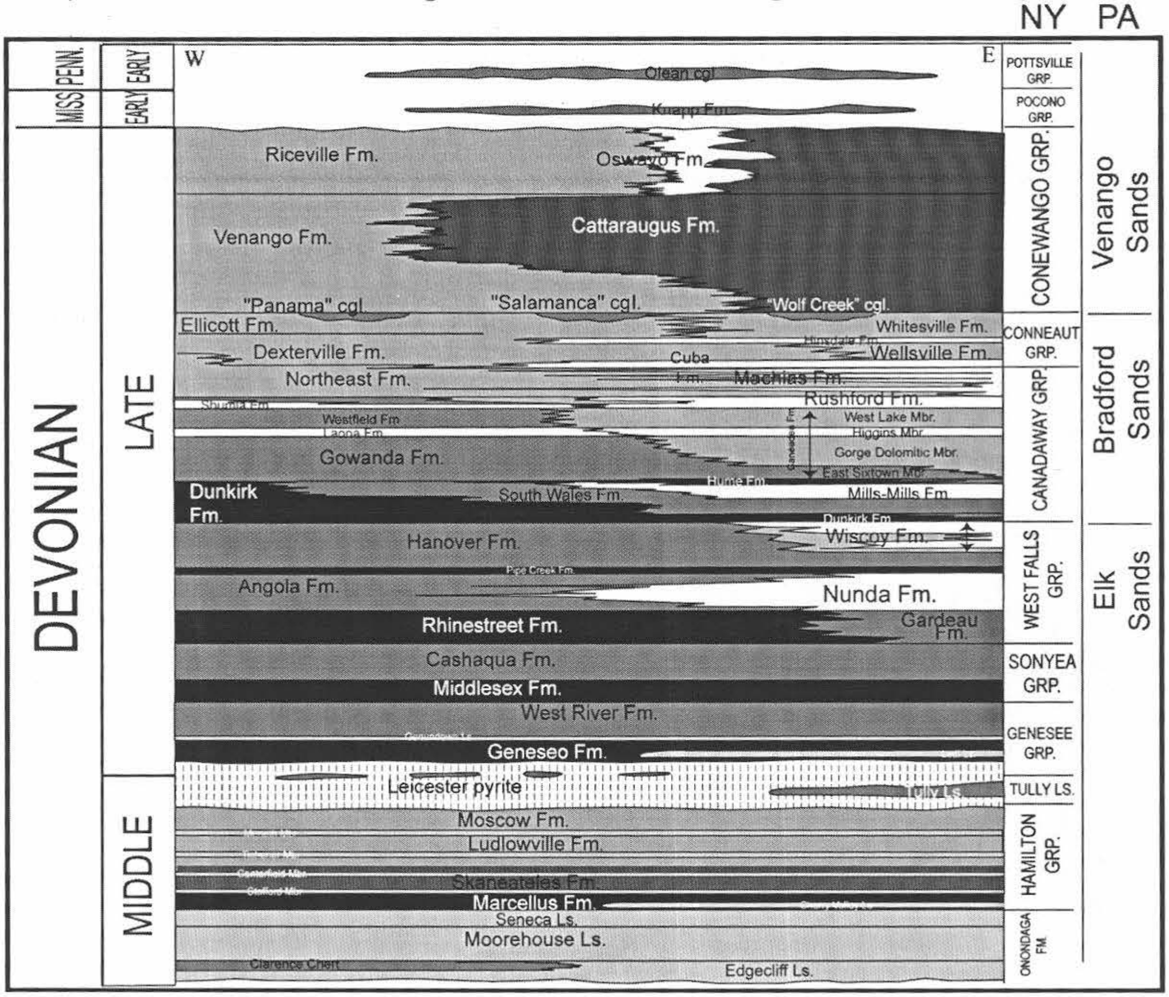


Figure 3. Devonian stratigraphic section for western New York State displaying oil and gas equivalent units. Based upon Rickard, 1975.

County also influenced the depositional trend of Upper Devonian reservoir and source rocks.

Other trends, such as the NE-trends, have a more complicated origin. Certainly Alleghanian folds form the structural closure that controlled many of the well-known fields such as the Sharon-Smithport anticline and pool. However, in addition, Iapetan opening faults that arc through Pennsylvanian and New York were reactivated and these fault-block reactivations controlled the depositional fabric of many of the sands, such as the Bradford field and the Elk sands (Jacobi et al., 2004, 2005, 2006)

The occurrence of northeast-trending sedimentary deposits such as shorelines and sand ridges were assumed to follow the paleoshoreline. Reconstruction of Devonian

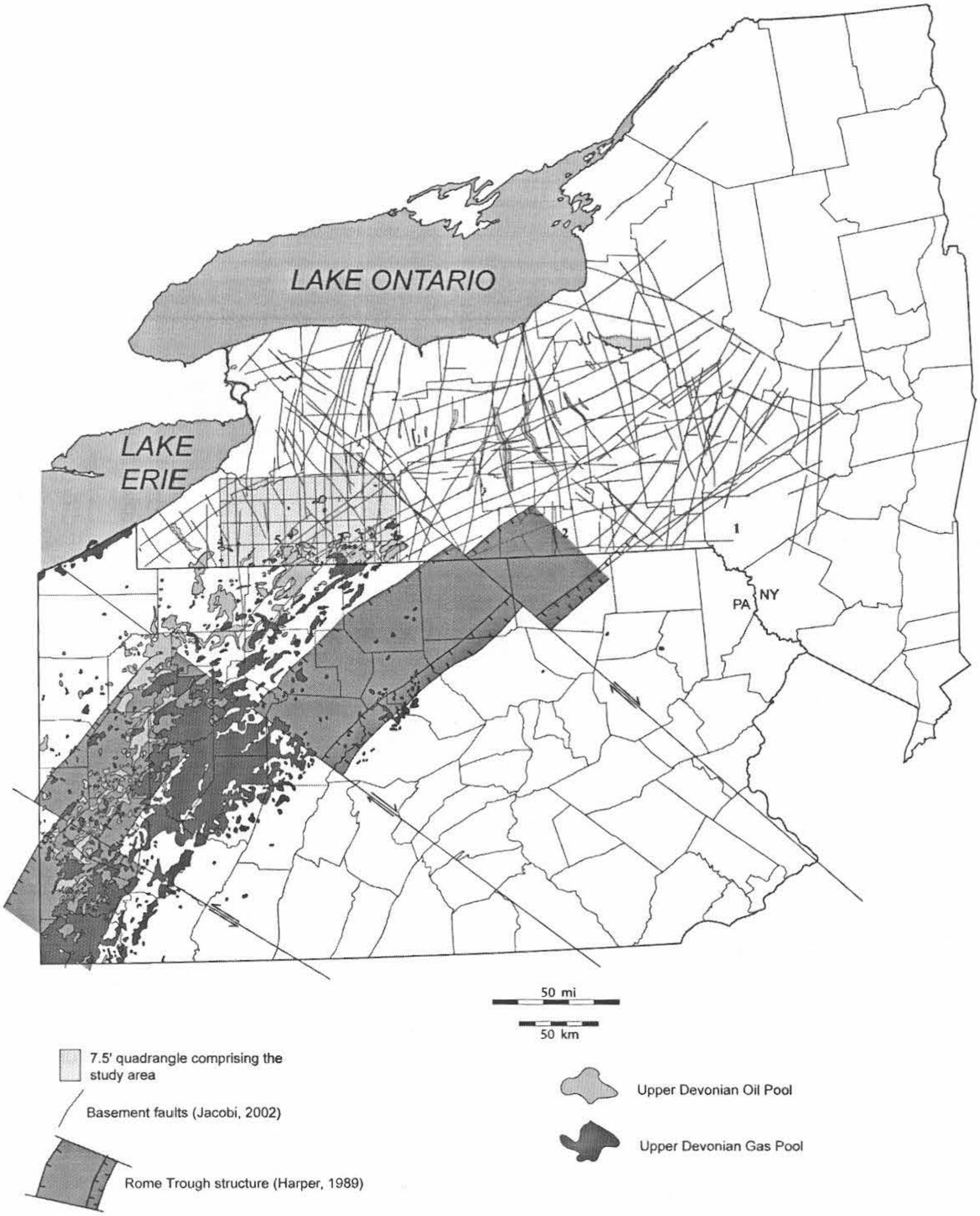


Figure 4. Upper Devonian oil and gas fields in relation to known and proposed basement structrures

paleoshorelines in West Virginia, Maryland and Pennsylvania by Boswell and Donaldson (1988), Dennison (1985) were derived from well-log data, based upon the sand

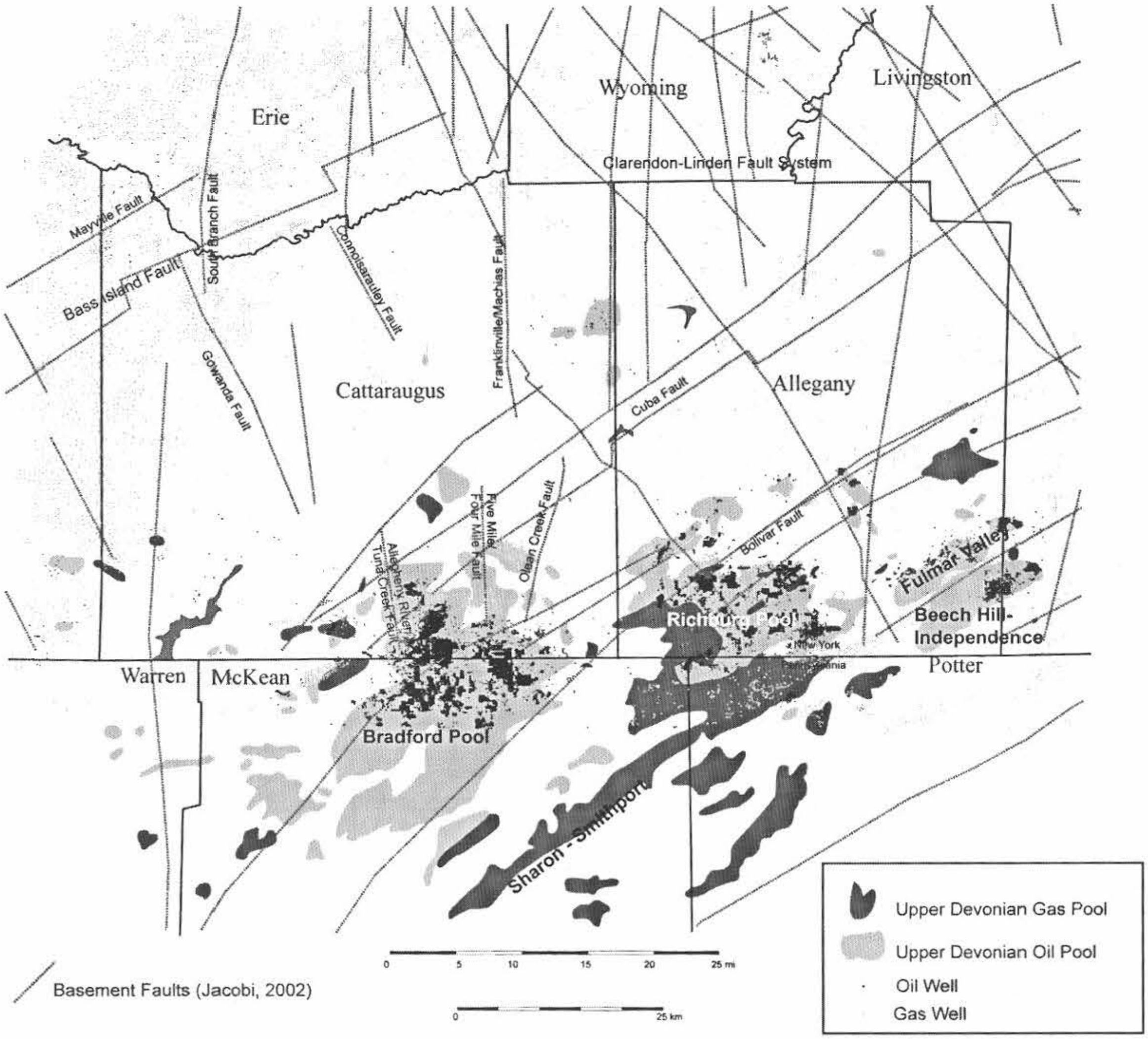


Figure 5. Apparent relationship between oil and gas fields and basement structures (after Jacobi et al., 2004)

percentage for a particular stratigraphic unit. Basinward limits and orientations for similar time periods seldom agree. For example paleoshorelines 4&5, 6&7 in Figure 6 both represent similar time periods in the Conneaut and Conewango groups, yet display differing trends in New York State. The coincidence between the basement structure and the location of the sedimentary deposits suggested that the paleoshoreline itself may have been controlled by reactivation of the Iapetan opening/Rome Trough fault system (Jacobi et al., 2004).

It is the intention of this study to correlate the "discontinuous bodies within marine shales", determine the architecture (size, shape and orientation) of the different sandstone packets and examine controls on the location and architecture of the sandstone packets. By understanding the architecture of the sandstone packets and what is controlling that architecture it is possible to better characterize known reservoirs and extrapolate the potential into unexplored locations.

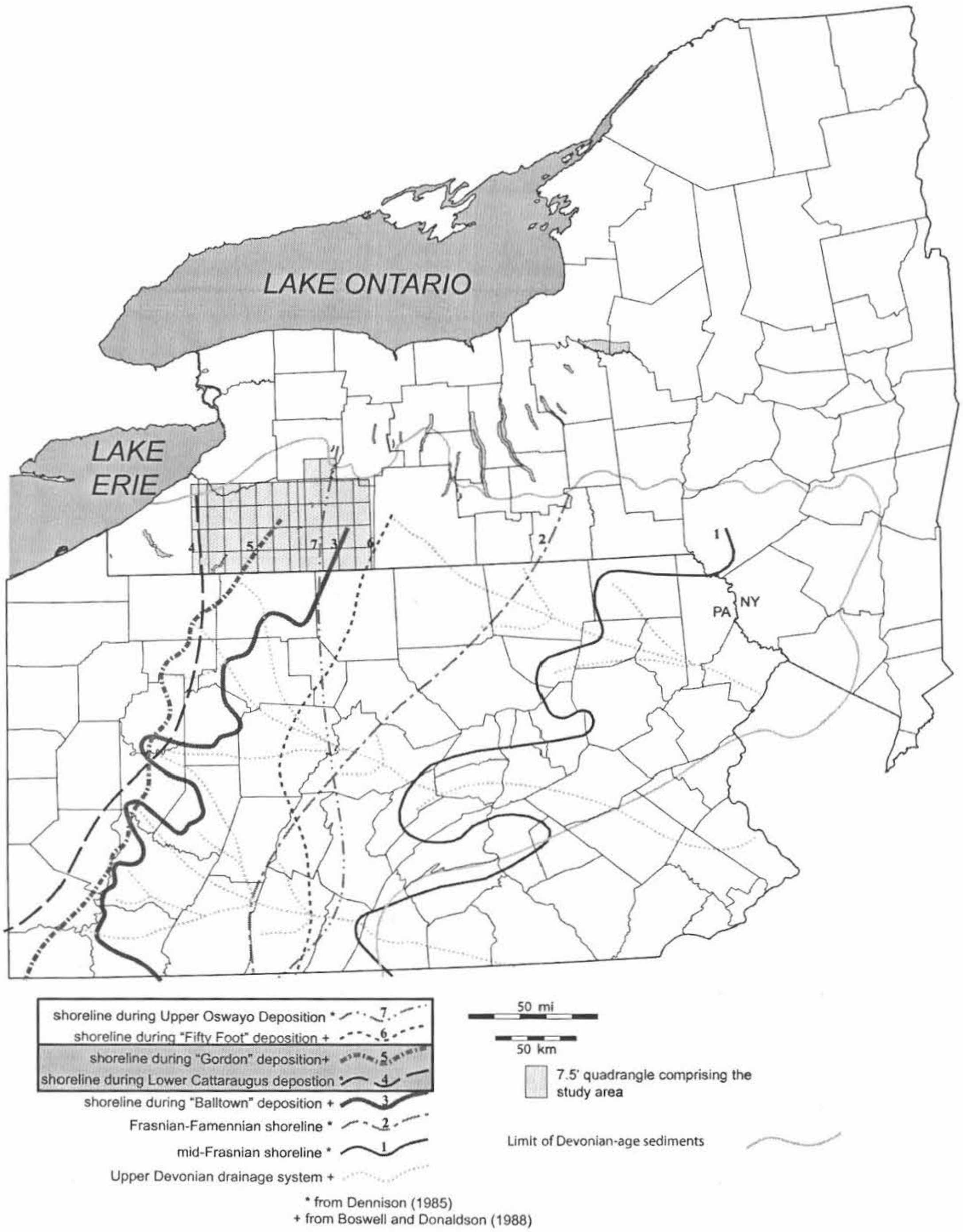


Figure 6. Paleoshorelines for different periods of the Upper Devonian. Note the "boxed" units (4&5, 6&7) represent very similar time frames in the Late Devonian.

Study Area Location

The study area is comprised of 46 7 ½' topographic quadrangles that cover the majority of Allegany and Cattaraugus counties (Figures 4, 5 and 6). The study area expands upon 15 years of fieldwork we have conducted in and around the Allegany – Cattaraugus county region. The outcrops within the area consist of Upper Devonian sandstones and shales of the West Falls, Canadaway, Conneaut and Conewango groups, and occur in streams exposures and road cuts. While a substantial number of oil and gas wells occur within the study area; the overwhelming majority of wells are located near the southern border of New York State, resulting in a sporadic covering of wells over approximately ¾ of the study area.

Methodology

The sedimentological and stratigraphical data were collected from 1991 through 2006 At each site, the location of the outcrop was obtained in recent years using a Garmin 76CSx GPS, and location coordinates were transferred to USGS 7 1/2' topographic maps. In earlier years, we used the topographic maps to locate the sites. The stratigraphic thickness of each distinct lithological unit was measured to the nearest millimeter; all sedimentary and bedding structures were also recorded for each bed.

Measurement of paleoflow orientations were taken with a Brunton compass corrected for the magnetic declination for the quadrangle studied at the time. Each outcrop was also carefully examined for trace fossils so that the ichnology and changes in ichnofauna could be used to supplement interpretations of the depositional environment. Annotated, scaled stratigraphic columns were made in Adobe Illustrator for each measured site.

Well log analyses were performed using scanned well-logs provided by the New York State Museum and the well-log viewer BlueView by Schlumberger. Only the gamma ray curves were examined, to enable consistent formation picks between wells. Of the 418 wells examined, 124 gamma ray curves were hand digitized in Adobe Illustrator to 1) allow better comparison between wells with different vertical scales, 2) enable the comparison between wells and outcrop stratigraphic columns, and 3) use both wells and outcrops in cross-sections. Only Devonian formations were examined for the purpose of this study. Well-log data were then entered into a GeoGraphix database, along with outcrop data for the Dunkirk Fm., Hume Fm., Lower Rushford Mbr., Machias 1st Sand, Cuba Fm. and Hinsdale Fm., to enable the generation of isopach and structure contour maps. Contouring in GeoGraphix was performed using a Kriging function at 30 iterations, without a geological bias to the data. The resulting contours were not subsequently modified to avoid imposing intentional or unintentional biases to the mapping.

Stratigraphic Nomenclature

One of the persistent problems in Upper Devonian stratigraphy is the plethora of unit names that can refer to 1) the same unit (but with different names), 2) any oil-producing unit, 3) regional location of the well, 4) part of the same unit but with different tops or bases, and 5) a gross simplification of several units into one catch-all name.
For this report, we will follow the lithostratigraphic names that use outcrop-defined units that we have established in earlier studies (i.e., Smith and Jacobi, 2000 and 2001; Smith

Pennsylvanian		Pottsville Grp.	Olean Cgl.	
Mississippian		Pocono Grp.	Knapp Creek Fm.	
DEVONIAN	Famennian	Conewango Grp.	Oswayo Fm. ★	
			Cattaraugus Fm.	
			Salamanca Cgl. ★	
		Conneaut Grp.	Whitesville Fm. ★	
			Hinsdale Fm.	
			Wellsville Fm. ★	Wellsville 3rd
				Wellsville 2nd
				Wellsville 1st
			Cuba Fm.	
		Canadaway Grp.	Machias Fm. ★	Machais 4th
				Machias 3rd
				Machias 2nd
				Machias 1st
			Rushford Fm.	Upper Rushford Mbr. ★
				Intermediate Rushfrod Mbr.
				Lower Rushford Mbr.
			Caneadea Fm.	West Lake Mbr.
				Higgins Mbr. ★
				Gorge Dolomitic Mbr.
				East Sixtown Mbr.
			Hume Fm.	
			Mills-Mills Fm. ★	
			South Wales Fm.	
			Dunkirk Fm.	
	Frasnian	West Falls Grp.	Wiscoy Fm.	
			Hanover Fm.	
			Pipe Creek Fm.	
			Nunda Fm.	

★ - see text for additional comments

Table 1. Stratigraphic units for outcrop in Allegany and Cattaraugus counties. Units we will visit during this field trip indicated with a box around the name.

Some noted changes and observations in this stratigraphic section include:

Mills-Mills Fm. – We had commonly used the name Canaseraga Fm (Chadwick, 1923) as the formal name for this unit (i.e., Smith and Jacobi, 1998, 2000 and 2001), but upon further study, the Canaseraga Fm., as defined by Chadwick (1923) encompasses both the South Wales Fm., and sandstones we refer to as the Mills-Mills Fm. Since both the South Wales and Mills-Mills formations are distinct lithologically and both are mapable over a large areal extent, it seems reasonable to identify the Mills-Mills as a formation, and to be correlative to only part of the Canaseraga Fm.

Higgins Mbr. (Caneadea Fm.) – This unit is the probable eastern correlative of the Laona Fm., and contains the stratigraphically lowest seismite zone observed within the Canadaway Grp.

Upper Rushford Mbr. (Rushford Fm.) – This unit is comprised of several lensing sandstones that are relatively thin (1-2 meters) in Allegany County, but form a thicker (3-6 meter) sandstone packet in Cattaraugus County.

Machias Fm. – The original description by Chadwick (1923) described the Machias Fm. as primarily shale. Later studies by Woodruff (1942) and Manspeizer (1963) noted one or two thick sandstones and/or limestones. From outcrops examined during the past 15 years of fieldwork, as well as from well-log analyses, we have informally identified four traceable sandstone packets referred to as the Machias 1st through 4th.

Wellsville Fm. – This unit is similar to the Machias Fm.; we have informally identified three traceable sandstone packets: Wellsville 1st through 3rd.

Whitesville Fm. – although we observed several thick (1-3 meters) sandstone packets in outcrop, too few examples occurred to provide convincing correlations.

Salamanca Conglomerate – likely to be correlative to similar conglomerates at Wolf Creek and Panama, NY; each separate conglomerate possibly representing a separate incision valley formed during the same lowstand-transgressive sequence(s) marking the base of the Conewango Grp. Elevation variation between the separate localities may reflect different depth of valley incision, as well as later faulting.

Oswayo Fm. – depending on the classification system, the Oswayo Fm. is considered Mississippian-age in Pennsylvania, but in New York it is the uppermost Devonian unit.

Paleogeography

There are three main controls on any depositional environment: sealevel, structure and sediment supply. The interplay of all three controls will affect what can be deposited or eroded, as well as the size, shape and orientation of the final deposits.
The Acadian Orogeny began during the Early-Middle Devonian, forming the Acadian Mountains (the primary sediment source for the study area) and the Acadian Foreland Basin (the northern area referred to as the Catskill Sea) (e.g., Woodrow and Isley, 1983).

Paleomagnetics studies (Ziegler et al., 1979; Ziegler, 1988; Witzke and Heckel, 1988; Scotese and McKerrow, 1990 and Witzke, 1990) vary in exact placement of paleolatitudes for the study area, but generally concur that the area was located in the tropical region of the southern hemisphere, 15° and 30° S. The paleogeographic location of the study area (Figure 7) has two significant effects: 1) the Catskill Sea would have counterclockwise surface current rotation such that longshore currents would trend from the southwest to the northeast; 2) the region would likely be affected by a monsoonal climate, alternating wet-dry seasons, with intense, large storms.

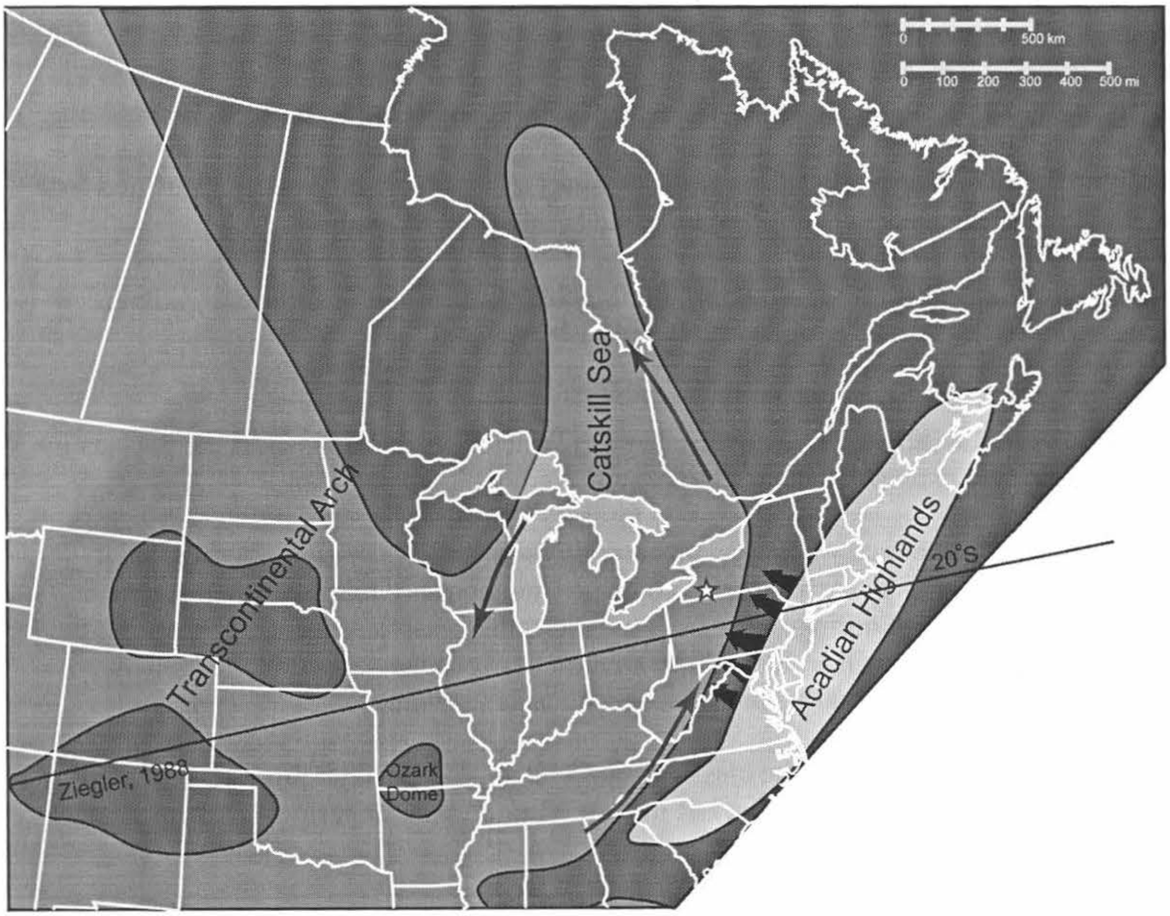


Figure 7. Catskill Sea paleogeography created by integrating models from Ettensohn (1985), Ziegler (1988), and Witzke and Heckel (1988). Paleolatitudes for the Late Devonain exhibit a wide variation in orientation, but all generally place the fieldtrip area (marked by the star) in the southern hemisphere from 5° to 32° south latitude.

Woodrow and Isley (1983) suggested that the Catskill Sea did not include the main bathymetric provinces of shore-shelf-slope-basin found in passive margin models; rather a gently sloping clinoform formed the margin. Such a province may be common for foreland basins (e.g., Pattison, 2005). The steady, shallow slope of the Catskill Sea would lead to a lateral gradation of clastic deposits that would shift dramatically with minor fluctuations in relative sealevel. Although the gently sloping clinoform of Woodrow and Isley (1983) describes the overall nature of the Catskill, it does not reflect smaller topographic variations within the basin caused by continual fault reactivations occurring during Late Devonian.

Tectonics and Regional Structure

Basement structures within the study area reflect the early geological history of eastern North America (see Figures 4 and 5). A reactivated intra-Grenvillian suture zone is locally expressed in the Paleozoic section as the north-south trending Clarendon-Linden Fault System (Jacobi and Fountain, 1993, 1996, and 2002).

Faults associated with Iapetan-opening/Rome Trough development are expressed as northeast-trending basement structures with complimentary northwest-trending cross-structures (Jacobi, 2002; Jacobi et al., 2004, 2005, and 2006). Syndepositional reactivation of these basement structures have been observed in seismic sections for the Ordovician Taconic Orogeny (Jacobi et al., 2004 and 2005) and for Devonian Acadian Orogeny (Jacobi, 2002; Jacobi and Fountain, 1996 and 2002). Syndeposition faulting is observable in the stratigraphic section; growth fault geometries are observed in the Upper Devonian Hume and Rushford formations (Smith and Jacobi, 2000, 2001 and 2002).

Further evidence for syndepositional fault activity are the numerous zones of seismites. Seismites are formed by the sudden dewatering of uncompacted sediments brought on by a sudden shock that is generally thought to be from a large magnitude earthquake (Figure 8). The ubiquitous occurrence of seismites at numerous stratigraphic horizons throughout the study area denotes seismic events of a magnitude greater or equal to magnitude 6. Earthquakes less than a magnitude 6 would only form seismites within 1 to 2 km of the epicenter whereas magnitudes of 6 or greater the distance from the epicenter expends to 20 to 110 km (e.g., Wheeler, 2002). At a magnitude of 5.5 earthquake, a maximum surficial displacement on the fault would be ~0.3m (Bonilla et al., 1984; dePolo and Slemmons, 1990; and Wells and Coppersmith, 1994). At greater magnitudes (M= 6-7), the surficial displacement can reach 1 to 2 meters along the fault. These small offsets may greatly impact deposition within the local area by reorienting currents, raising some areas into the fair-weather wave base, dropping other regions, and generally altering the accommodation space for the region. It is obvious that the interplay among fault block motion, eustatic sea level changes (Figure 9; Johnson et al., 1985), and sediment supply in this shallowly sloping basin can significantly alter the sediment architecture.

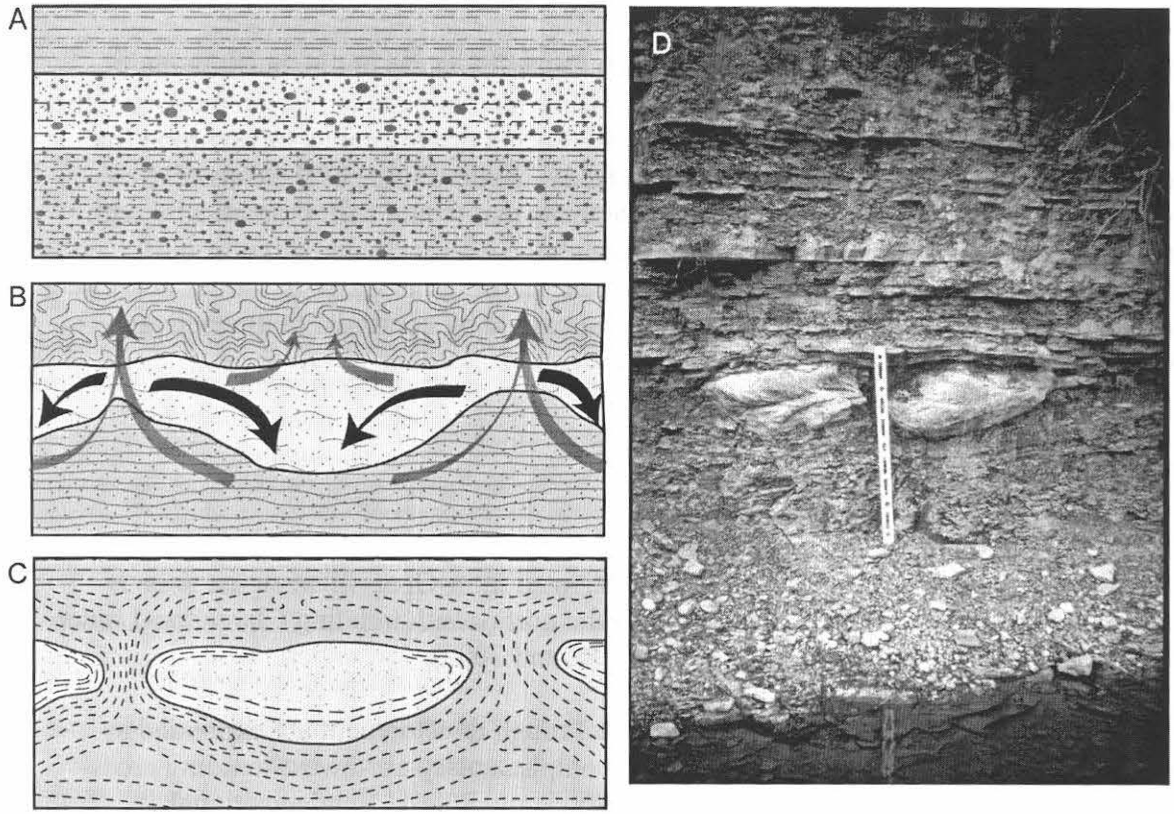


Figure 8. Stages of seismite formation.
A) In regions of high depositional rates, water saturated sediments are common.
B) A triggering mechanisn (generally assumed to be a seismic event greater than M=6) will cause rapid dewatering which will cause the water to escape toward the surface, bringing some of the clays up toward the surface and also cause the denser, heavier sands to sink into the underlying, unconsolidated clays.
C) The resulting structures will have sands which have turned or rolled-up edges, surrounded by a matrix of deformed shales.
D) A typical seismite occurring in outcrop in the Caneadea Formation. Ruler is 122 cm long.

DEPOSITIONAL ENVIRONMENTS

Assemblages of sedimentary structures, lithologies and ichnofauna enable the determination of water depth, current strength and salinity, which can be used to distinguish different depositional environments (Figure 10). From the collected field data we have determined the depositional environments that cover the stratigraphic section for Cattaraugus and Allegany counties (Figure 11).

Black shale deposits in the Late Devonian are thought to have formed from anoxia events (Ettensohn, 1994) rather than from great depths. High organic input combined with a stratified water column would produce a low-oxygen to anoxic system at shallower depths. Storm modified sandstones observed in outcrop within the Dunkirk and Hume formations suggest that these black shales formed within storm wave base. Above the anoxia boundary, the high organic content within the shales would oxidize, producing medium to light gray shales.

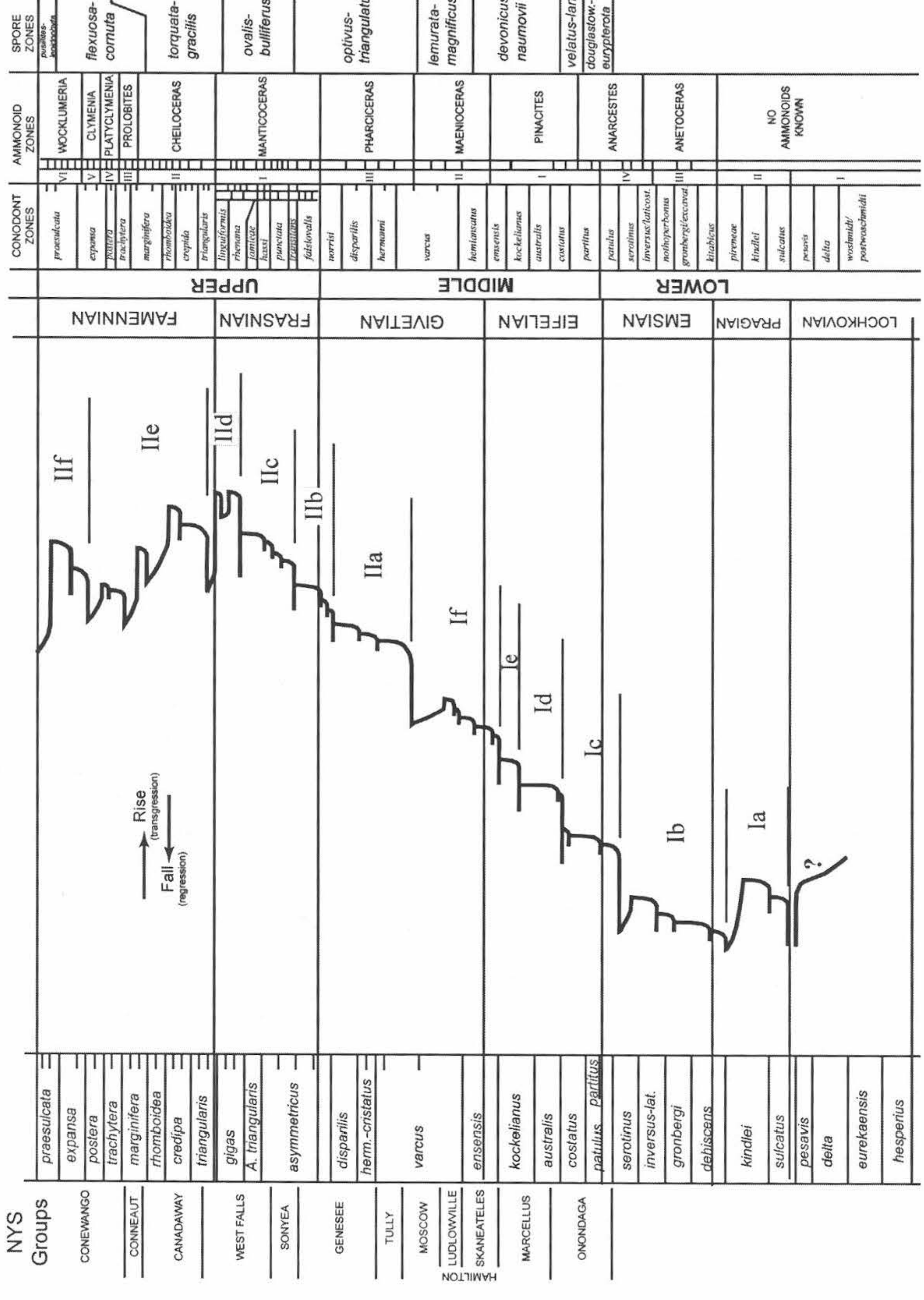


Figure 9. Eustatic sealevel curve from Johnson and others (1985) with biostratigraphic zones from Woodrow and others (1988) and House (2000).

Sand and coarser clastic materials were deposited in four major depositional environments: turbidites, shelf ridges, shoreface system, and fluvial. All four depositional environments are part of a larger deltaic system, but for describing depositional patterns and controls it is easier to examine the four parts separately.

Turbidites

Turbidites are formed from relatively dense, sediment-entrained currents flowing down-slope from a disturbance that introduces the sediment into systems (see for example, "Submarine Fans and Related Turbidite Systems", edited by Bouma, Normark, and Barnes, 1985; and "Fine-grained Turbidite Systems", edited by Bouma and Stone, 2000). Turbidites, in general, form sharp-based, fining-upwards deposits. Methods for initiating turbidites include up-slope slumps, storms or waves stirring up or eroding bottom sediments, suspended sediments introduced by rivers in flood stage and earthquakes. The resultant turbidity currents will construct submarine fans with a form controlled in part by the bathymetry of the basin and the sediment size carried by the turbidity flow. For steep slopes (such as the slope in a passive margin) and/or sand-rich environments (with a relatively close source of coarse sediment), the resultant submarine fan will be radial in shape (assuming relatively smooth pre-submarine bathymetry, Figure 12). For gentler slopes and/or mud-rich environments (with a distant sediment source), the resultant submarine fan may be relatively elongate in shape, since the finer sediment will be able to be carried farther into the basin (e.g., Stow, 1986; Bouma, 2000; Figure 12). For example, the turbidity current pathways on the west African margin can extend over 1000 km downslope to the abyssal plains (e.g., Jacobi and Hayes, 1982, 1992). The gentle slope of the Catskill Sea along with the fine-grained composition of the Nunda, South Wales and Mills-Mills formations would suggest that these units would form relatively elongate fan deposits orientated perpendicular to the paleoshoreline. However, characteristics (e.g., paleoflow data) of the South Wales Formation in western New York west of our present study area suggest a radial flow pattern, consistent with lobe fringe sands of Mutti (1977) (Jacobi et al., 1994). Shallower turbidite deposits, e.g., pro-delta fans of Pattison (2005) and wave-modified turbidites of Myrow and others (2002) form within storm-wave base. These shallower turbidites share similarities to turbidites with the exception of combined-flow ripples in the C part of the Bouma model (Brett, 1983; Myrow et al., 2002). The thick sandstones of the Nunda Formation and the Higgins Member of the Caneadea Formation, correlative to the Laona Fm. to west, possess many of the features attributed to wave-modified turbidite fans. West of the present study area, distal beds of the Nunda Formation were thought to represent sand lobes on a submarine fan, based on the massive nature, abrupt pinchouts, and lobate form of isopach maps of the thick sand beds (Jacobi et al., 1994).

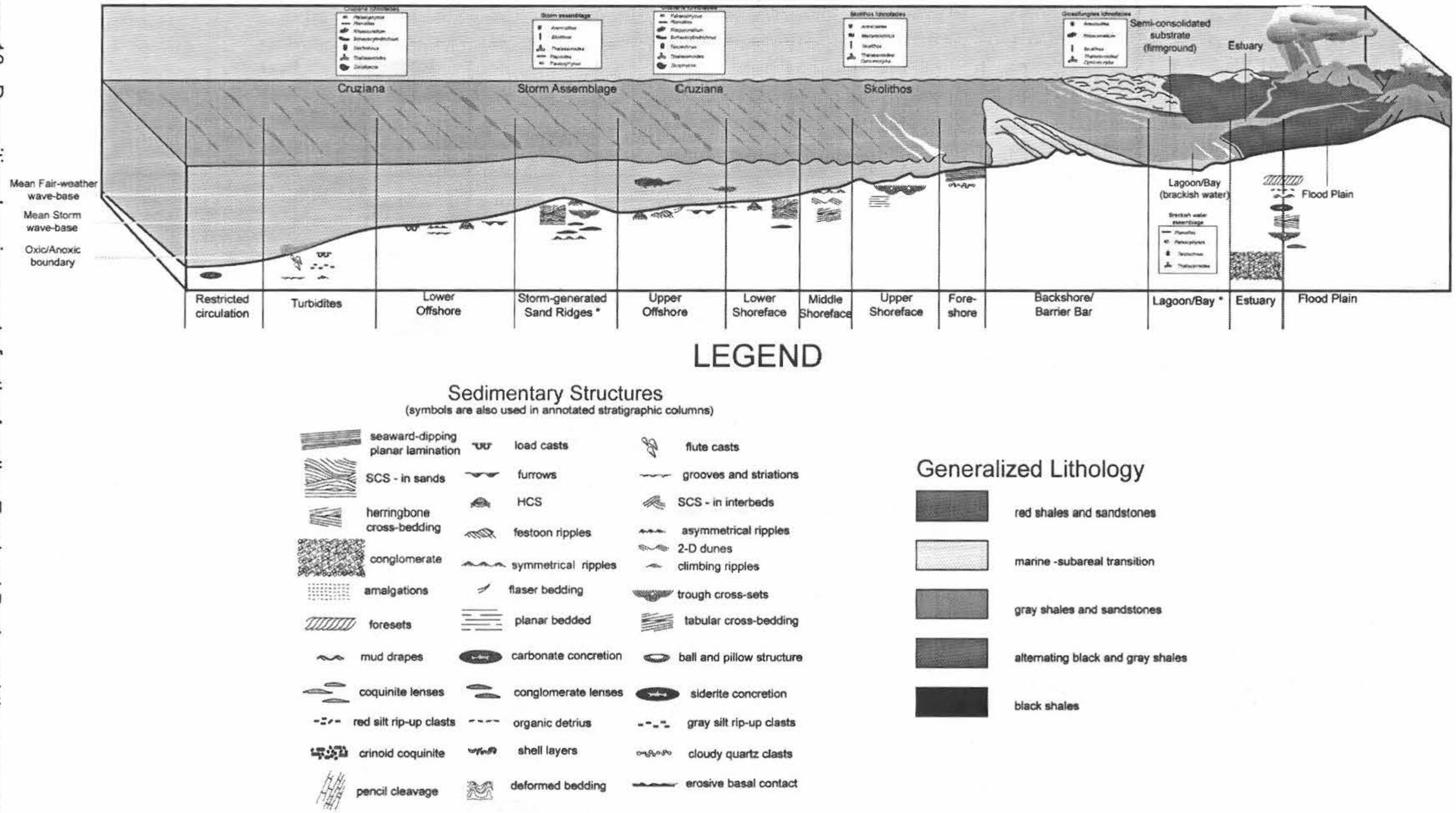


Figure 10. Depositional environments for the Acadian Foreland Basin and the common assemblages of sedimentary stuctures and ichnofacies.

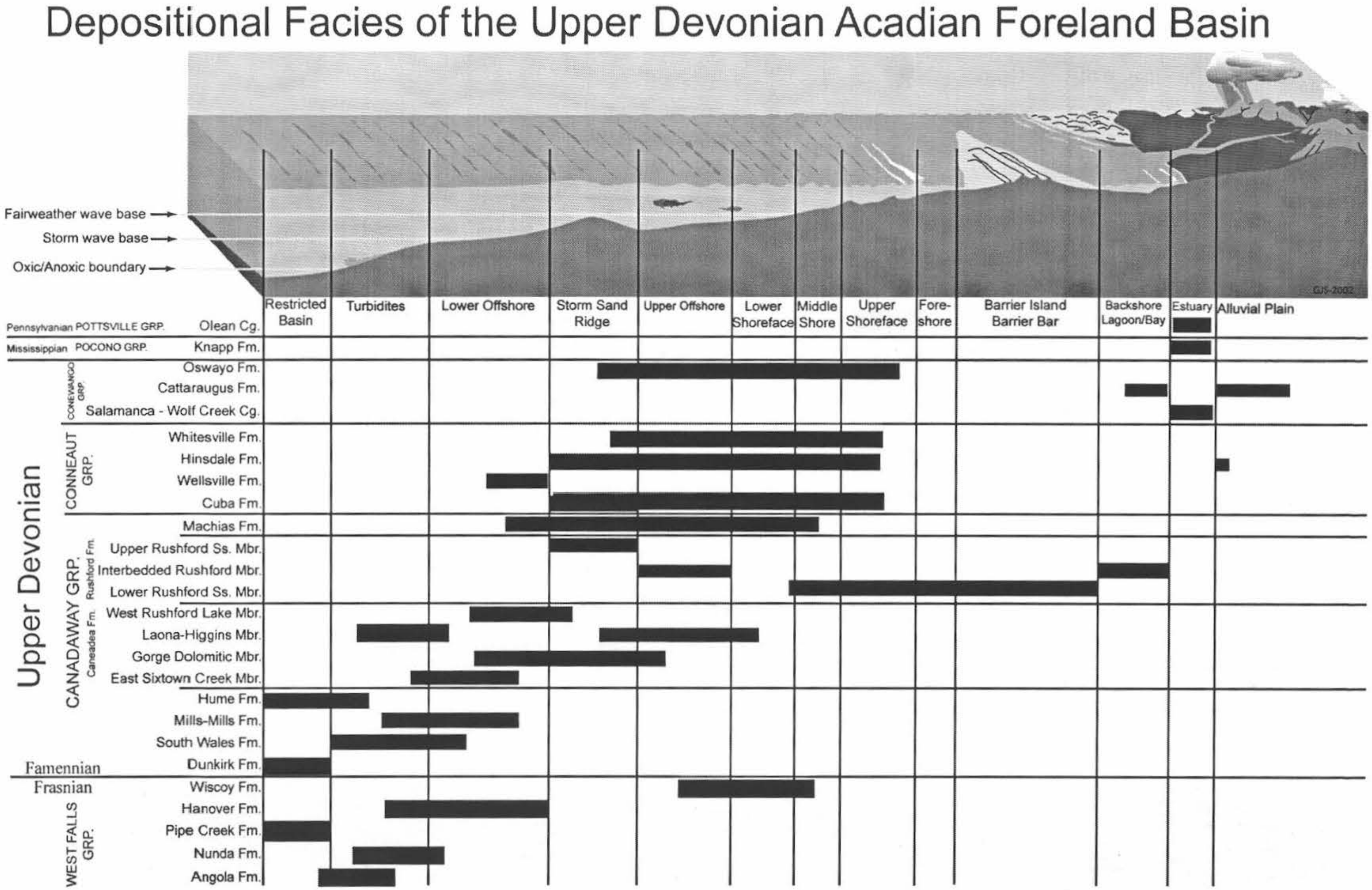


Figure 11. Depositional environments for the stratigraphic units within Allegany and Cattaraugus counties.

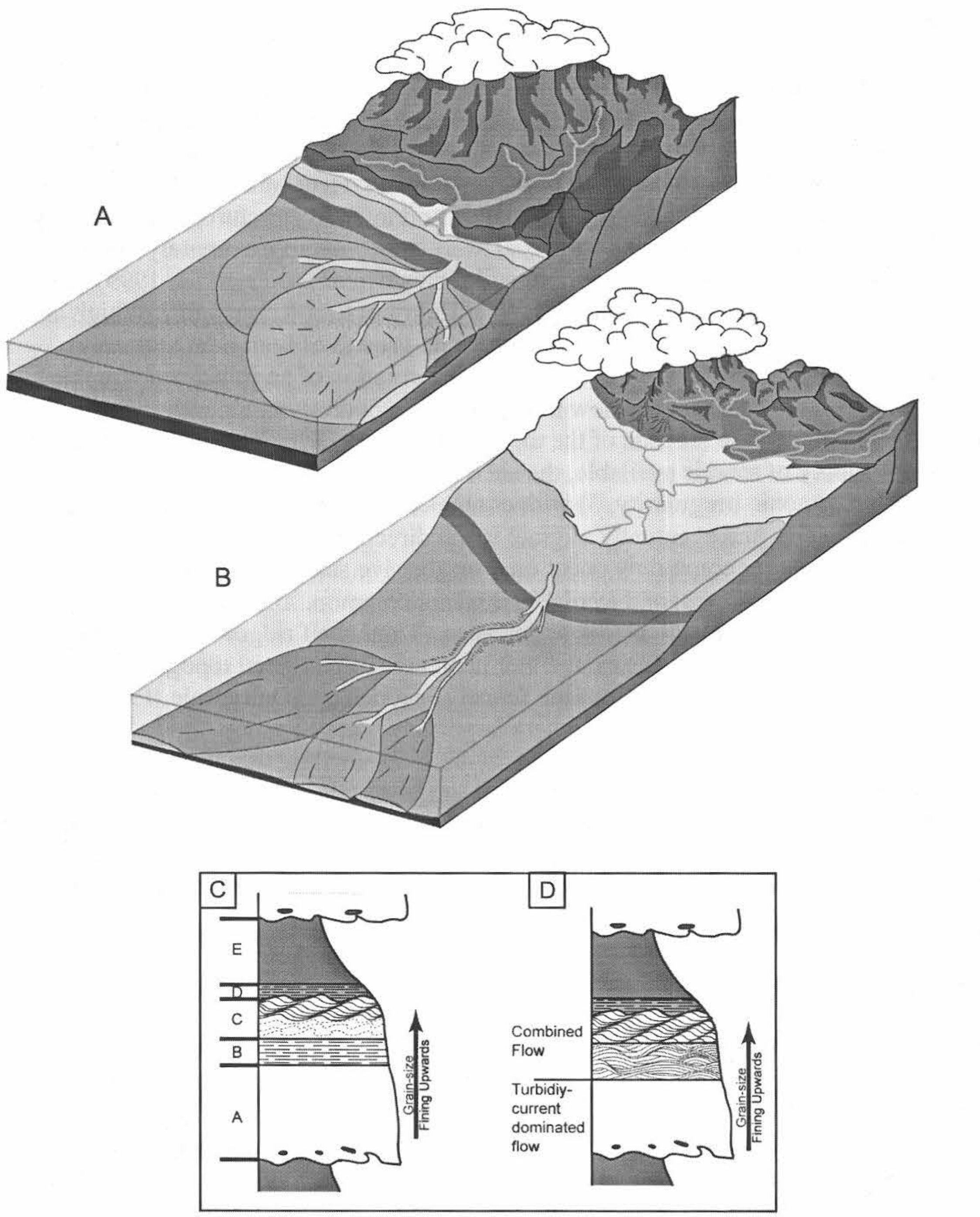


Figure 12. A) Where the sediment source is close to the shelf and typically comprised of sand and gravel and the slope is relatively steep and the pre-submarine fan bathymetry is relatively smooth, the resulting turibidite fan will be round and lobate. B) Where the sediment source is far from the shelf and typically comprised of shale, silts and fine sands the turbite fan will be farther from the shore, longer and narrower than the coarser endmemer in "A". The turbidites in the Late Devonian of western New York State may follow this model (modified from Bouma, 2000). C. Idealized turbidite model proposed by Bouma (1962), depicts the fining-upwards sequence and changes in bedding causes by decreasing flow velocity. Bouma (1962) model is proposed for a sand-rich system (depositional model A) but can still be applicable to finer-grained systems (depositional model B), although turbidite rarely exhibit a complete sequence (T_{A-E}). D) Generalized wave-dominated turbidite model from (Myrow et al., 2002) for comparison.

Shelf Ridges

Shelf ridges are generally thought to be relict sandstones reworked and reshaped by later currents (Snedden and Dalrymple, 1999). Shelf ridges have three main varieties: detached beach barrier bars, tidal shelf ridges, and storm shelf ridges. Whereas detached beach barrier bars can be considered part of the shoreface system (discussed below), both tidal and storm shelf ridges are similar in formation and internal structure, differing only in scale (Snedden and Dalrymple, 1999). Tidal shelf ridges are approximately 10-60 km in length, 5 to 40 m high and 0.7 km to 8 km wide (Snedden and Dalrymple, 1999). In contrast, storm shelf ridges are generally less than 15 km in length, average 7 m high and less than 8 km wide (Miall, 2000). The formation and growth of both tidal and storm shelf ridges can be described by the same model: 1) formation of an initial topographic irregularity oblique to the dominant flow that generates a hydrodynamic instability leading to deposition on the lee-side of the topographic irregularity (Huthnance, 1982), 2) if a sufficient supply of sand is available, the shelf ridge will grow, and the ridge now becomes the topographic irregularity, 3) with continued current activity, the ridge will grow to maximum size and eventually migrate in the direction of the current (Figure 13). Both types of shelf ridges commonly occur on transgressive surfaces where topographic irregularities (from ravinement) and available sand are common. In well-logs, shelf sand ridges will be sharp-based and blocky in appearance. Tidal shelf ridges form in areas of the shelf where strong tidal currents exist either in an area of restricted topography, such as the English Channel, or near an estuarine funnel (Snedden and Dalrymple, 1999). Storm shelf ridges can form in any area where large storms are common.

We suggest that many of the sandstone lenses (or lentils described in older papers) can be attributed to shelf sand ridges. The Upper Rushford Member, the sandstone packets in the Machias, Wellsville and Whitesville formations may all be shelf sand ridges, as would the basinward extent of shoreface sandstone in the Lower Rushford Member, Cuba and Hinsdale formation.

Beach and Shoreface

Beaches and shoreface systems in the Late Devonian Catskill Sea were thought to be non-existent, with only a transition from non-marine, muddy tidal systems to offshore muddy systems (Walker and Harms, 1971). Shoreface sequences are formed in wave-dominated systems and can be found as beaches attached to the land; as attached or detached barrier bar systems separated from the land by a shallow lagoon; or as cheniers where isolated beach ridges occur within the coastal mudflats (Elliott, 1986).

Our past work (Smith and Jacobi, 1998, 2000, and 2001) has shown that that the Lower Member of the Rushford Formation is comprised of three, stacked sandy shoreface sequences. Each shoreface sequence containing identifiable lower, middle, upper and foreshore zones (from lower shoreface ripples to trough cross sets to foreshore seaward-dipping planar laminae, Figures 10, 14). Organic-rich shales containing abundant examples of the brackish-water trace-fossil *Teichichnus* (Figure 14) within the Lower Rushford Member and overlying Intermediate Rushford Member indicate the presence of lagoonal or bay facies associated with the shoreface sequences, suggesting that in the northern part of the study area, the Lower Rushford Member shorefaces are part of a barrier bar system. In the Cuba and Hinsdale formations, lower to middle shoreface zones have been identified in outcrop. In well-logs the coarsening upward gamma ray curves are typical for normal (not subsequently modified) shorefaces. By default, beaches are parallel to the shoreline, although beach barrier bars can become more oblique as they become farther from land and transition to shelf sand ridges.

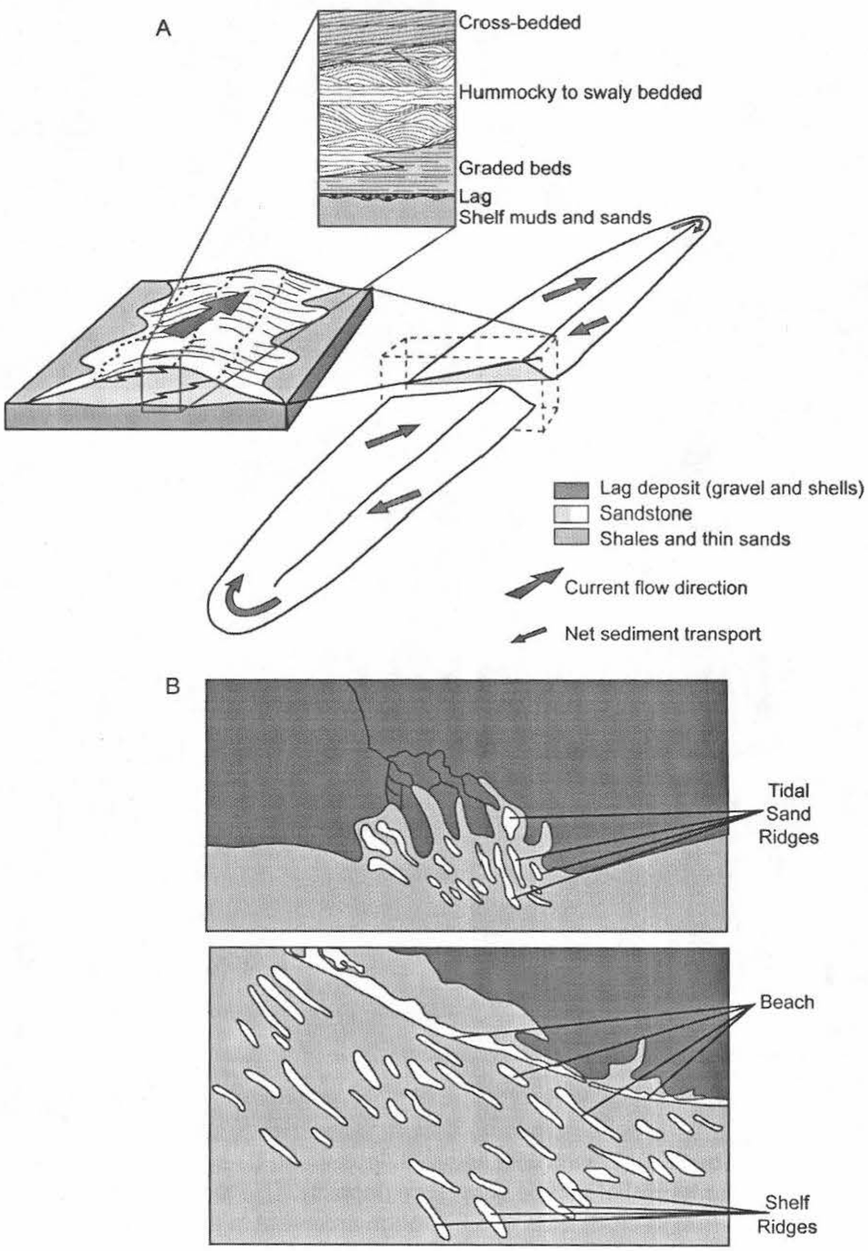


Figure 13. A) Anatomy of a shelf sand ridge incorporating models from Swift et al., 1986 and Dalrymple, 1992. B) Orientation of sand-ridges compared to shorelines, tidal sand ridges (top) are typically orientated perpendicular to shore. Beach barrier bars (bottom) trend parallel to shore, whereas shelf ridges (bottom) become more oblique offshore.

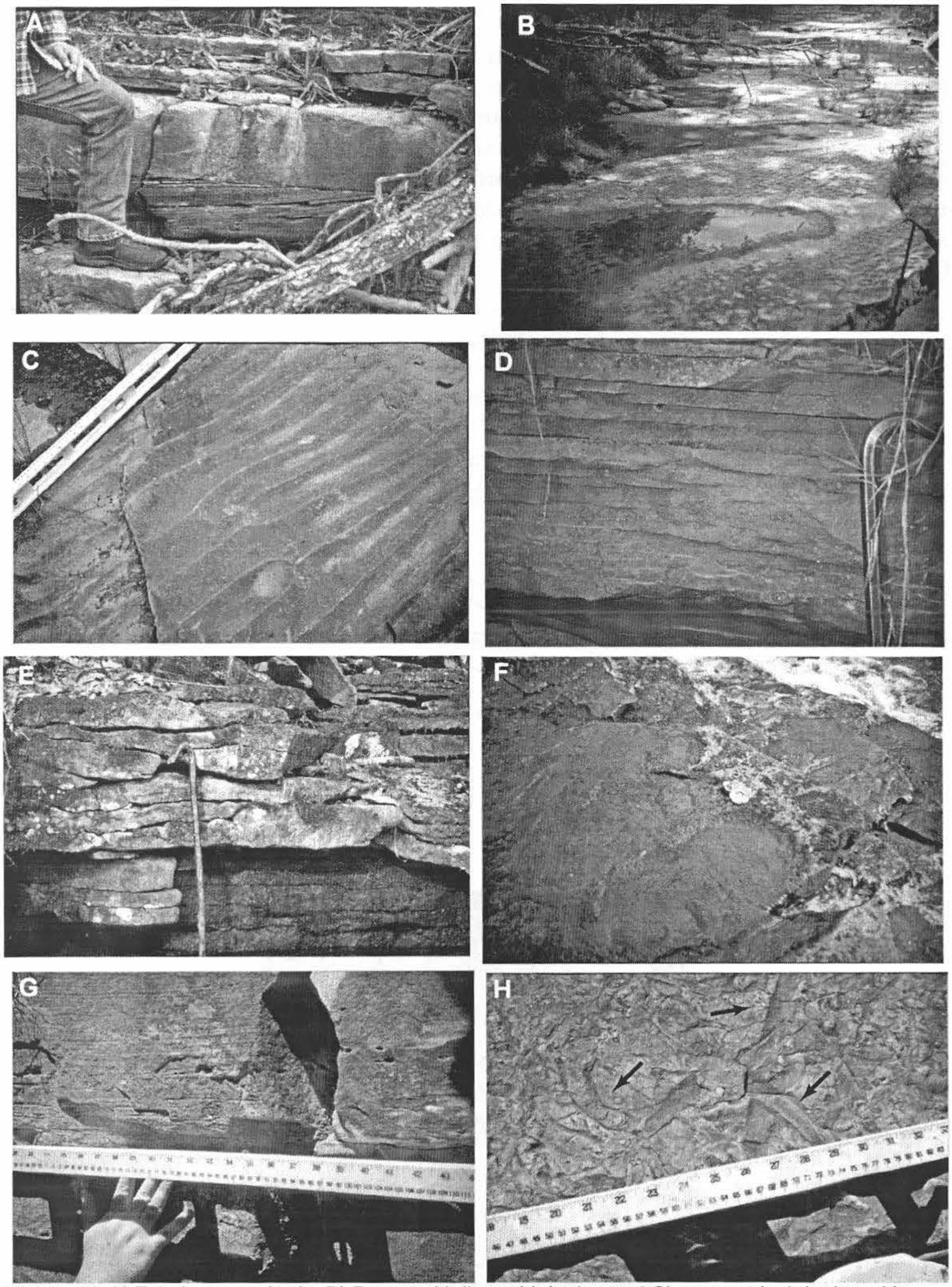


Figure 14. A) Tabular cross-beds, B) Dunes with linguoid ripples and C) symmetrical ripples with "tuning fork" and ladderbacks typical of middle shoreface deposits. D) Planer-bedded sandstones, E) Planer beds transitioning to trough cross-sets and F) trough cross-set in plan view typical of upper shoreface deposits. G) Planar lamiated sandstone with sporadic quartz pebbles indicative of foreshore deposits. Photos A-G taken from the Lower Rushord Member. H) Teichinchus (arrows) in thin sandstones in the Intermediate Rushford Member.

Fluvial

Fluvial systems are found within the non-marine red-beds of the Cattaraugus Formation and more commonly in Pennsylvania closer to the source area. Common fluvial system sandstone beds within the study area are small tidal channel deposits that are light gray, steeply cross-bedded sandstone that is 1-2 meters in thickness and laterally limited to a few tens of meters. The tidal channel sandstones are generally heavily burrowed with *Skolithos*, *Arenicolites* and *Ophiomorpha* common to high-energy, shallow or tidal environments.

Less common, but more prominent, are the thick conglomerates that occur at the base of the Conewango Group. The Salamanca, Panama, Pope Hollow, Killbuck, Tunangwant and Wolf Creek conglomerates are thick orthoconglomerate deposits that range in thickness from 2 to 12 meters. These conglomerate deposits do not appear to be laterally continuous and have been previously interpreted to be non-contemporaneous (Tesmer, 1975). The problem with correlating the conglomerates is that outcrops are rare, although the large blocks of conglomerate can be easily found, but rarely in place. The sporadic occurrences across Chautauqua, Cattaraugus and Allegany counties produce exposures at varying elevations that would lead to the assumption that the conglomerates are different units. However, elevation differences between widely separated localities can also result from 1) different erosional depth for lowstand incision, and 2) post depositional faulting.

Tesmer (1975) interpreted the conglomerates with flat pebbles to represent a marine environment, whereas he suggested the spherical pebble conglomerates represent fluvial deposits. We interpret the conglomerates to represent incised-valley fill, where the fluvial system intensely erodes in response to a lowstand, and generates clastic sediment through erosion as well as transporting fluvial gravels farther basinward. During the ensuing transgression, the eroded, or incised, river valley is inundated by the rising sealevel forming an estuary. Tidal, fluvial and wave components rework and redeposit the lowstand sediments into to coarse-grained estuarine deposits. Components of an incised valley system include the "tripartite" systems: marine (barrier bar at the mouth of the estuary); mixed marine-fluvial and tidal (estuarine mud and tidal bars), and fluvial (bay-head delta, fluvial bars and overbank deposits) (Figure 15). In wave-dominated systems, the tripartite system is typically: sandy marine barrier, muddy estuarine lagoon, and sandy bay-head fluvial delta (Dalrymple, 1992). It is important to note the difference between estuarine tidal bars and tidal shelf ridges. Both can, and have been, referred to as tidal bars (e.g. Willis, 2005), but differ in size, orientation and formation. Shelf tidal ridges (as previously discussed) form in deeper water, oblique to the current and are limited in size by the strength of the current, supply of sediment, and depth of the water. Estuarine bars form parallel to the tidal current, and are limited in height by the depth of low tide (Willis, 2005). Estuarine sand bars are typically separated by ebb-tidal channels, which force the bar to parallel flow. Incised valley systems and tidal channels will generally by orientated perpendicular to shoreline.

To summarize, in the absence of additional controls (structural), sand packets will be orientated: perpendicular to the shoreline (turbidite fans, estuarine and fluvial deposits), parallel to the shoreline (beaches, beach barrier bars), or oblique to the shoreline (shelf sand-ridges) (Figure 16).

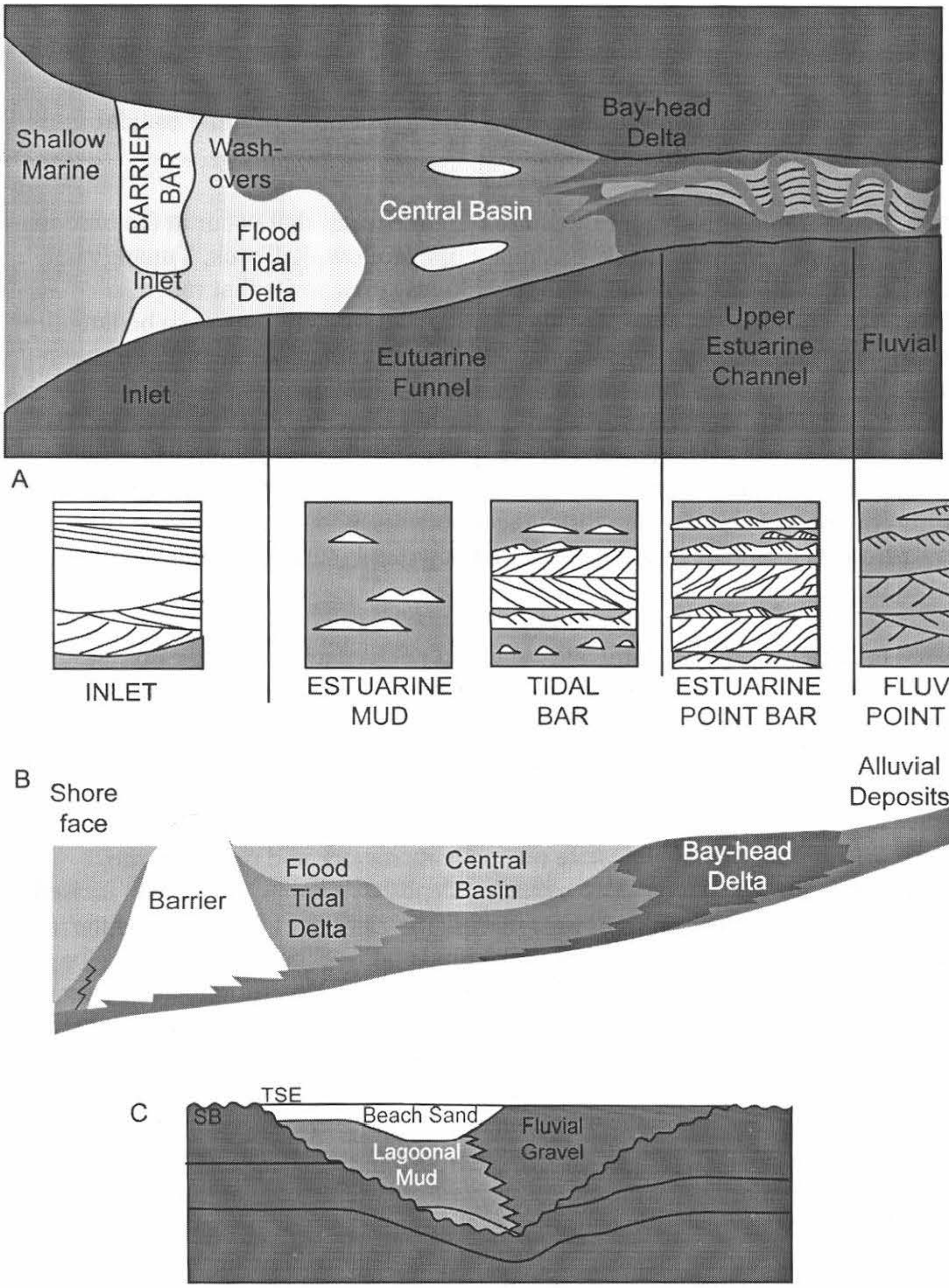


Figure 15. A) Estuarine deposition showing the "tripartite" components in a wave-dominated environment. B) Cross-sectional view of the estuarine system, parallel to the valley trend. C) Cross-sectional distribution of sediments, perpendicular to the valley trend. All modified from Dalrymple et. al., 1992 and Reinson, 1992.

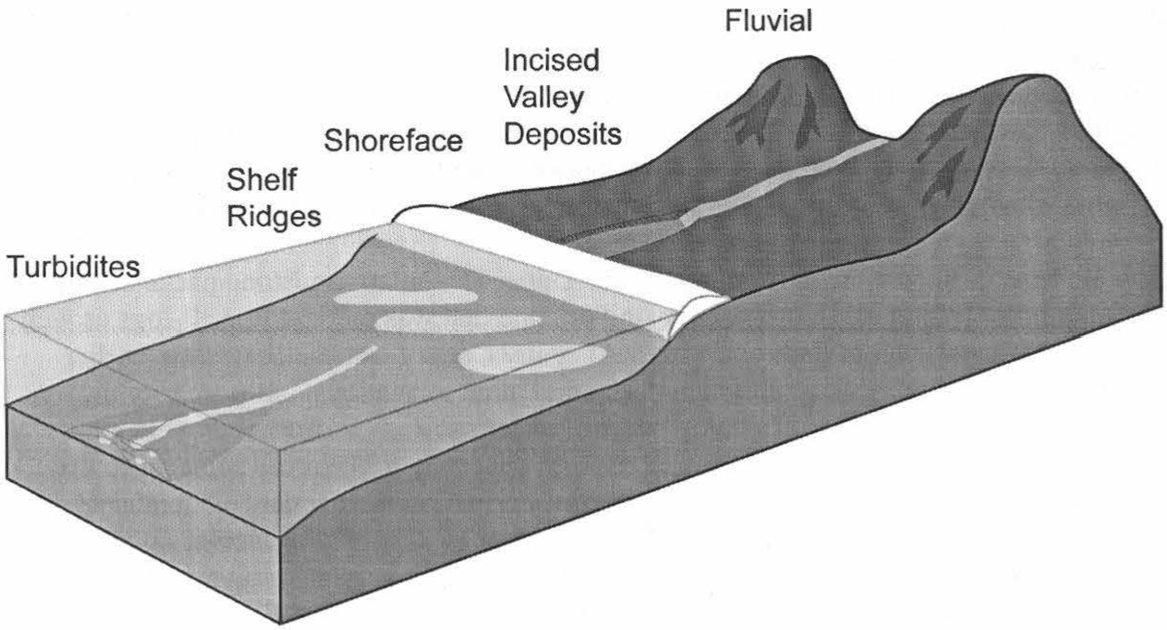


Figure 16. Summery of sandstone and coarser grained clastic depositional systems and the typical orientations to shoreline

Correlations

To correlate the Upper Devonian sandstone packets over the entire study area it was necessary to utilize well-logs to trace units observed in outcrop into the subsurface. Numerous well-logs were available for wells within the study; approximately 2,000 wells had logs on file or online with the New York State Museum. However, this number is far less than the 23,000+ wells the New York State Department of Environmental Conservation has listed in their well database. The discrepancy between the number of wells and the number of logs reflects the age of the oil and gas fields within the area; many of the wells predate electronic logging and more importantly predate state regulations requiring the submission of well logs.

Of the well-logs available, we focused on gamma-ray curves to provide stratigraphic correlations between wells, and between wells and outcrop. In many of the well-logs, only the zones of interest were logged, or logged up to casing which in turn removed much of the Upper Devonian section for most, if not all, of the Upper Devonian stratigraphic sections. Despite these limitations, we examined 418 well-logs that contained recognizable stratigraphic units and/or units that we might be able to correlate to adjacent wells or outcrop. For the purposes of this study we examined only the Devonian stratigraphic section, using the Onondaga Formation as our lower stratigraphic limit. Picked formations included in ascending stratigraphic order: Onondaga Formation, Cherry Valley Limestone, Marcellus Formation, Centerfield Limestone, Tichenor Limestone, Tully Limestone, Lodi Limestone, Geneseo Formation, Middlesex Formation, Rhinestreet, Nunda Formation, Pipe Creek Formation, Wiscoy Formation, Dunkirk Formation, South Wales Formation, Mills-Mills Formation, Hume Formation, Higgins

Member of the Caneadea Formation, Lower Rushford Member, Upper Rushford Member, Machias 1st Sandstone Packet, Machias 2nd Sandstone Packet, Machias 3rd Sandstone Packet, Machias 4th Sandstone Packet, Cuba Formation and the Hinsdale Formation. In effort to save time, formation tops for the Nunda Formation through the Higgins Member stratigraphic section were disregarded outside of areas adjacent to cross-sections as the subtle curve responses between the fine-grained sandstones, from the sandy shales and the silty black shales made formation picks ambiguous.

Crucial to our study are the correlations of the Upper Devonian sandstone packets, and most important was to identify a marker unit or zone that was evident in well-logs as well as outcrop. Correlation of outcrops within Allegany County was facilitated by marker units such as the black shales of the Dunkirk and Hume formations, but primarily the Upper Devonian sequence stratigraphy was of greater use for a wider geographic area, since exposures of the black shales were confined generally to the areas adjacent to the Genesee River Valley. The base of the Rushford Formation is a sequence boundary with lowstand shoreface deposits overlying offshore sediments (Smith and Jacobi, 2001 and 2003); the top of the Lower Rushford Member is marked by a transgressive surface of erosion, overlain by deeper-water deposits (Smith and Jacobi, 1996, 1998, 1999, 2001, 2002, 2003 and 2004). The three stacked, coarsening-upwards, shoreface sequences could be trace from outcrop to outcrop from the Fillmore quadrangle in Allegany County to the Ellicottville quadrangle in Cattaraugus County. By comparing these identified outcrops with adjacent well-logs in the northern section of the study area, we were able to identify a sequence with a coarsening-upward sequence at the base (typically with both a sharp base and top), overlain by a shale (which represents the Lower Rushford Member). A sequence representing the Lower Rushford – Upper Rushford – Machias 1st sequence was identified for wells in the major oil fields (Figure 17). Cross-section A-A' (Figure 18) is a north-south cross-section that incorporates both outcrop and well-logs. Outcrops of the Rushford Formation correlate with adjacent well-logs, and the resultant dips in the correlated Rushford match dips from the Onondaga and Tully. Two east-west cross-sections were made for this study to trace changes in stratigraphy as units were traced westward, deeper into the basin. The northern east-west cross-section B-B' (Figure 19) and the southern east-west cross-section C-C' (Figure 20) show an overall thinning of the stratigraphic section toward the west; the thinning appears more pronounced in the southern cross-section).

Significant lateral variations in shale thickness between outcrops and between well-logs indicate either syndepositional faulting along local fault basins (e.g., along the Clarendon-Linden Fault System, e.g., Jacobi and Fountain, 1996; Smith and Jacobi, 2002), or in some areas, tectonic thinning or thickening of the section. Such tectonic effects were expected from outcrops in Allegany and Cattaraugus counties that displayed pencil cleavage, bedding thrusts, disrupted bedding, and in one outcrop, a recumbent fold (Jacobi and Fountain, 1996; Peters, 1998; Zack 1998; Smith, 2002). These structural features are thought to represent post-depositional (Alleghanian Orogeny) crustal shorting (e.g., Engelder and Geiser, 1979) that could produce the apparent black shale thinning or especially thickening along duplexes within the interbedded sequences.

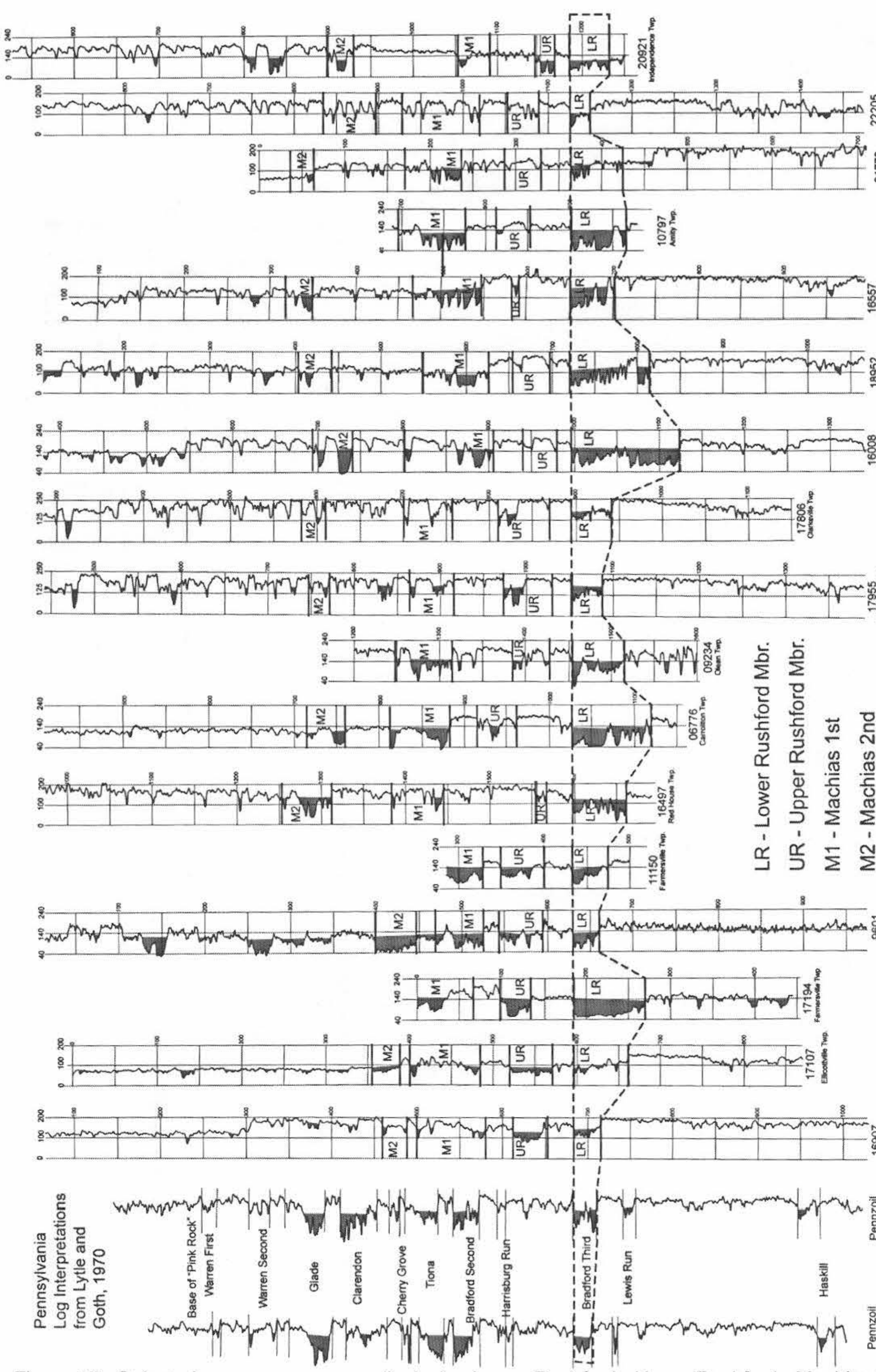


Figure 17. Selected gamma ray curves displaying Lower Rushford - Upper Rushford - Machias 1st sequence for selected wells ; Pennsylvania well-logs from eastern Warren County, are used for comparison of sands and common sandstone names.

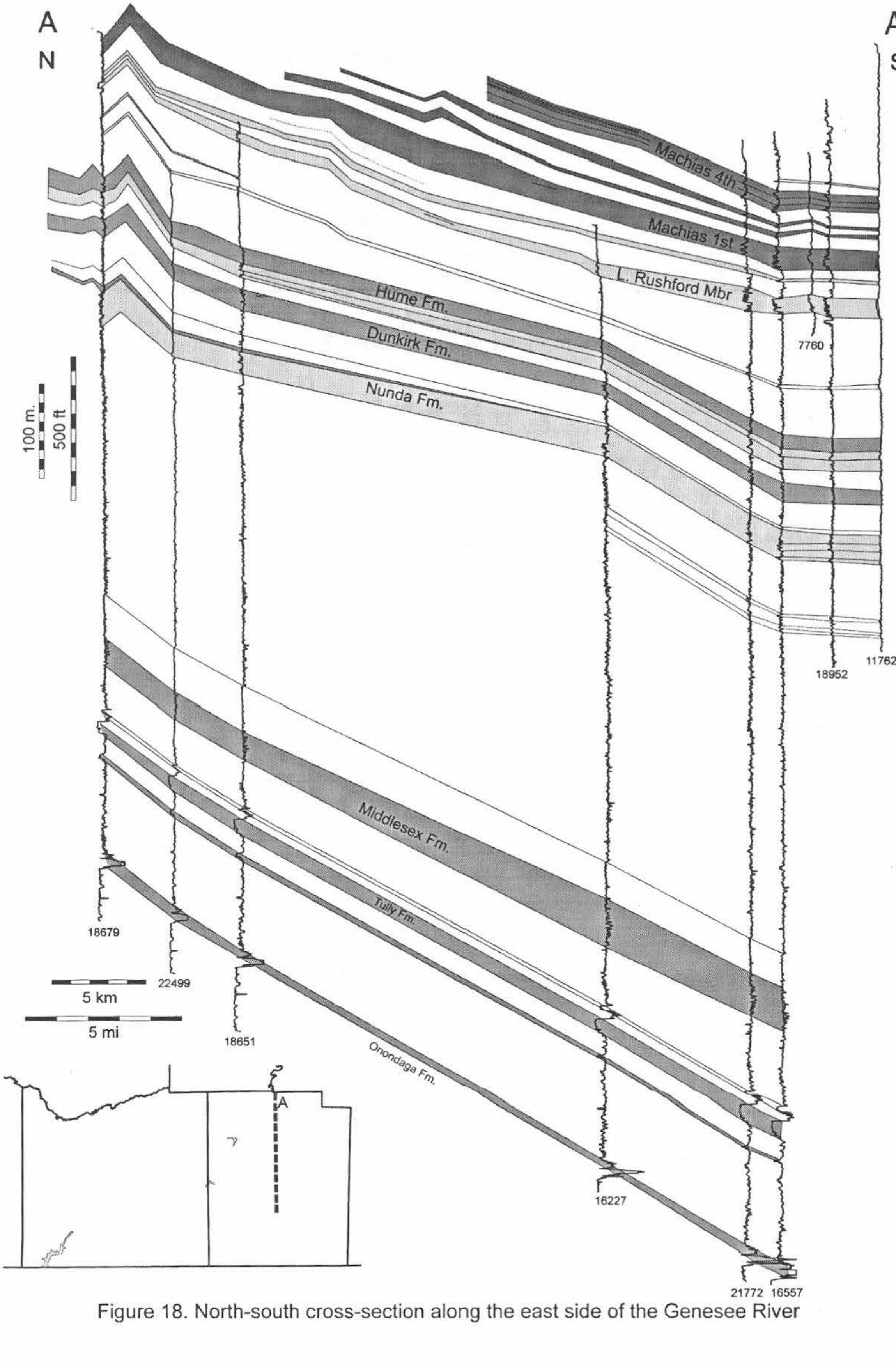


Figure 18. North-south cross-section along the east side of the Genesee River

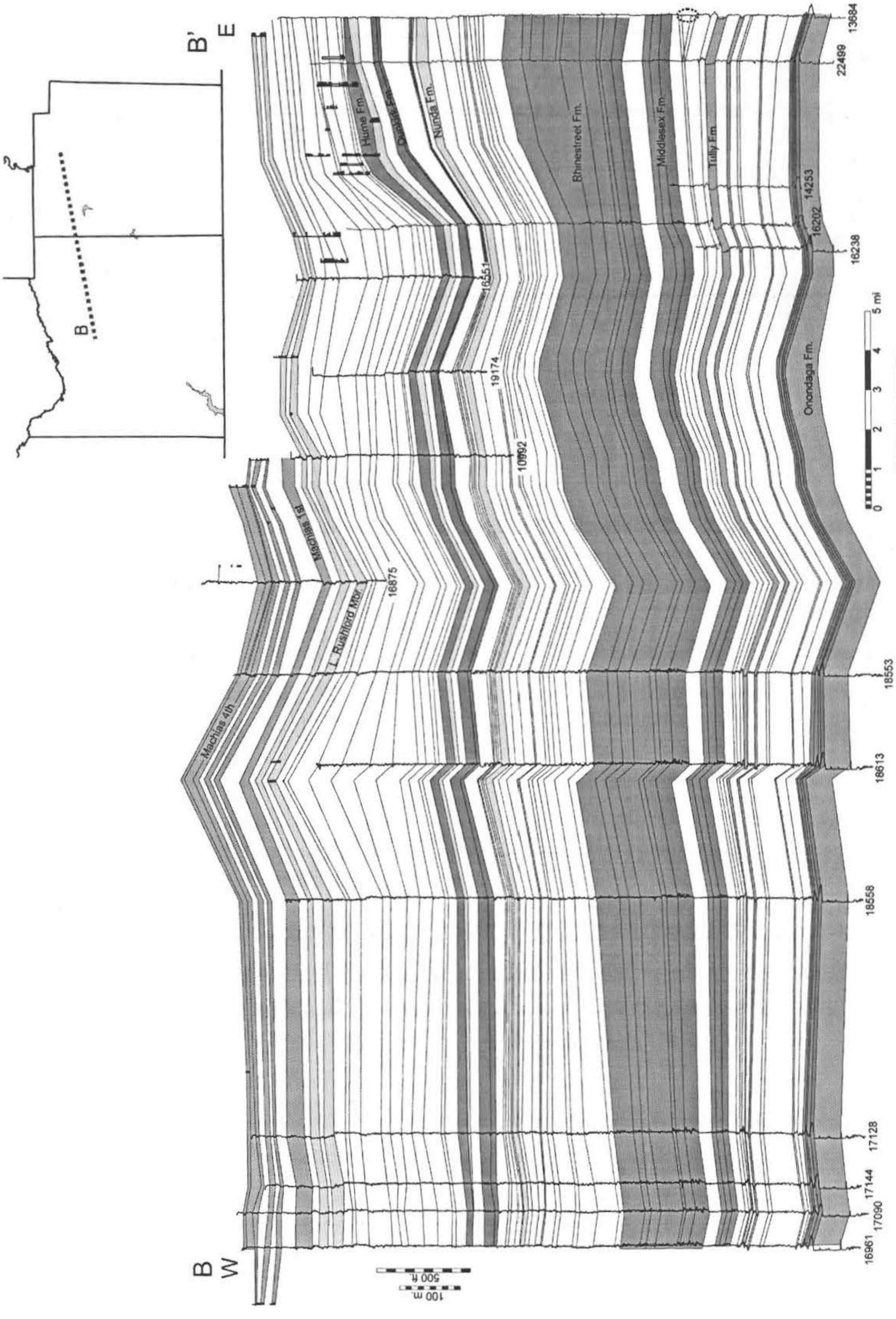


Figure 19. Northern east-west cross-section with ground elevation for datum. Repeated section circled in dashed circle.

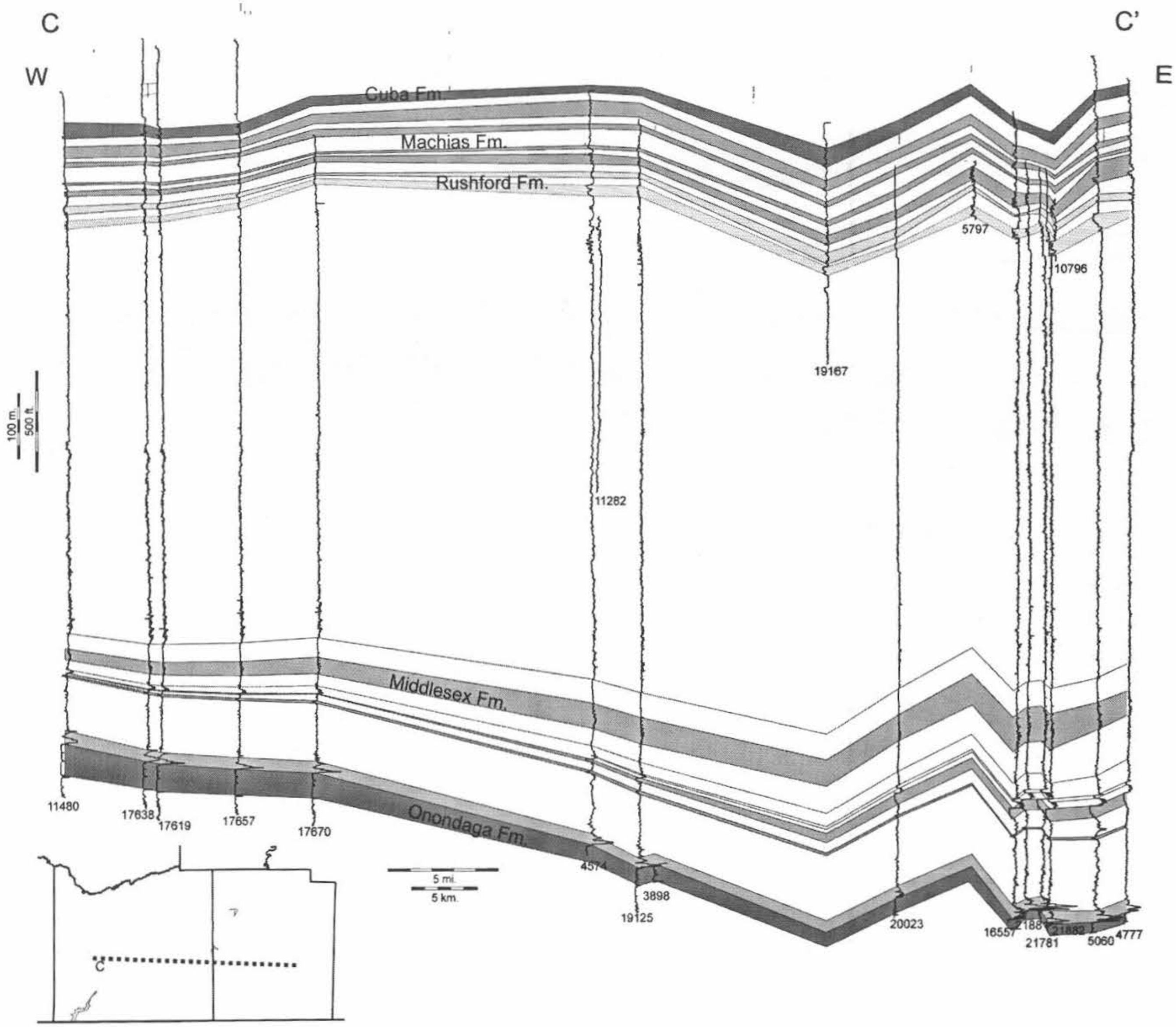


Figure 20. Southern east-west cross-section

OUTCROP EXAMPLES OBSERVED ON THE FIELDTRIP

Our fieldtrip will examine outcrop examples of the depositional environments that represent the major clastic reservoirs in the Upper Devonian that correlate with the Elk, Bradford and Venango reservoir sands to the south. Additionally, the fieldtrip shows examples of syndepositional faulting and the faulting influence/control on depositional environments. To cover as much ground as possible in the time allotted, we have chosen outcrops near roads, although these outcrops do not necessarily represent the best outcrop possible. Several stops are located on state property, or along roadcuts, and do not require obtaining landowners permission (although entry to Letchworth State Park will require an admission during certain periods of the year). Two stops are located on private property and will require obtaining permission from the landowner prior to visiting. Most landowners we have encountered are reasonable and offer no objection, unless they regularly discover trespassers or signs of trespasser. Please respect the wishes of the landowner.

Directions	Distance	Cumulative
START - LEAVE ADAMS MARK	0	0
Head west on Church St toward Bingham St.	0.1mi	0.1
Merge onto I-190S/NYS Thruway via the ramp on the LEFT towards RT-5 West/Skyway	5.1 mi	5.2
Merge onto I-90 W/NYS Thruway W via EXIT 54-61 toward ERIE	1.6 mi	6.8
Take the RT-400/ RT-16 exit - EXIT 54 - toward WEST SENECA/EAST AURORA	0.5 mi	7.3
Merge onto NY-400 S / AURORA EXPY	12.4 mi	19.7
Take the US-20A ramp toward EAST AURORA, turn LEFT (West) onto US 20A	0.2 mi	19.9
Turn RIGHT onto RT-78 S	2.1 mi	22
Strykersville - Turn LEFT onto PERRY RD/ CR-9	7.3 mi	29.3
North Java - Continue straight on WETHERSFIELD RD/ CR-32	6 mi	35.3
Wethersfield Spring - Continue straight on WETHERSFIELD RD	7.2 mi	42.5
Turn RIGHT on NY-19 South	3.4 mi	45.9
Turn LEFT onto NY-19A	2 mi	47.9
Turn SLIGHT LEFT to stay on NY-19A	2 mi	49.9
Turn LEFT onto DENTON CORNERS RD/ CR-38	0.2 mi	50.1
Bear RIGHT enter LETCHWORTH STATE PARK- Castile Entrance.	<0.1 mi	50.2
Past the Park Entrance, turn RIGHT and head toward the Middle Falls Parking Lot.	2.7 mi	52.9
STOP 1 - Letchworth Upper Falls Trail		

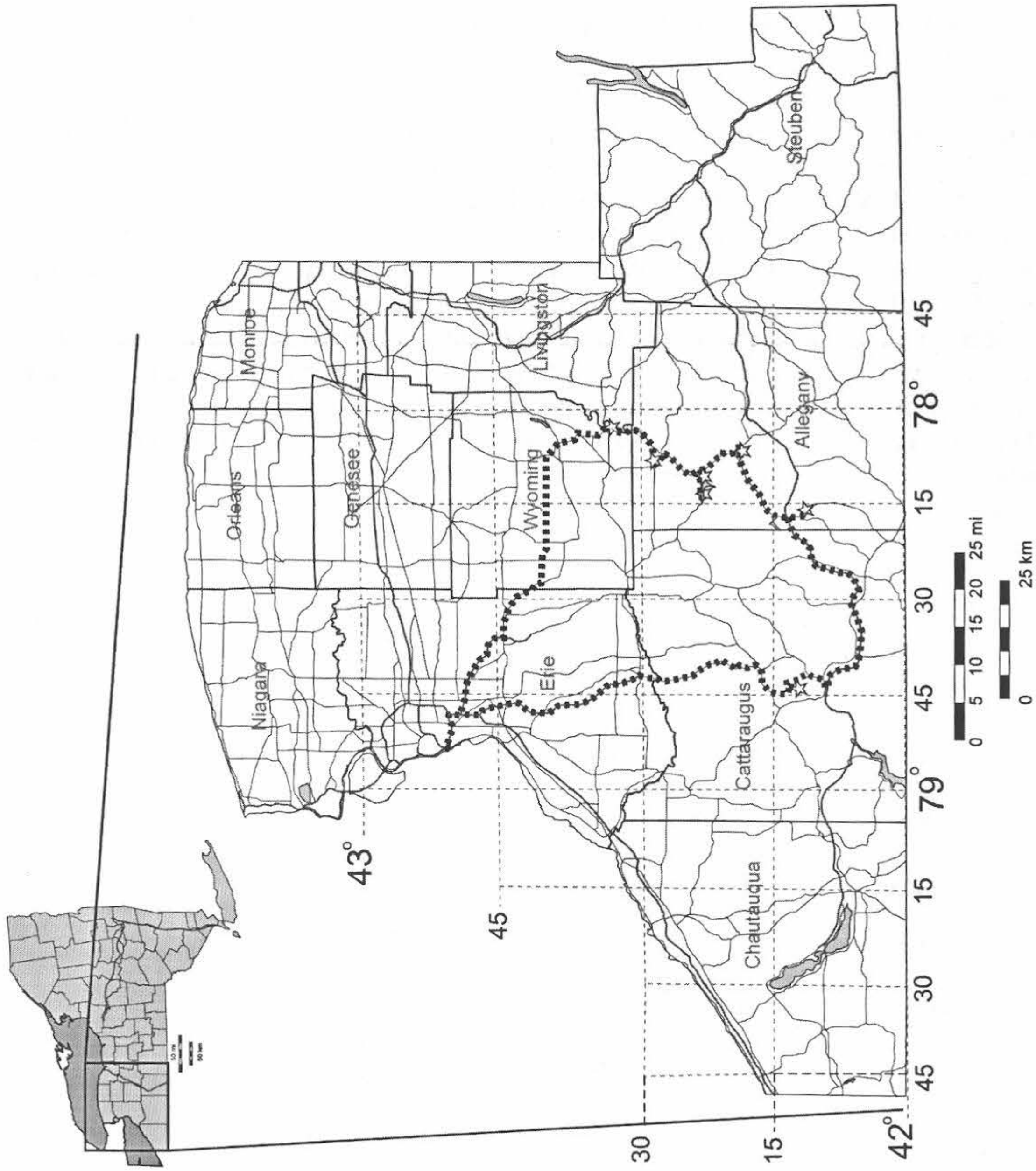


Figure 21. Overview of the route for the fieldtrip. Stops are indicated by stars (☆).

STOP 1 – Letchworth State Park – Upper Falls Trail (Figures 22, 23).
Location Coordinates (WGS 1984):
Lat – N42.57889°; Lng – W78.04927°
UTM Zone 17; N4718278, E0742154

Letchworth State Park contains a remarkable exposure of the Frasnian stratigraphic section along the 400 foot high cliffs and waterfalls carved by the Genesee River. Three major waterfalls are viewable in the park, the Lower, Middle and Upper Falls. Whereas much of the cliff sections are comprised of Gardeau Fm. shales and silty shales interbedded with thin turbidite/storm sandstones, the stratigraphic section above the

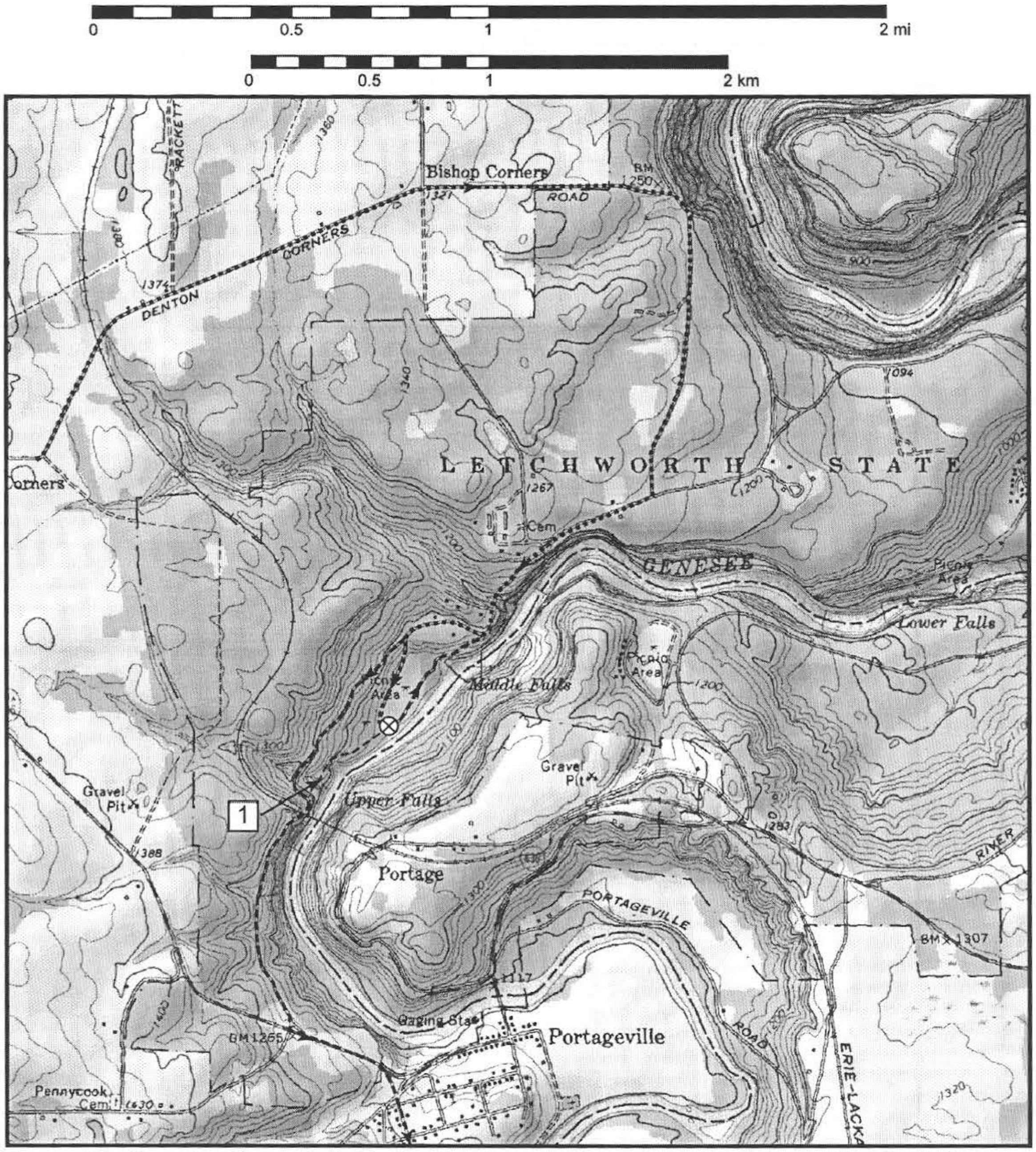
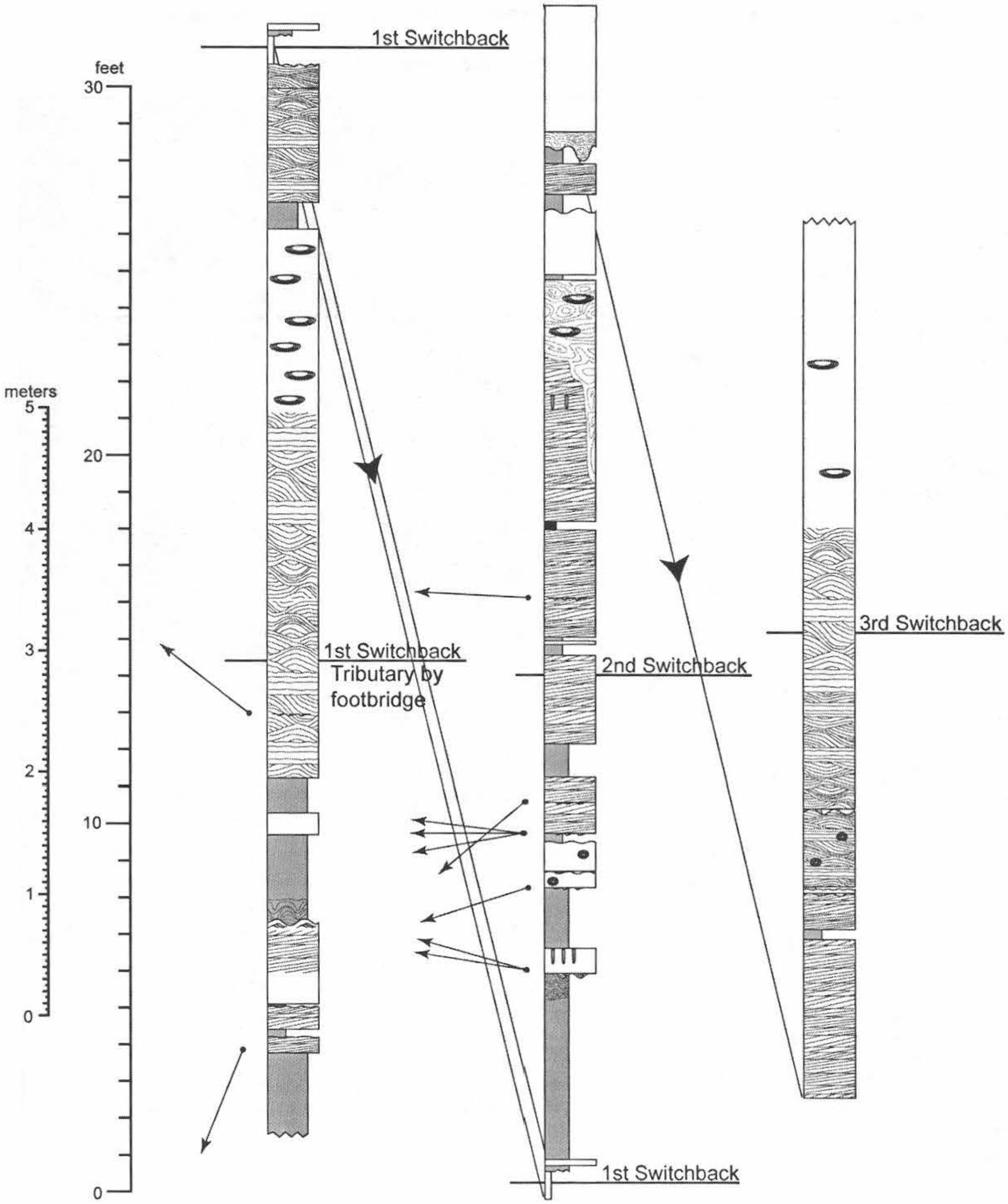


Figure 22. Stop 1 - Upper Falls Trail at Letchworth State Park. Park in the Middle Falls parking lot, and take the Upper Falls Trail south and west; accessible outcrop starts at the footbridge crossing the southeast flowing tributary to the Genesee River (1).

Upper Falls exposes outcrops of the Nunda Formation (Clarke, 1897; Figure 23). The Nunda Formation is correlative to Elk reservoir sands, and is a good example of a thick turbidite sandstone T_{AB}; however dewatering structures and wave-modification provide more variability than the original Bouma model (See figure 12).



STOP 1 - Letworth State Park Upper Falls Trail - Nunda Fm.

Figure 23. Measured annotated stratigraphic column for the Upper Falls Trail. Arrows represent measured paleoflow directions, key to the symbols may be found on Figure 10.

The outcrop we will examine is exposed along the trail toward the Upper Falls and the High Bridge. The path and stairs were cut through 24 m of stratigraphic section, mostly fine-grained sandstone. In addition to examples of primary bedforms and sedimentary structures typical of turbidites, soft-sediment deformation and liquefaction features including sandstone dikes, roll ups, load casts, and zones of liquefaction with sharp boundaries are quite common in both the sandstones and the interbedded shales. Small down-on-the-east stratigraphic offsets (~6 cm) along northwest-trending (317°) fractures typically coincide with observable deformation zones in the outcrop.

Bedding within some of the thicker sandstones appears similar to swaly-cross stratification, which suggests storm-wave modification/influence. Storm influence within the Nunda Formation is consistent with observations made higher in the stratigraphic section where "deep" water depositional environments (black shales and turbidites) show numerous bedforms (such as hummocky cross-stratification, 3-D ripples, linguoid ripples, e.g., Smith and Jacobi, 2002) that indicate depositional depth was within storm-wave base (Myrow et al., 2002; Pattison, 2005).

The specific depositional environment of the thick sandstones of the Nunda exposed at Letchworth is ambiguous. The thick nature of the sandstone beds, the striations on the base of the sandstone beds, the massive to planar (to possibly swaly bedding) all indicate proximal turbidites (T_{AB} to $T_{A\text{-}C}$). Such turbidites were typically are thought to represent submarine channel deposits. However, the lack of coarse basal sediments, and the constant thickness of the thick beds exposed along the High Cliffs at Inspiration Point suggest that either the channels were very wide, or that these represent a non-point source, more on the order of a storm-generated wash with a shoreface source. To the east we have found an erosional channel wall, indicating a channel (or slump scar) (Jacobi et al, 1994), and to the west, we suggested relatively thick Nunda sand beds represented sand lobe at the end of a channel (Jacobi et al, 1994), so it is possible that these sands at Letchworth do represent wide channels (or storm-generated wash that funneled into channels downslope).

Directions	Distance	Cumulative
STOP 1 - Letchworth Upper Falls Trail		
Exiting the Middle Falls Parking Lot, turn left heading south towards the Portageville Entrance. At the Portage Entrance Turn LEFT onto RT-486/RT-19A	0.8 mi	53.7
Keep RIGHT on RT-19A	0.7 mi	54.4
Rossberg - Turn RIGHT onto Wiscoy Rd/CR-27	5.2 mi	59.6
Wiscoy - Turn LEFT onto Wiscoy-MillsMills Rd	0.8 mi	60.4
Turn RIGHT at the stop sign at the western end of Wiscoy-MillsMills Rd park along right side of the road, outcrop is below the bridge crossing Wiscoy Creek.	1.9 mi	62.3
STOP 2 – Mills-Mills		

STOP 2 – Mills-Mills – Wiscoy Creek exposure (Figures 24, 25).
Location Coordinates (WGS 1984):
Lat – N42.50121°, Lng – W78.04927°
UTM Zone 17; N4718278, E0742154

Wiscoy Creek at Mill-Mills where Mills-Mills Rd. crosses Wiscoy Creek. This is the upstream end of a fairly continuous outcrop that extends down to the village of Wiscoy. Roadcuts on either side of the bridge contain weathered exposures of the Hume Fm. black shale (Pepper and deWitt, 1951). The basal contact of the Hume Fm. with the underlying Mills-Mills Fm. (Smith and Jacobi, 2000) occurs below the Mills-Mills dam. The sandstones near the contact have a distinct petroliferous odor.

The underlying Mills-Mills Fm. (Smith and Jacobi, 2000) is lowest, thick sandstone package in the Famennian stratigraphic section in Allegany County. The Mills-Mills Fm. can be separated into upper and lower sandstone packets. The lower sandstone packet is a thick (~3m), upward fining sequence of fine- to medium-grained sandstone interbedded with thin gray shales and siltstone. The lower sandstones have tabular cross-stratification and displays prominent climbing ripples. The lowermost sandstone is thick (20-40cm) and fine-grained sand; the sand bed has an erosional base and is channelized in cross-section. The upper sandstone packet ranges from 1 to 2 m thick and consists of thick beds (40 to 60 cm) of medium sandstones. Flute casts are observed on the soles of sand beds. Between the two sandstone packets is an interbedded section of thin gray sandstones, siltstones and shales that change upsection from gray to black. We have interpreted the Mills-Mills Fm. as offshore to restricted circulation deposits (Figure 12), with the sandstone packets deposited as T_{AB} to T_{AC} turbidites with T_C–starting turbidites in the upper packet. Many of the lower turbidites are consistent with proximal turbidites deposited in channels, whereas the T_C–starting turbidites appear more like lobe fringe sands.

The contact between the Mills-Mills Fm. and the underlying South Wales Fm. (Pepper and deWitt, 1951) is located downstream at the waterfall near the old RG&E powerhouse. The caprock of the waterfall is the basal thick fine-grained sandstone bed of the Mills-Mills Fm. The exposure of the Mills-Mills continues upstream with the sandstone beds forming small cascades up to the dam at Mills-Mills. The interbedded shales and siltstones grade from gray to black at the top of the formation.

Farther downstream, additional outcrops of the South Wales Fm. and Dunkirk Fm. (Clarke, 1903), are exposed along the southern bank, and in the north-flowing tributaries. Correlation of marker beds, such as the basal sandstones of the Mills-Mills and South Wales formations) indicate possible uplift along the fault east of STOP 2 (Figure 26).

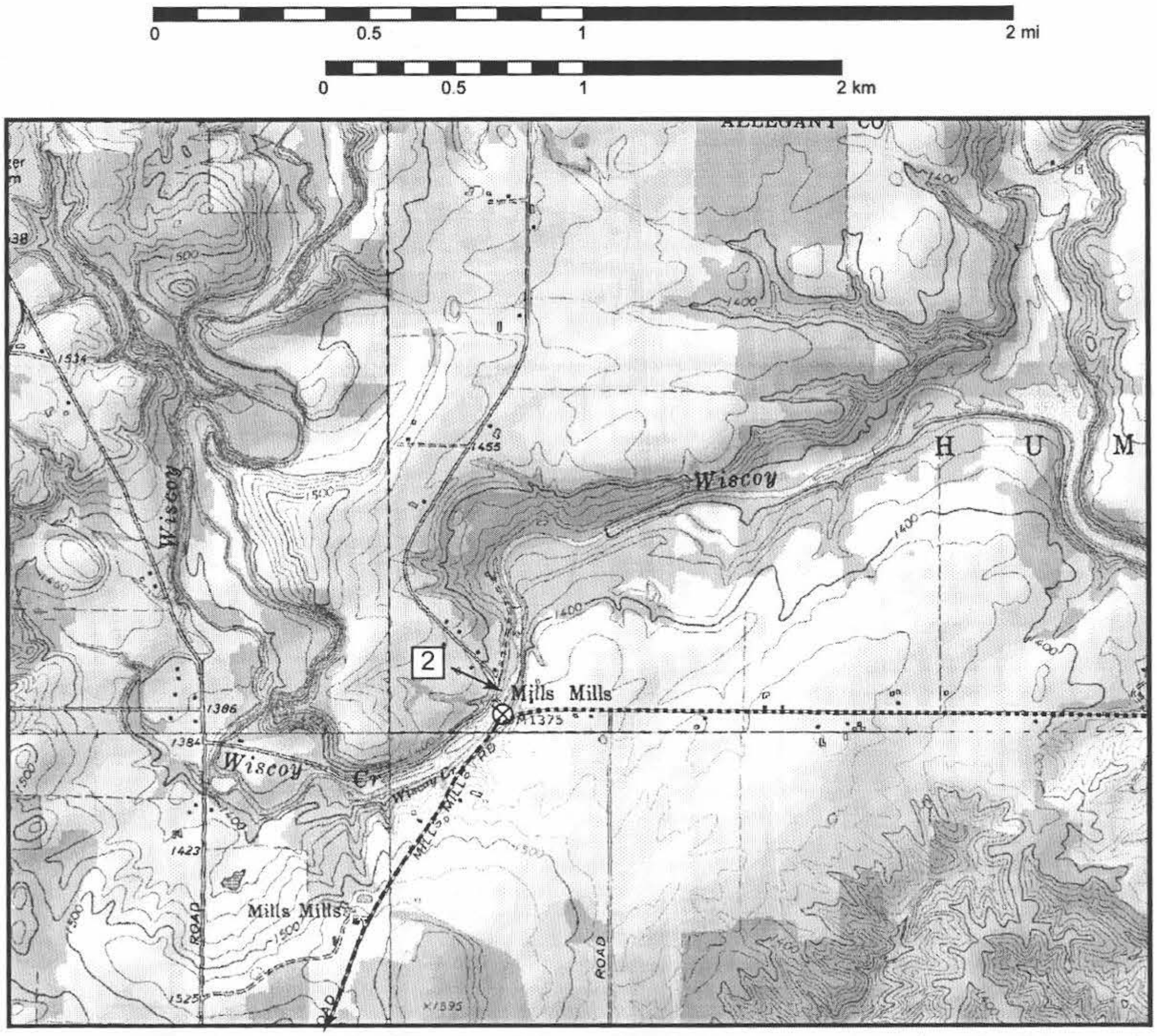
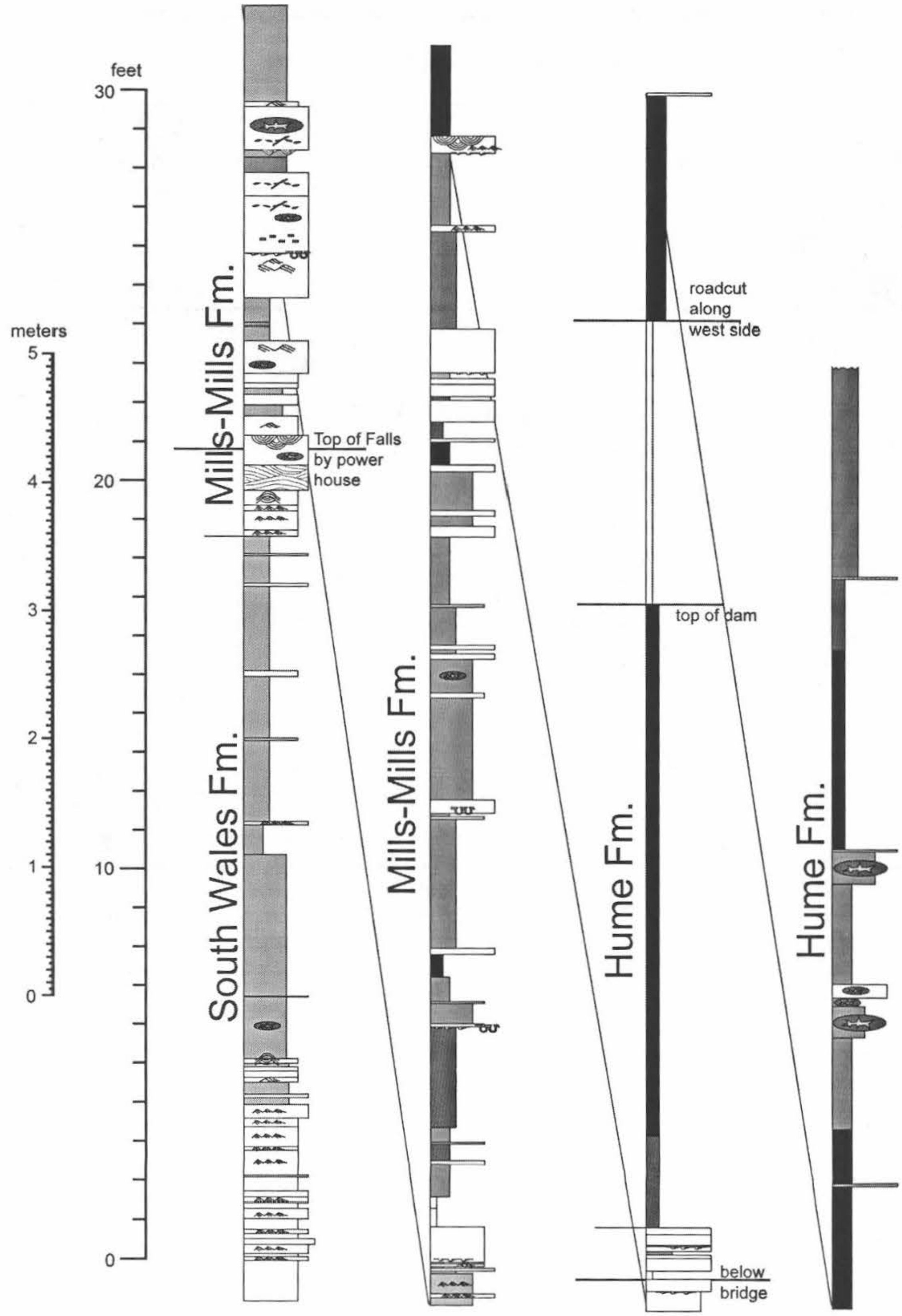


Figure 24. Stop 2 - Mills-Mills Falls along Wiscoy Creek. Park along the right side of the road, east of the bridge crossing Wiscoy Creek. The top of the Mills-Mills Fm. is below the bridge (2). Note: since this is private property, please contact the landowner prior to visiting and respect his wishes.



STOP 2 - Wiscoy Creek at Mills-Mills.
South Wales, Mills-Mills and Hume formations

Figure 25. Measured annotated stratigraphic column for the outcrop exposure along Wiscoy Creek at Mills-Mills. Key to annotations may be found on Figure 10.

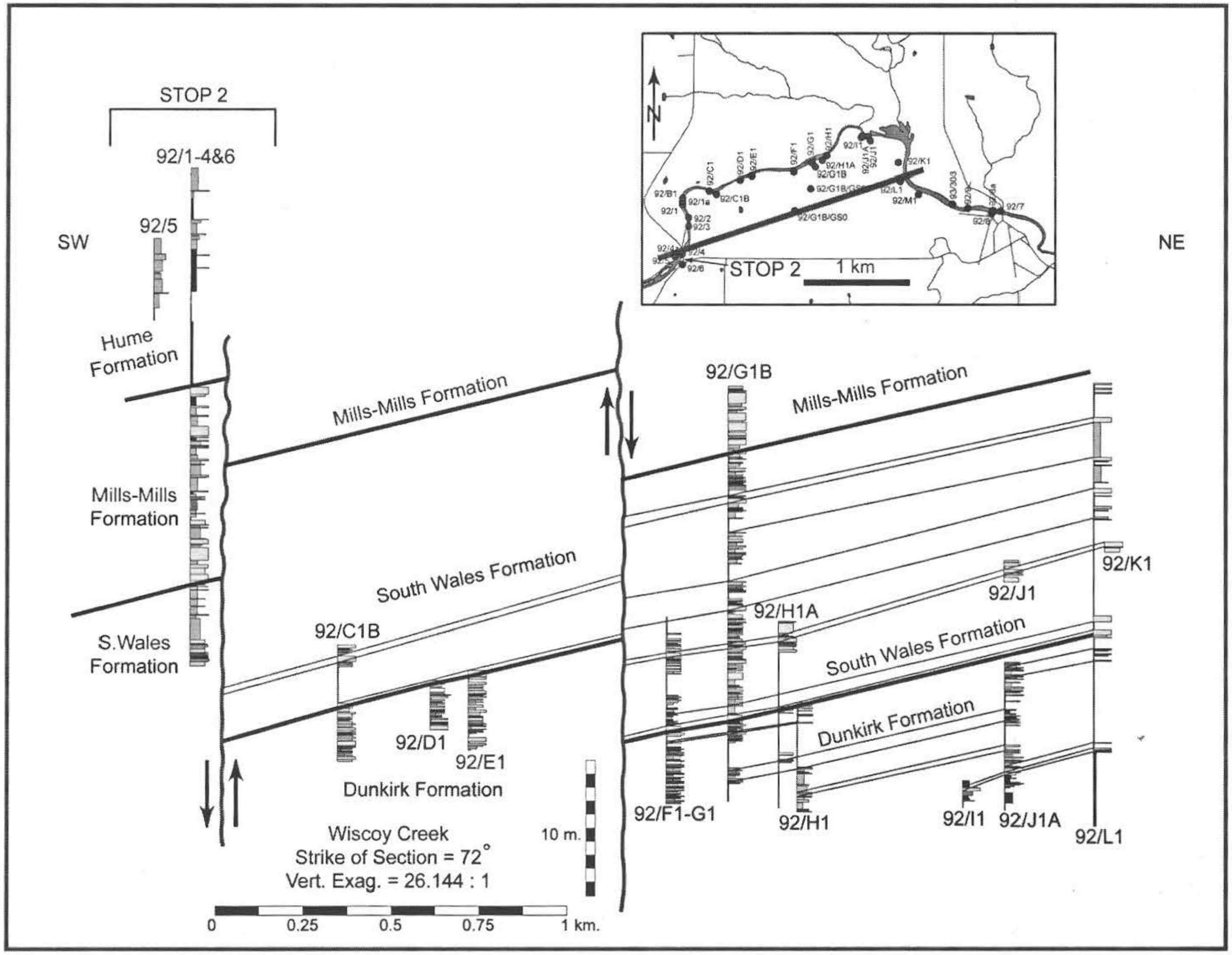


Figure 26. Cross-section of outcrops along Wiscoy Creek displaying down-dropped block based upon marker beds at the base of the South Wales Fm., and the distictive base of the Mills-Mills Fm.

Directions	Distance	Cumulative
STOP 2 – Mills-Mills		
From Stop 2, make a U-Turn and head west along Mills-Mills Rd. (Not Wiscoy-MillsMills Rd. taken before Stop 2).		
Continue STRAIGHT onto CR-23/Hume Rd.	1.9 mi	64.2
Turn LEFT onto SR-243	8.3 mi	72.5
Turn RIGHT onto CR-9/Hillcrest Rd.	0.2 mi	72.7
Outcrop for STOP 3 is on both the Left and Right sides of the road - park with care along the right since this is essentially a blind curve.	0.4 mi	73.1
STOP 3 - Hillcrest Rd.		

STOP 3 – Hillcrest Rd. Roadcut (Figures 27, 28)
Location Coordinates (WGS 1984):
Lat – N42.38221°, Lng – W78.22596°
UTM Zone 17; N4695944, E0728370

Thick laminated sandstone (> 3 meters thick) is exposed on either side of the road, where Hillcrest Rd. begins to climb the hill. The western side of the road exhibits the sharp contact between the Rushford Fm. (Luther, 1902) and the Caneadea Fm. (Chadwick, 1933), whereas the fracture controlled exposure on the eastern side gives an excellent view of the planar laminations and small, white, cloudy quartz pebbles incorporated into the fine-grained sandstone matrix.

The exposure of the Lower Rushford Mbr. here is interpreted as three stacked shorefaces. The upper shoreface (planar beds) and foreshore (planar laminations) facies of each shoreface is eroded by an overlying transgressive surface of erosion. The Lower Rushford Member is correlative to the Bradford 3rd reservoirs along the New York-Pennsylvania border (Fettke, 1939; Manspeizer 1963; Smith, 2006).

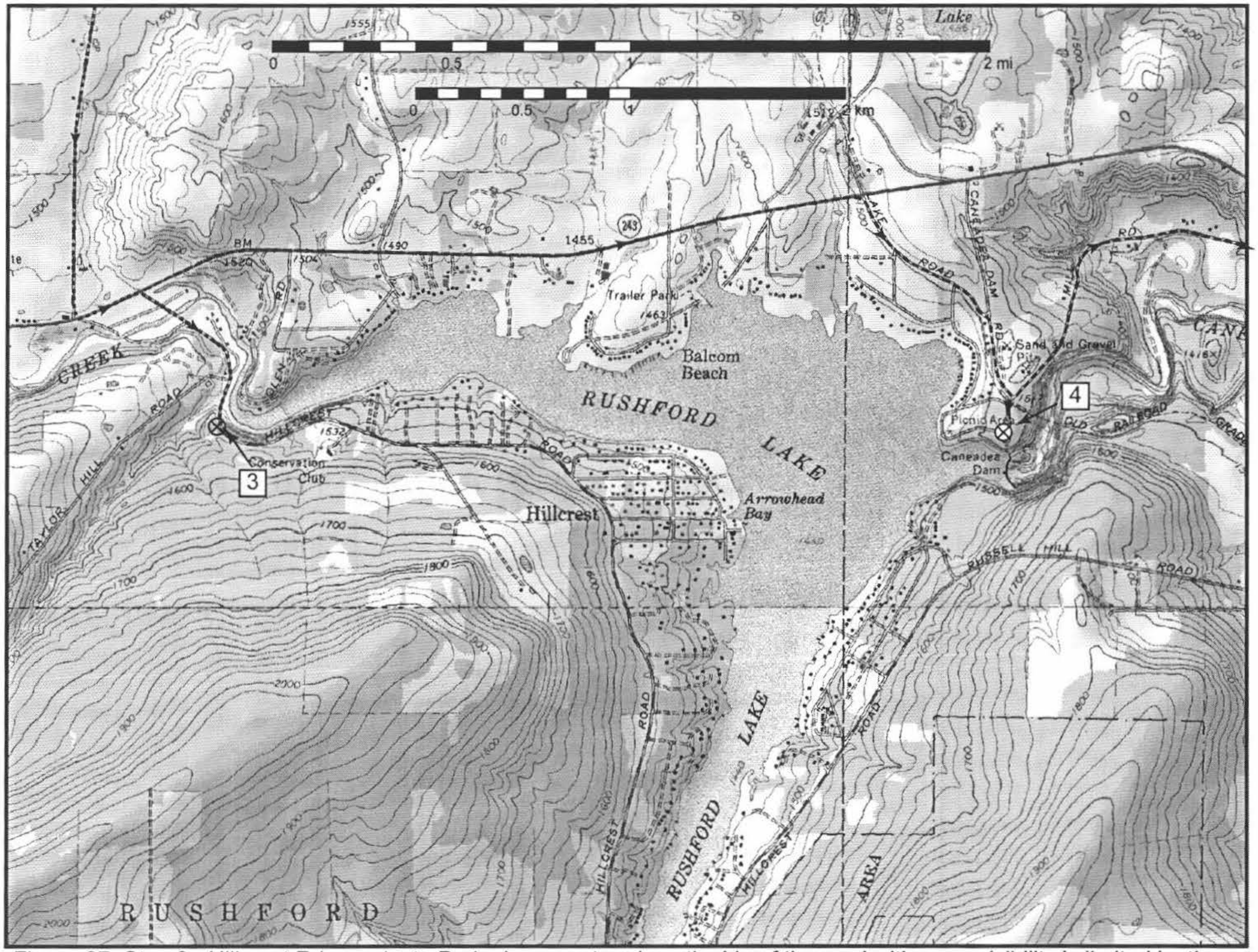
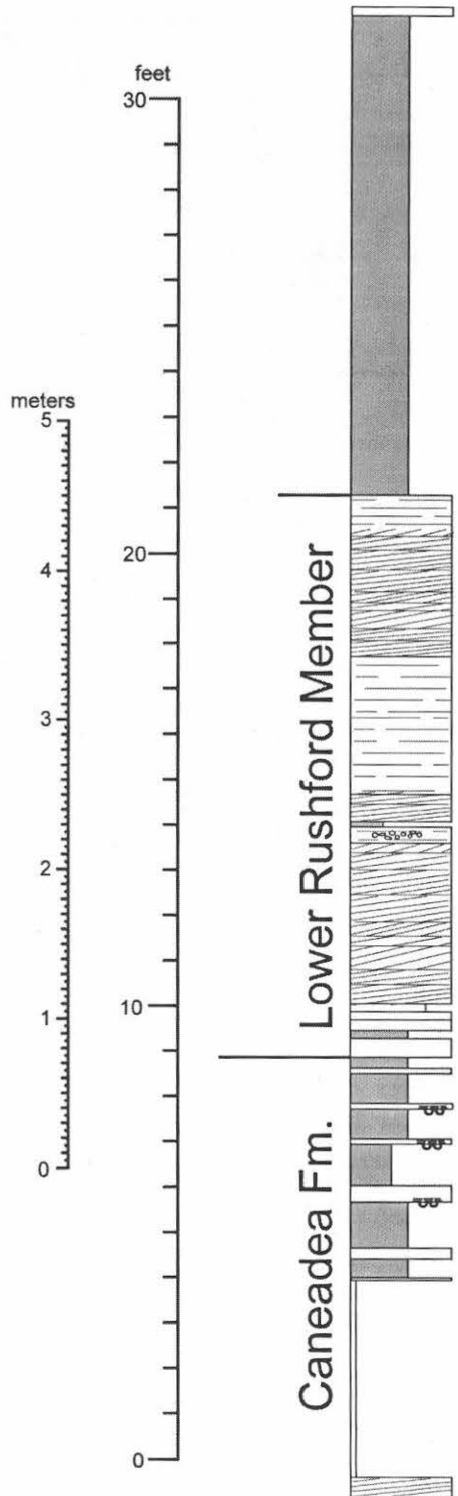


Figure 27. Stop 3 - Hillcrest Rd., roadcut. Park along west and south side of the road with care, visibility is limited by the curve of the road. Stop 4 - Lunch stop at Caneadea Dam Pinic Area.



STOP 3 - Hillcrest Rd. Rushford Fm.

Figure 28. Measured, annotated stratigraphic column for the Hillcrest Rd. outcrop. Key for the annotations may be found on Figure 10.

Directions	Distance	Cumulative
STOP 3 - Hillcrest Rd.		
Turn around and return to SR-243, Turn RIGHT onto SR-243	0.4 mi	73.5
Turn RIGHT onto Lake Rd.	2 mi	75.5
Turn RIGHT onto Caneadea Dam Rd.	0.5 mi	76
Continue STRAIGHT into Caneadea Dam Picnic Area	0.3 mi	76.3
STOP 4 - Lunch Stop		

STOP 4 – Caneadea Dam Picnic Area – Lunch
Location Coordinates (WGS 1984):
Lat – N42.38178°, Lng – W78.18311°
UTM Zone 17; N4696008, E0731902

Beyond the fence is Caneadea Dam and Rushford Lake. Old photographs of the construction of the Caneadea Dam in 1929 show the north wall of the gorge (the side we are on) devoid of trees, and show quarries of the Rushford sandstones. By the dam, and around the eastern end of Rushford Lake, is the type exposure of the Rushford Formation as defined by Luther (1902), although close, hands-on access to these outcrops can be difficult depending on the lake level. When the lake level is lowered in the Fall, along the south gorge wall of the dam can be seen a series of north-striking, east-dipping, closely-spaced step faults; each with minor offsets (on the order of a few centimeters; Jacobi and Fountain, 2002). These small faults are consistent with well-logs that indicate a down-on-the-west fault of the Clarendon-Linden Fault System (Jacobi and Fountain, 2002) just west of this location. Other indicators of the fault include: 1) the north-striking valley (topographic lineament) where well-logs indicate a fault (Jacobi and Fountain, 1993), and 2) surface stratigraphy that suggests about 15.25 meters of offset across the fault (from Hillcrest Rd. outcrop (STOP 3) to the lake-level outcrops along the eastern edge of Rushford Lake.

Directions	Distance	Cumulative
STOP 4 - Lunch Stop		
From Caneadea Dam Picnic area, Turn RIGHT onto Mill St.		
Turn RIGHT onto SR-243	1.2 mi	77.5
Turn RIGHT onto SR-19	0.6 mi	78.1
Turn RIGHT onto CR-17/White Creek Rd.	5.7 mi	83.8
Off to the left is a roadcut of Lower Rushford Member, which is also exposed at Stop 4	0.2 mi	84
Turn RIGHT onto LittleJohn Rd.	0.45 mi	84.45
Park along the right side of the road, past the bridge over White Creek., outcrop for is north of the bridge.	0.05 mi	84.5
STOP 5 - White Creek		

STOP 5 – White Creek at the Little John Rd. crossing (Figures 29, 30)
Location Coordinates (WGS 1984):
Lat – N42.31537°, Lng – W78.10702°
UTM Zone 17; N468848, E0738417

White Creek contains some of the best exposures of Luther's (1902) Rushford Fm. Roadcuts north of Little John Rd. display the vertical section; creek exposures show the three, stacked shoreface sequences as well as exhibiting syn- and postdepositional deformation.

North of the bridge is a large exposure of the three-shoreface sequences that comprise the Lower Rushford Mbr. of the Rushford Formation. Basically similar to the Lower Rushford Mbr outcrops we examined near Rushford Lake, the outcrop here is much thicker (~8.5m). The variation of thickness of the Lower Rushford Mbr. coincides with faults of the Clarendon-Linden Fault System (Figure 31). The thicker regions possess preserved transgressive lags on offshore sequences, whereas the thinner regions possess eroded upper-shoreface and foreshore zones.

At the northernmost exposure of the outcrop, the 1st shoreface sequence forms the lowest step in the small falls of the outcrop. The top bed is a fossiliferous, medium to coarse sandstone that typifies the transgressive lags that occur on top of the shoreface sequences in the lower sandstone packet. Overlying the 1st shoreface is a thick interbedded section that is well exposure on the east cliff exposure. Soft-sediment deformation, primarily ball-and-pillow structures are well exhibited in the sandstone beds beneath the 2nd shoreface sequence. The 2nd shoreface sequence displays planer laminated beds typical of the shoreface sequence, but also includes a sediment slide (debrite) with a large sandstone olistolith surrounded by a muddy debris flow with deformed sand clasts. The top of the 2nd shoreface sequence occurs near the main falls/cascade. The sandstone displays small dunes (amplitude - ~0.5m, wavelength ~2.0 m) that are a classic example of symmetrical ripples with tuning forks. The 3rd sandstone sequence forms the upper part of the cascade. The most noticeable feature is the transgressive lag that caps the sequence and the falls. This lag deposit contains large clasts of white, cloudy quartz as well as numerous brachiopod shell fragments and large red silt clasts. The underlying sandstone contains *Rhizocorallium, Arenicolites and Thalassinoides*, typical of a *Glossifungites* firmground. Overlying the 3rd shoreface sequence is the thick interbedded sequence that separates the lower sandstone packet from the upper sandstone packet.

We propose that the stress associated with the Acadian Orogeny caused a reactivation of basement faults, including those of the Clarendon-Linden Fault System and the Iapetan opening faults. Motion on these faults resulted in a number of small fault blocks that affected the basin topography. The motion and rate of slip and rotation among the faults was not identical, with the result that minor topographic highs and lows within the basin were created by small fault block motion that would create variable accommodation space over the faulted area (Figure 32).

We first proposed fault control of Upper Devonian sandstones to explain the occurrence and preservation of shoreface sandstones in the Lower Rushford Member (Smith and Jacobi, 1998, 2000, 2001, 2003 and 2005; Jacobi and Smith, 2004). Syndepositional fault activity would enable coarser sediments to accumulate along active fault blocks at and adjacent to the intersection of the north-south trending Clarendon-Linden Fault System, and the northeast trending shoreline would act as a groin (Figure 33).

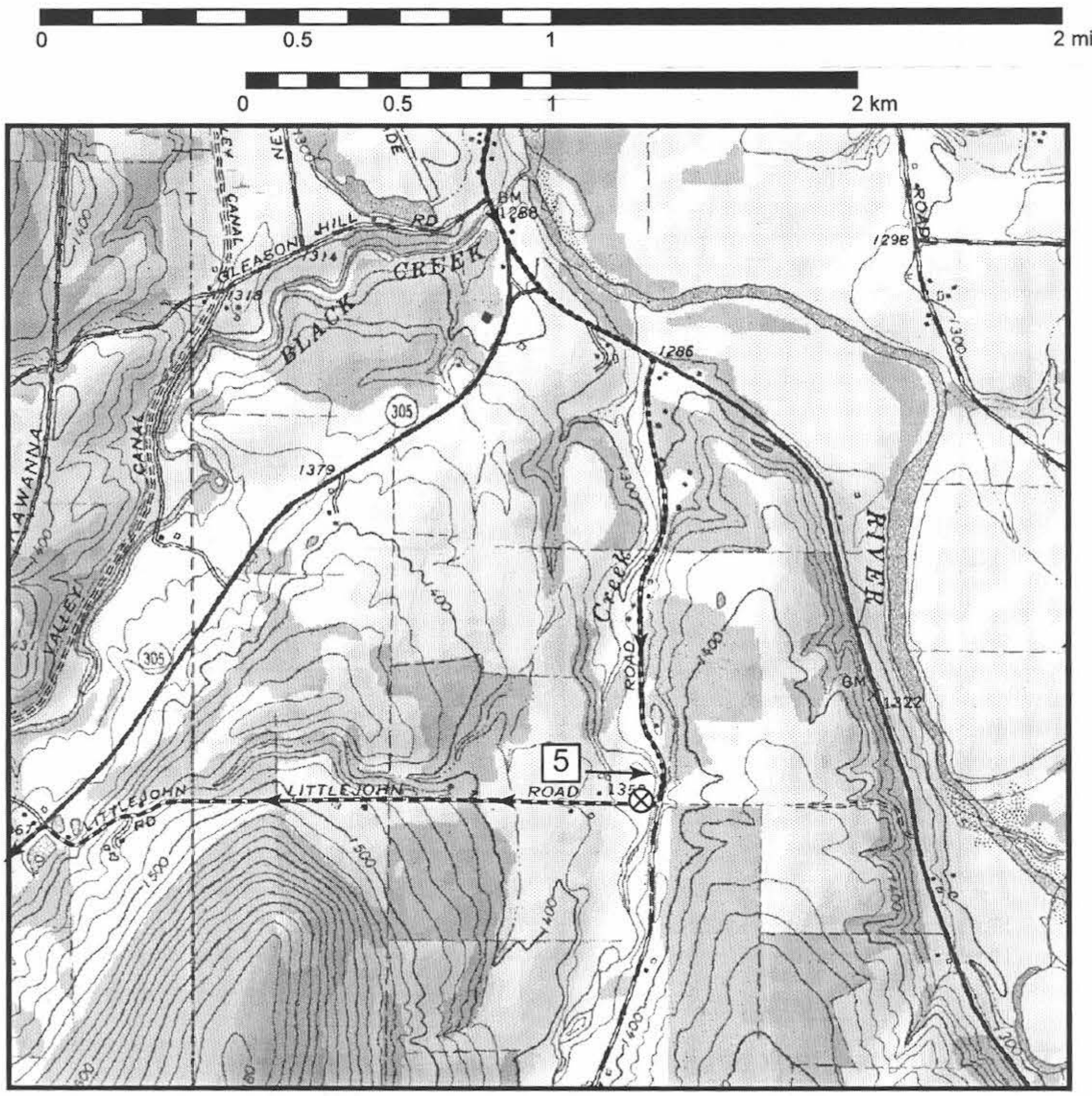
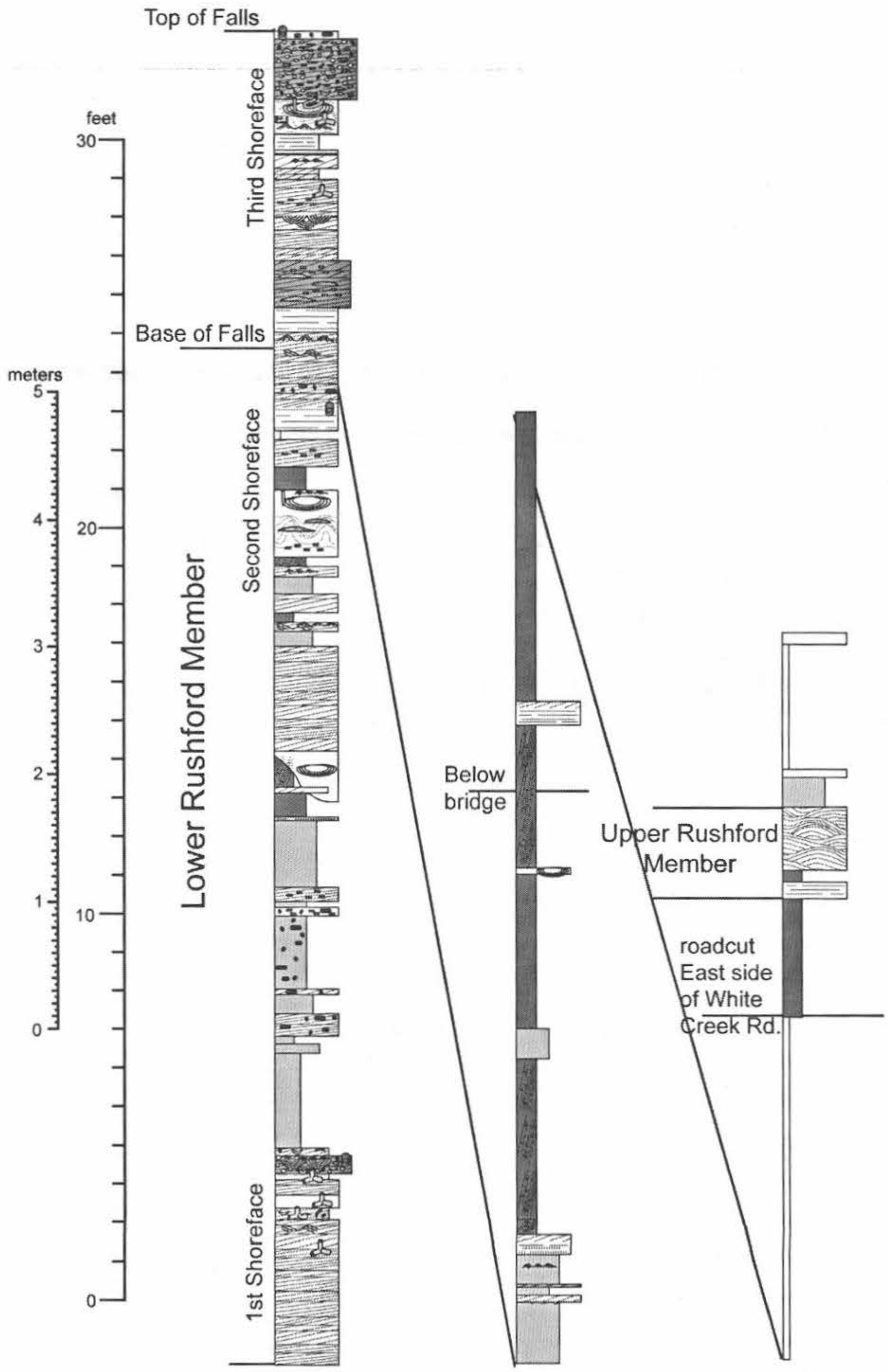


Figure 29. Stop 5 - White Creek at LittleJohn Rd. bridge. Park along north side of the road, west of the bridge crossing White Creek. Take path north along the creek to the cascade (5).



STOP 5 - White Creek at LittleJohn Rd. Rushford Formation

Figure 30. Measured, annotated stratigraphc column for outcrop along White Creek at LittleJohn Rd. Key to the symbols may be found on Figure 10.

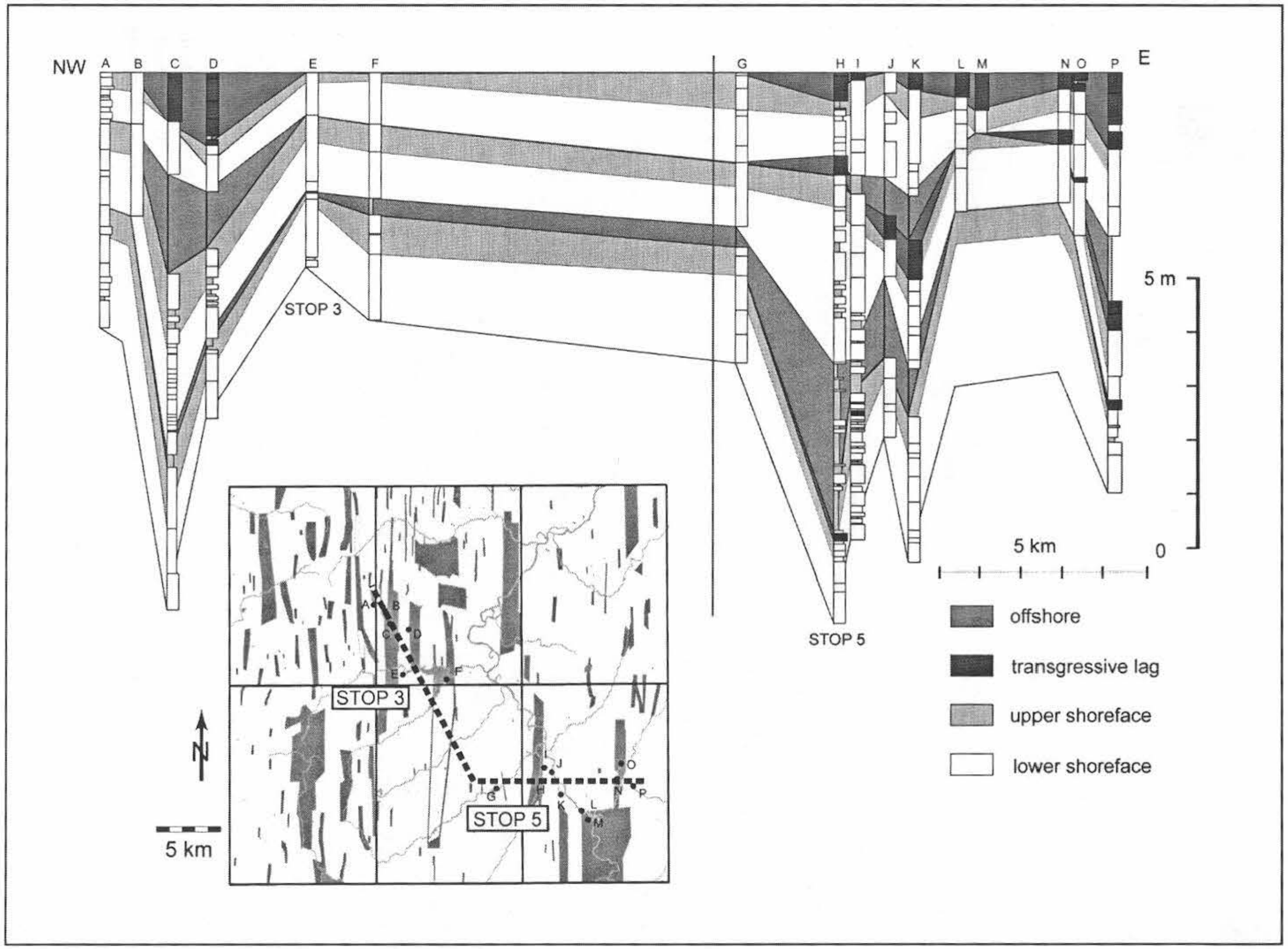


Figure 31. Cross-section of the Lower Rushford Member depicting the three stacked shoreface sequences, Note that the transgressive deposits (lags and offshore) generally coincide with the thicker outcrops, and that the upper shoreface deposits thin where the overall thickness of the Lower Rushford Member is thinner. Map at the bottom shows the relationship between the outcrops and north-south faults of the Clarendon-Linden Fault System (Jacobi and Fountain, 1996).

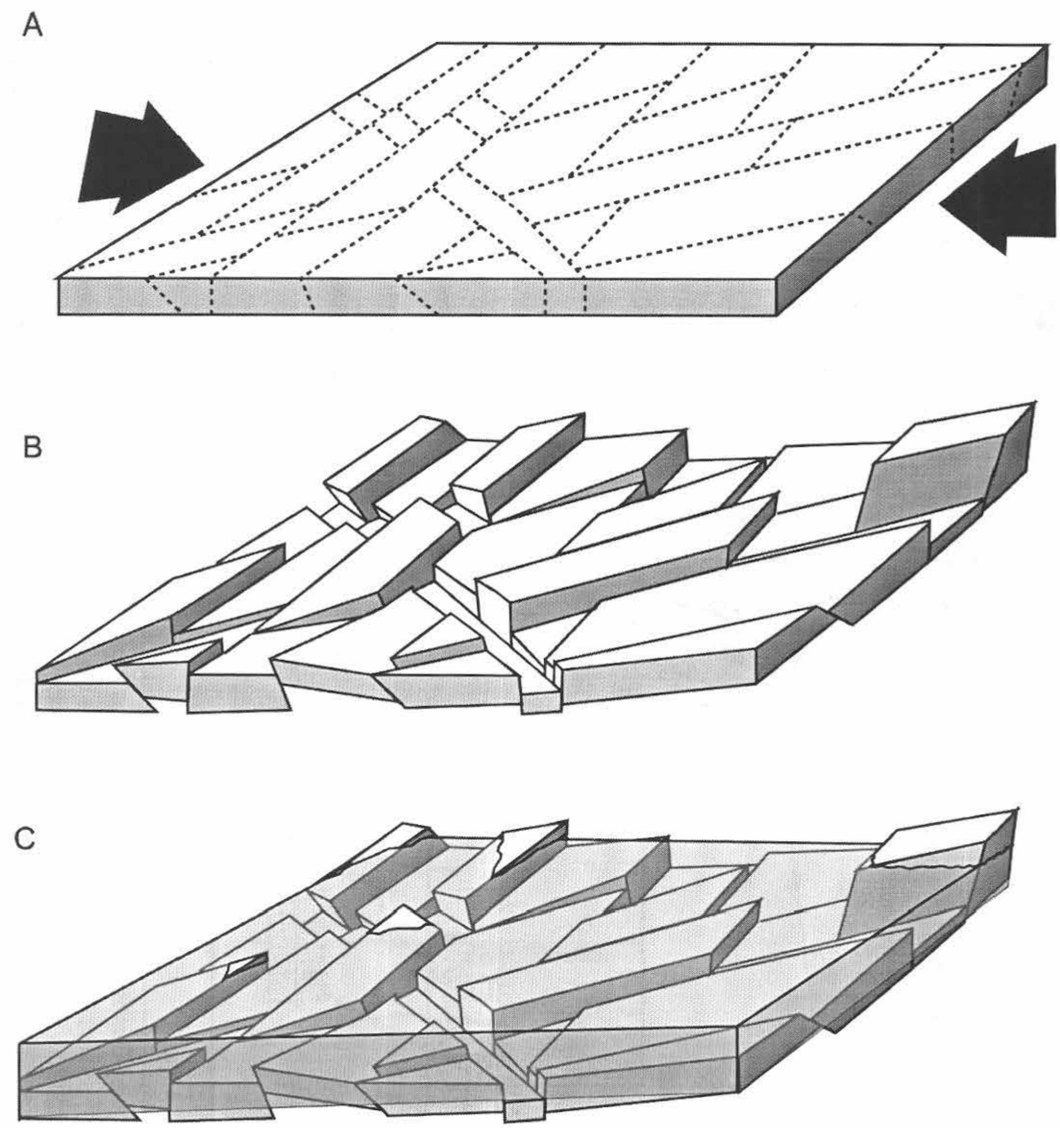


Figure 32. Hypothetical model of fault-block-derived accommodation variability. A) In an area with pre-existing faulting, reactivation is likely during periods of high stress such as compression during an orogenic phase (orange arrows). B) To relieve the stress, the faults may move, but at different rates creating highs and lows between faults blocks. Rotational motion will also create higher and lower regions. C) The overall effect on deposition will be to generate lows that accumulate sediment, and highs that will erode, deflect currents and create multiple trending sediment packets. If the fault activity occurs during deposition, then the changes may be observed within the sedimentary deposits. It is important to note that neither scale or orientation is implied by this model - this model could represent the scale of basins or outcrop jointing.

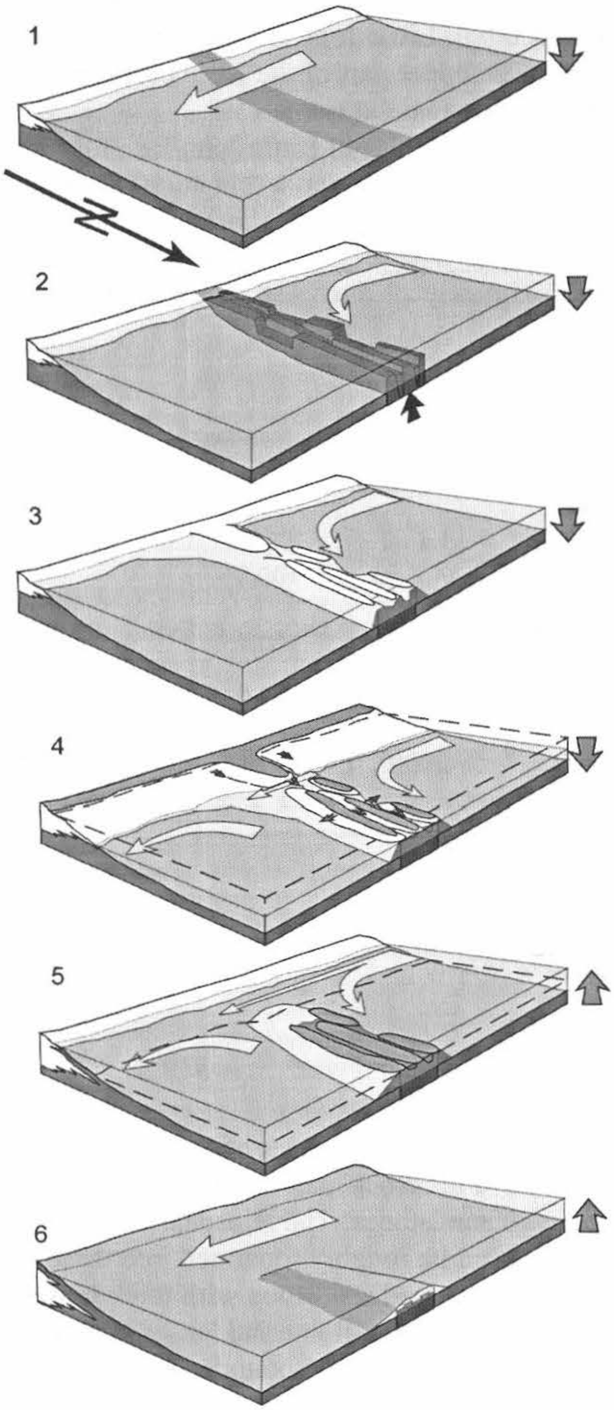


Figure 33. Model proposed for north-south depositional trend observed within the Lower Rushford Member in northern Allegany and Cattaraugus counties.
1) Preexisting fault zone (Clarendon-Linden Fault System) intersects the Late Devonian paleoshoreline at either an oblique or perpendicular angle. 2) Syndepositional fault activity causes variable uplift on the fault-blocks comprising the fault zone blocking the longshore current. 3) The fault zone behaving as a groin would redirect the sediment transport causing the shoreline to follow the trend of the fault zone. 4) During periods 1 through 4 relative sealevel has been falling, by time period 4 the shoreface is now subaerially exposed increasing erosion. Coarser materials (quartz clasts) are redeposited in immediately adjacent accommodation zones, while finer-grained material is transported away. The result is a generally cleaner and coarser sediment. 5) Sealevel rises during the ensuing transgression, reworking and redepositing the coarser material as a transgressive lag deposited at the top of the shoreface sequence. 6) With continued rise in sealevel during the transgression and later highstand tract, the fault-controlled shoreline becomes isolated offshore and is now buried under deep-water sediments. The resulting topographic anomaly may later serve as a depositional focus for shelf-ridge creation.
(Gray arrows represent the rise or fall of relative sealevel. The black arrow represents fault motion. The white arrows represent sediment transport. Small dark gray arrows represent erosion and redeposition of coarser sediments.

Upstream from the waterfalls located north of the bridge, deformation in the shales, silty shales and siltstones are readily observed; both deformed sandstones (ball and pillow) and pencil cleavages occur. At 0.25 mi (0.4 km) south of Little John Bridge, a repetition of the section appears to outcrop in White Creek. The upper part of the 2nd shoreface sequence and all of the 3rd shoreface sequence outcrops again at a small waterfall/cascade. Unlike the outcrop to the north of the bridge at Little John Rd., this outcrop does not have the thick transgressive lag deposits. However, a thin coquinite is observed at the top of the 3rd shoreface sequence and on top of the 2nd shoreface sequence, large cloudy quartz pebbles can be observed. The offset between the outcrop by the bridge and the repeated (?) section is approximately 11.6 m (~38 ft), down on the west (or north).

Directions	Distance	Cumulative
STOP 5 - White Creek		
Continue west along LittleJohn Rd.		
Turn LEFT onto SR-305	1.1 mi	85.6
Turn LEFT onto Stout Rd.	14.7 mi	100.3
Drive straight through Cuba NY		
Turn RIGHT onto Farnsworth Rd.	0.1 mi	100.4
Park along the right side of the road, outcrop is exposed along the left side.	0.2 mi	100.6
Additional outcrop exposed farther along the road	300ft	
STOP 6 - Farnsworth Rd.		

STOP 6 – Farnsworth Road – road cut (Figures 34, 35).
Location Coordinates (WGS 1984):
Lat – N42.18096°, Lng – W78.25373°
UTM Zone 17; N4673512, E0726810

The roadcuts along the northern side of Farnsworth Road are exposures of the Hinsdale Fm. (Chadwick, 1933), (possibly a correlative to the "Pink Rock", an old drillers name), one of the stratigraphically higher Bradford reservoir sandstones. The Hinsdale Formation is comprised of thick sandstone beds that can be roughly separated into three packets. The thick sandstone lenses are typically fine-grained sandstone with thick lenses of coquinite that typically form basal lags, but also occur as small internal lenses (Figure 35). Thickness variability within the sandstones is high, further compounded by variable erosion of lower units by overlying sandstones.

Sedimentary structures are typically SCS with less frequent occurrences of TCS and tabular cross-beds. Small amplitude (and wavelength) symmetrical and interference ripples are commonly observed on the inner surface of large troughs and swales, implying complex, combined-flow during time of deposition (Figure 17). Thinner sandstones within the interbedded sections contain both asymmetric and 3-dimensional ripples and HCS; several thin sandstones also contain grooves and striations on the sole of the beds. Both the sandstones of the interbedded sections and the thick sandstone packets contain an abundant, but still low-diversity *Skolithos* ichnofacies. High frequency of *Arenicolites* and *Skolithos* occur in both the thinner sandstones as well as in the thick sandstone packets. A poorly developed *Cruziana* ichnofacies consisting of sporadic occurrences of Planolites and Teichichnus is found in the thicker interbedded sections.

We interpret the Hinsdale Formation as being deposited well above storm wave base, but still below fair weather wave base, most likely upper offshore to lower shoreface. The dominance of SCS and typical tempestite packets imply the Hinsdale Formation was deposited primarily by storms; the large amounts of fossils and the singular occurrence of red shales and sandstones suggest the shoreline was moving westward and terrestrial sediments were now being incorporated into the marine sedimentary deposits. The interpreted depositional environment is likely to be either nearshore storm-generated sand ridges or a storm-influenced barrier bars (Miall, 2000)

The roadcut displays the sharp-based sandstones that have lenses of coquinite throughout, but generally a thicker lag deposit at the base. Bedding in the sandstone is distinctly swaly, with large erosional surfaces cutting down into underlying units. At the lower roadcut, above the prominent, thick sandstone is a red to purplish red shale which aids the identification of the formation. The depositional environment for the Hinsdale and many of the similar thick sandstones between the Lower Rushford Member (STOPS 3, 4 and 5) and the Hinsdale Fm. are probably storm shelf ridges to storm-dominated barrier bars. The lateral variability observed at this location and many similar outcrops may explain the "lensing" nature of the Upper Devonian reservoir sands.

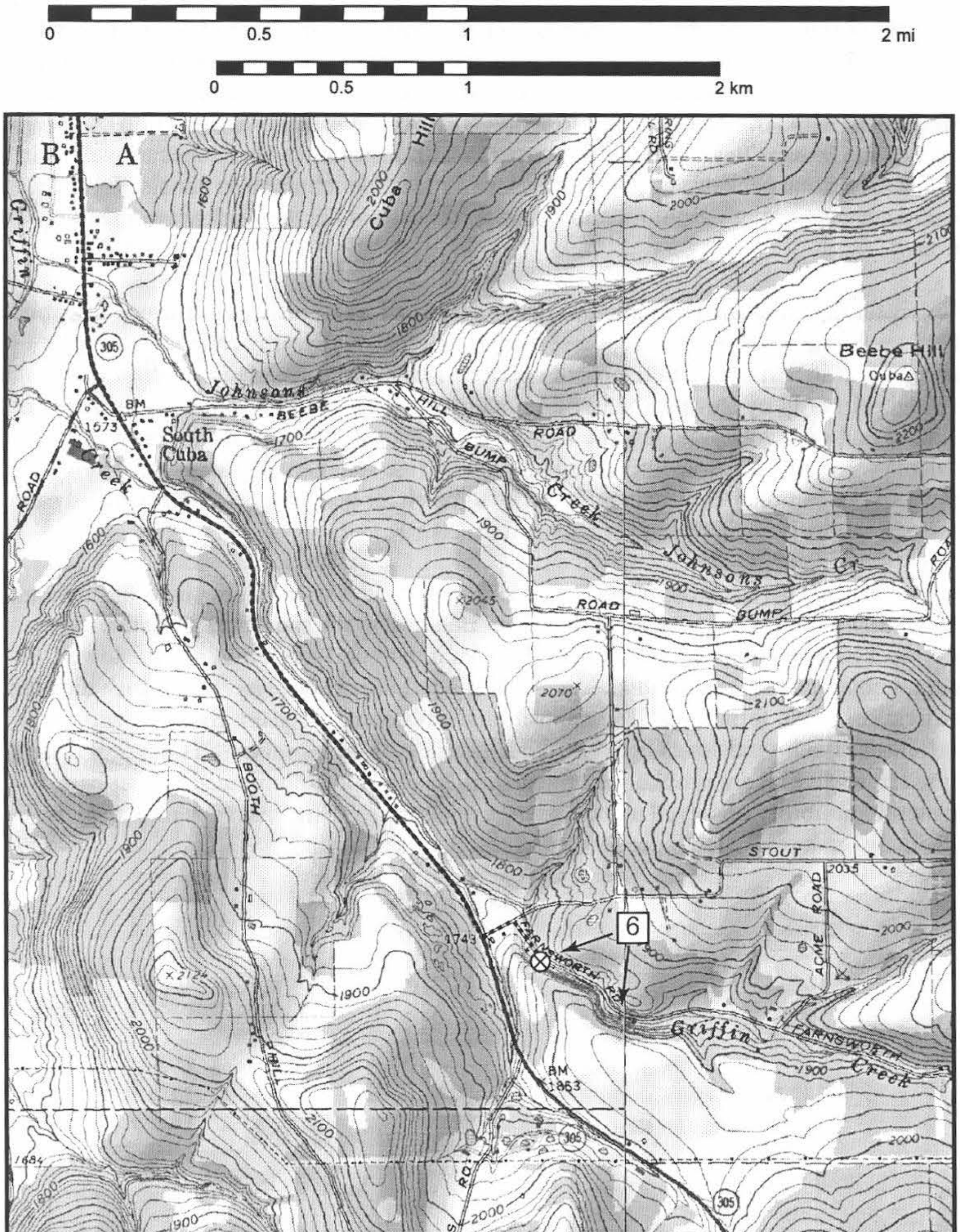
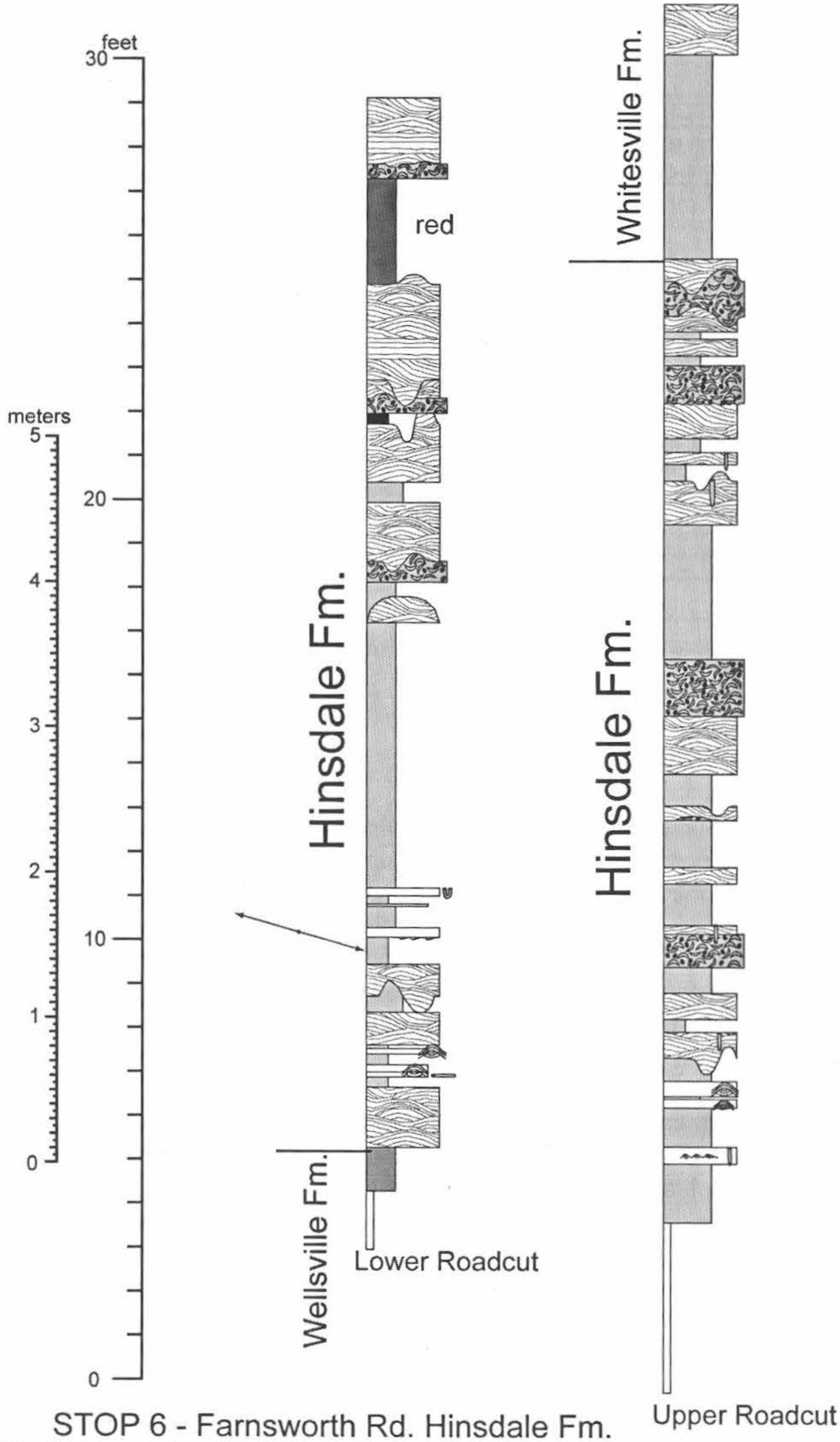


Figure 34. Stop 6 - Farnsworth Rd., roadcut. Park along south side of the road, past the house and garage, opposite is the lower roadcut. Outcrop is also present in Griffin Creek, but you will need permission from the landowner. Additional outcrop is accessible farther east along the road (marked by the second arrow).



STOP 6 - Farnsworth Rd. Hinsdale Fm.

Figure 35. Measured annotated stratigraphic columns for roadcuts along Farnsworth Rd. Arrow represents measured paleoflow direction. Key for symbols used may be found on Figure 10.

Directions	Distance	Cumulative
STOP 6 - Farnsworth Rd.		
Turn around and Turn LEFT onto Stout Rd.	0.2 mi	100.8
Turn RIGHT onto SR-305 heading N	0.1 mi	100.9
Turn LEFT onto ramp to US-86 W (old SR-17)	3.5 mi	104.4
Merge onto US-86	0.4 mi	104.8
Take EXIT 21 at Salamanca	30.4 mi	131.2
Turn RIGHT onto Parkway Dr.	0.1 mi	131.3
Keep RIGHT onto SR-417	0.1 mi	131.4
Turn LEFT onto US-219	0.5 mi	131.9
Turn LEFT onto Hungry Hollow Rd.	2.4 mi	134.3
Turn LEFT following Hungry Hollow Rd. (straight is McCarthy Hill Rd)	3.6 mi	139.9
Turn LEFT onto Little Rock City Rd	0.4 mi	140.3
Continue South to Parking Area	1.7 mi	142
STOP 7 - Little Rock City		

STOP 7 – Little Rock City (Figures 36, 37).
Location Coordinates (WGS 1984):
Lat – N42.20853°, Lng – W78.70802°
UTM Zone 17; N4675470, E0689203

Little Rock City is a State Forest under the supervision of the NYS Department of Environmental Conservation. The Salamanca Conglomerate (Carll, 1880) marks the basal contact of the Conewango Group with the underlying Conneaut Group. Sands and coarser units of the Conewango Group are correlative to the Venango reservoir sands in Pennsylvania. At this site exposures of thick orthoconglomerate (between 2 and 12 m thick) occur along the northern and eastern sides of the hill. House-size blocks separated along joints by periglacial expansion are abundant throughout the area; at least one was redeposited upside-down on top of another block by glacial activity.

Lithologically the Salamanca is an oligomictic quartz orthoconglomerate. The conglomerate consists of a medium to coarse sand matrix with clasts of cloudy white quartz. The clasts are oblate discoids that range from 0.5 centimeters up to 6+ centimeters in diameter. The oblate discoid shapes of the pebbles are typical of a wave-dominated shoreface. Large downlapping, graded cross-sets (1 to 4 meters high) and channels occur throughout the lower section of the outcrop. These bedforms may represent alluvial gravel bars. The form of the large cross sets is also similar to that found in estuarine tidal sand bars, but the thickness of the sets is high for the low (about 2 m) tidal fluctuations thought to characterize the Acadian Catskill Sea (e.g., Woodrow and Isley, 1983). However, the coarseness of the conglomerate indicates a high energy system, consistent with a tidal prism or bore in an estuary (and also consistent with the tidal gravels in the English Channel) that could have had a higher tidal range than the assumed 2m Catskill Sea tidal range. Thus, these large crossbeds could be fluvial bars or tidal. Tabular crosssets and crossbeds with a herringbone pattern are observed above the large cross sets (although some of the herringbone is a "pseudo-herringbone" that results from trough cross sets in opposite directions). Both herring bone patterns, however, are consistent with reversing currents of tidal flux, as in a tidal channel. Overall, the large scale crossbeds, trough cross-sets, and the herringbone pattern suggest an estuarine/estuarine mouth depositional environment (Dalrymple, 1992) with a strong tidal influence. Paleoflow indicators from sedimentary structures are few, ripples and troughs on the upper surface occur sporadically.

We recognized several distinct facies within the Salamanca Conglomerate: facies A is the caprock of the Salamanca Conglomerate with the largest quartz clasts (>5cm). The conglomerate is generally disorganized with clasts orientated obliquely and perpendicular to bedding, except for the top surface; large carbonized wood fragments also occur. Facies B is a less resistant, more friable unit that exhibits herringbone bedding marking current reversals. Facies C is comprised of two thick, graded foreset packets each ranging 1 to 4 meters thick with large clasts accumulating along bedding planes with some asymmetric ripples. Facies D is a less resistant, more friable unit that contains fewer clasts, but generally does not exhibit reversals in bedding (Figure 47).

We have observed similar arrangement of facies at conglomerate blocks loose in Allegany State Park and adjacent areas, and at the "Bear Caves" area outcrops at Mount Seneca in Allegany State Park. The outcrops and loose blocks examined south of Little Rock City were generally sandier than the Little Rock City conglomerates, with clasts occurring only along bedding planes, and the clasts typically are not as large (maximum clast ~ 2 cm) as the clasts measured at Little Rock City.

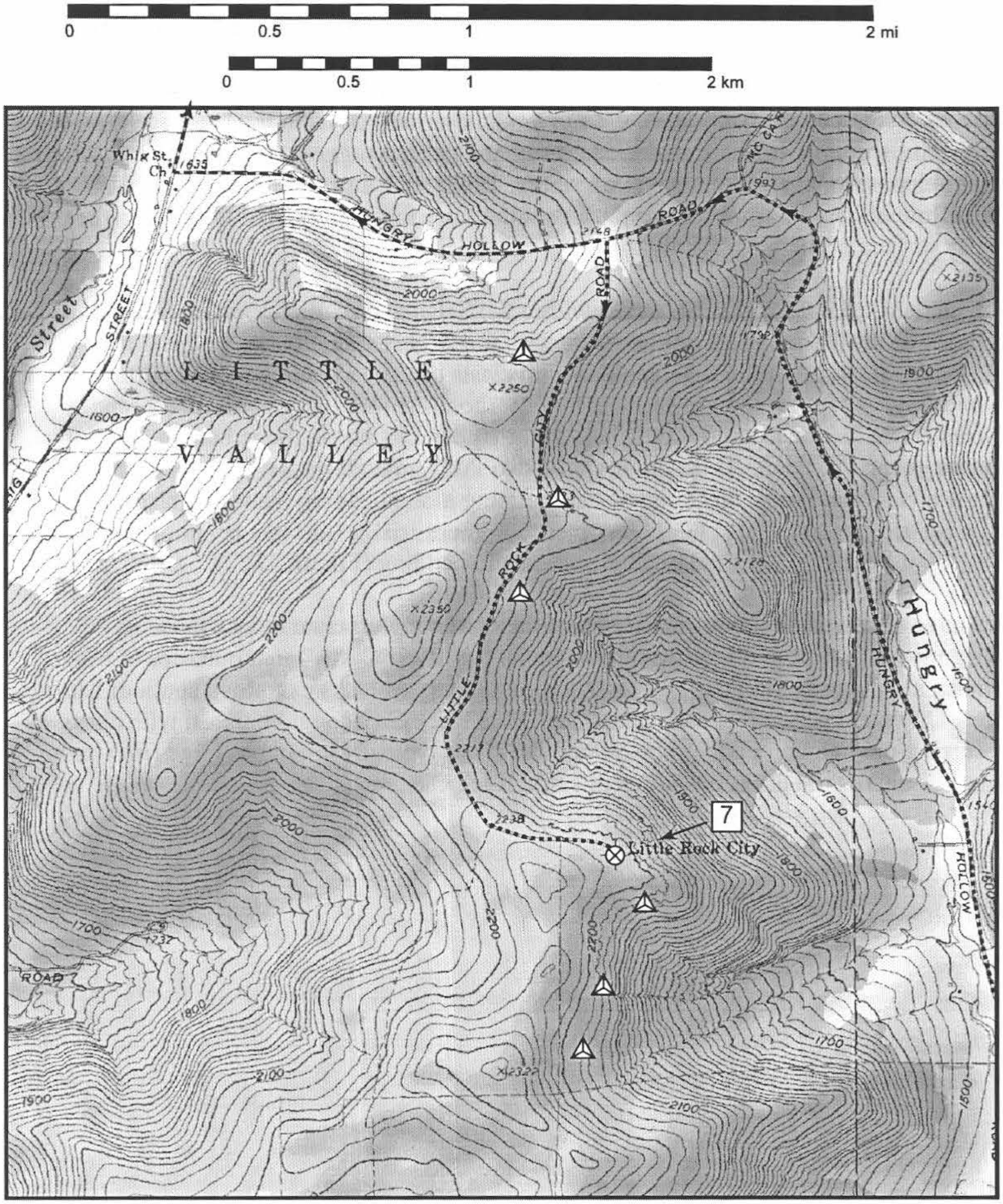
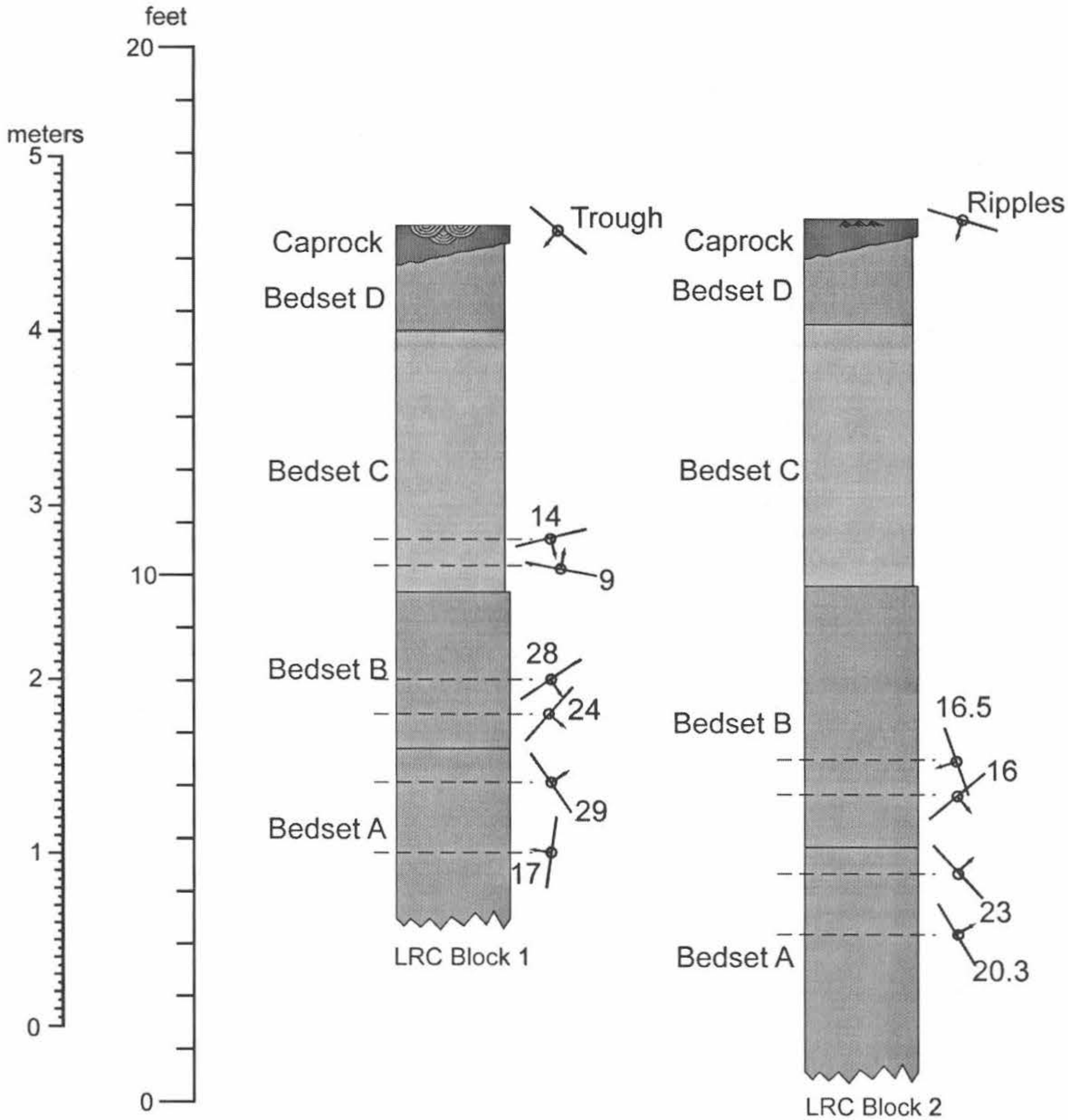


Figure 36. Stop 7 - Little Rock City. Park at the parking circle at the end of Little Rock City Rd. Outcrop is accessible immediately north of the parking area, but is easily found thoughout the area (marked by triangles).



STOP 7 - Little Rock City. Salamanca Conglomerate.

Figure 37. Measured, annotated stratigraphic columns for some of the outcrops at Little Rock City. Strike and dip symbols represent the calculated true strike and dip of measured crossbeds within the conglomerates. Block 1 is due north of the parking area; Block 2 is immediately east of Block 1.

Directions	Distance	Cumulative
STOP 7 - Little Rock City		
Travel north along Little Rock City Rd.		
Turn LEFT onto Hungry Hollow Rd.	1.7 mi	143.7
Turn RIGHT onto Whig St	1.1 mi	144.8
Turn RIGHT onto RT-242	2.2 mi	147
Turn LEFT onto US-219	3.8 mi	150.8
Turn LEFT following US 219 (also SR-39)	17.3 mi	168.1
Turn RIGHT onto ramp for US-219 (Southern Expy)	0.1 mi	168.2
Keep STRAIGHT onto Ramp towards I-90	23.6 mi	191.8
Keep LEFT to stay on ramp towards I-90/Thruway/ Ridge Road West/Buffalo/Lackawanna	0.4 mi	192.2
Take ramp (LEFT) onto I-90 toward I-90/Thruway/Buffalo	0.3 mi	192.5
At exit 53, take ramp (RIGHT) onto I-190 toward I-190/Downtown Buffalo/Niagara Falls	3 mi	195.5
At exit 7, turn RIGHT onto ramp toward Church St.	5.5 mi	201
Bear RIGHT (East) onto Broadcast Plaza (Church St)	0.2 mi	201.2
FINISH - RETURN TO ADAMS MARK		

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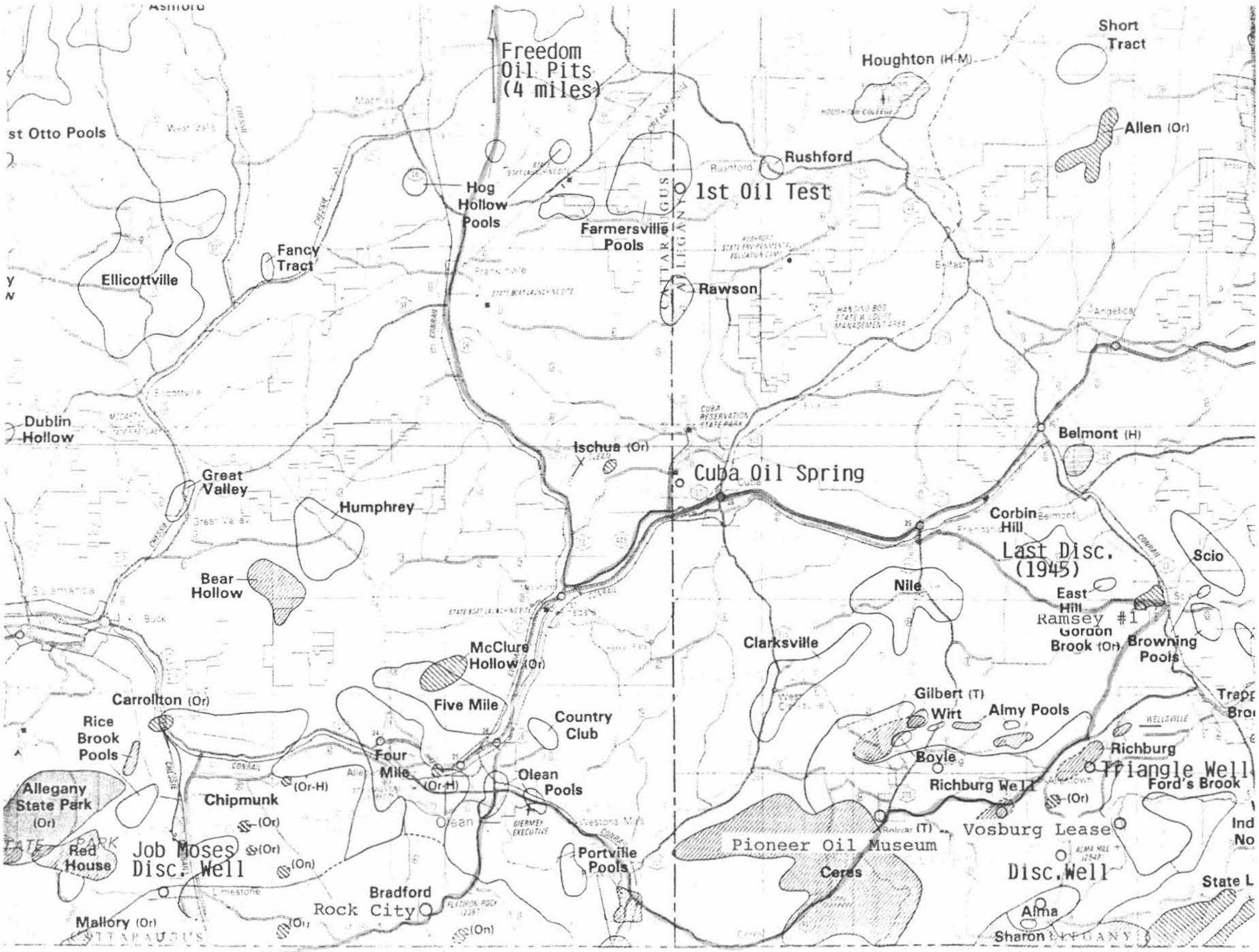
NORTHERN APPALACHIAN BASIN OIL AND GAS: HISTORICAL PERSPECTIVE

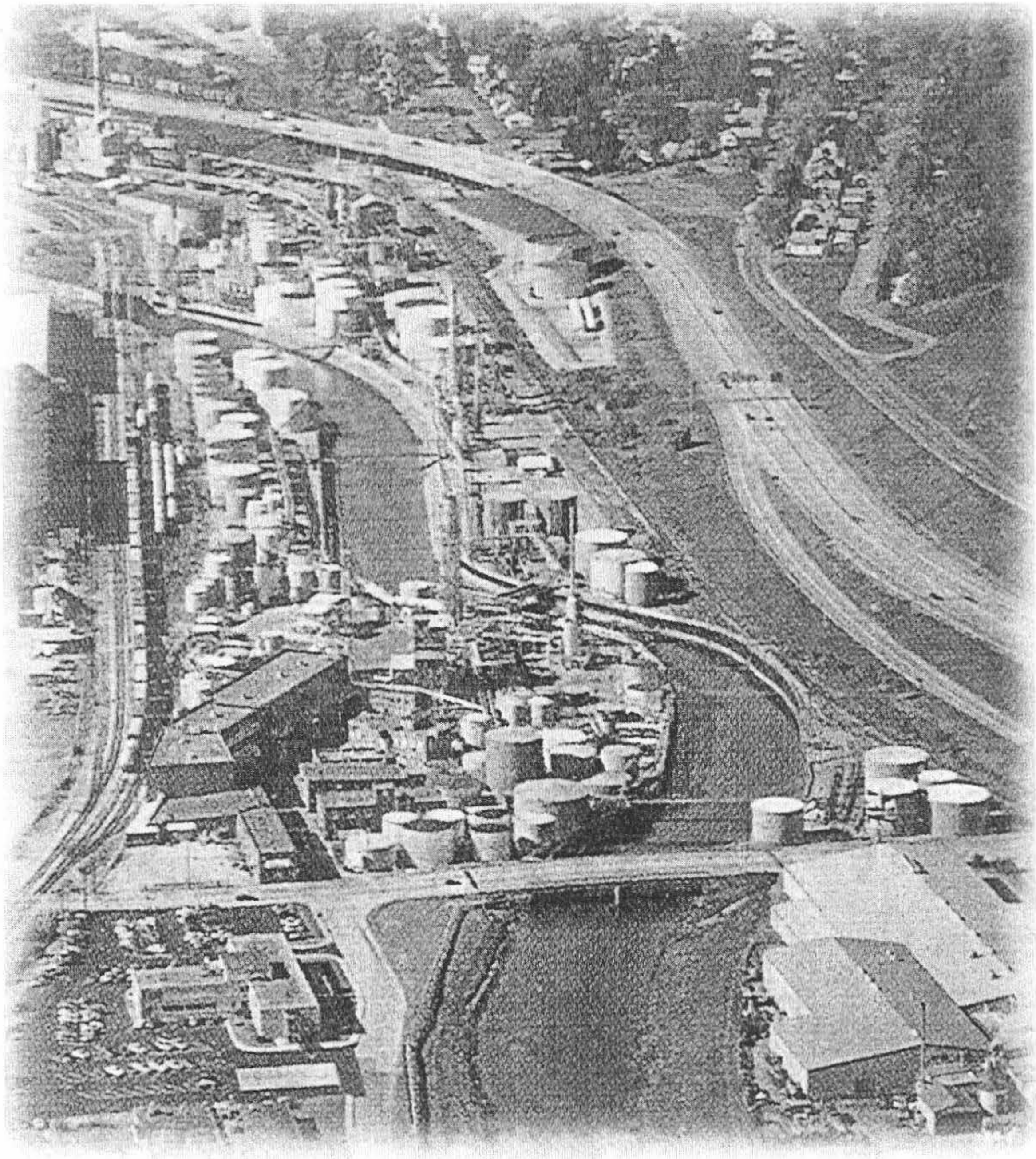
Field trip to Bradford, Pennsylvania and various locales in southern Cattaraugus and Allegany Counties, New York.

Held on Saturday, October 7, 2006 in conjunction with the joint annual meetings of the New York State Geological Association and the Eastern Section - American Association of Petroleum Geologists in Buffalo New York.

Arthur M. Van Tyne, Field Trip Leader.



Bradford Refinery 1 inch = 4 miles Van Tyne Consulting, Wellsville, NY 9/06

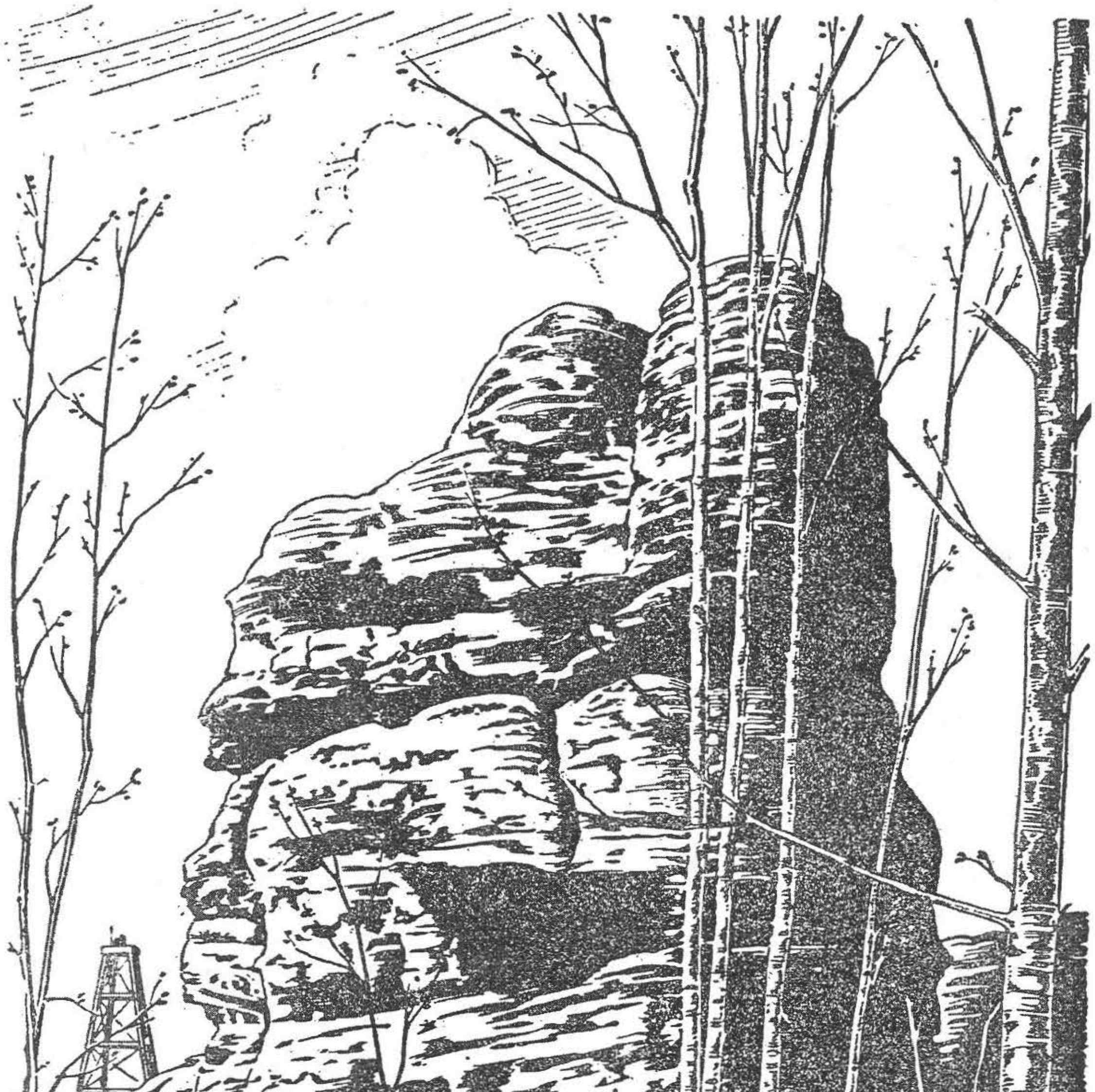


BRAD PENN® Refinery
Bradford Pennsylvania

Stop 1. Our trip will proceed first to Bradford, Pennsylvania where we will visit an oil refinery owned by the American Refining Group. This refinery was started in 1881 and was formerly owned by the Kendall Refining Corp. This summer its 125 anniversary was celebrated by American Refining. It is now said to be, "the oldest continuously operating refinery in the U.S. and the oldest refinery in the world dedicated solely to processing 100 % Pennsylvania Grade crude oil." It is one of only two small refineries still operating in all of central and western Pennsylvania. The Kendall company became famous for its two fingers in the air ad signifying that by the use of Kendall Pennsylvania grade oil you only needed to change the oil in your car motor every 2,000 miles instead of sooner with the use of other motor oils!

Stop 2. After leaving the refinery we will proceed northeastward through Bradford and the Bradford oil field up into a hilly area between Bradford and Olean, New York. Just across the NY-PA State line we will stop at the tiny hamlet of Rock City. This place received its name from the huge rock formations which occur here. These rocks are the Olean conglomerate and sandstones of the Pennsylvanian Pottsville Group. These have been described as fluvial deposits by some researchers. Coming up the hill we passed outcrops of the underlying shales and sandstones of the Mississippian Pocono Group. At this locale the Olean conglomerate has been eroded into a literal "Rock City" and you can now walk down the enlarged joints as "streets" between the rock "buildings". Rock City is actually the technical geologic term for such erosional features. Because of the high purity of the silica in the Olean conglomerate nearby outcrops were quarried extensively during WWII for their silica content. Those pebbles were used in the manufacture of high fidelity radio frequency crystals which were installed in military and aircraft radios.

This rock city has been a Sunday outing site for people from the surrounding area since at least the early 1900s. The current owners, Cindy and Dale Smith, purchased the park from its former long-time owner only a few years ago. They have made a number of improvements to the trail system and have built a new pavilion with tables for picnickers which we may use for our lunch time today.



Lone Rock Towers High Above Oil Derrick at Rock City

Stop 3. Our next stop will be at the Pioneer Oil Museum in the Village of Bolivar, NY. The museum was started in the 1970s by several local oil producers in an effort to save as much as possible of the artifacts of the early oil industry of this area. Old oil field equipment, data, maps and other information was usually discarded as various operations closed down. The museum has been quite successful in obtaining many gifts of old equipment and records from families and acquaintances of pioneer oilmen.

The building in which the Museum is now located was originally the McEwen Brothers Oilfield Supply Store. It was purposely located here in the heart of the Allegany oil field. An adjacent addition to the north was recently erected with the help of a New York State grant but the Museum has now outgrown its present site. A larger, and better constructed, building nearby has recently become available and the Museum will be moving to that site as soon as possible. That building, the old Hahn & Schaffner Supply Co. establishment, has been purchased by the New York State Oil Producers Association for the Museum.

Stop 4. After leaving the Pioneer Oil Museum we will proceed eastward on Route 417 to the Vosburg oil lease owned by Plants & Goodwin. Mr. Paul Plants, the owner, is currently the President of the New York State Oil Producers Association and also of the board of the Pioneer Oil Museum.

The lease consists of 95 oil wells and 102 water inmput wells on a 275 acre tract. It has been operated for more than 60 years. This is a fully developed secondary oil recovery waterflood operation. The last major development took place in 1982 but one new oil well was drilled in the late 1990s. The oil wells are producing from the Richburg sandstone which is quite thick here. The average depth of these wells is 1,300 to 1,400 feet. About 2½ million barrels of oil has been recovered from this lease over a long period of time. We will be walking in about 600 to 700 feet to visit a central power operation with a large eccentric shaped band wheel which pumps several wells with one rotation. This central power is one of only a few still operating in the New York-Pennsylvania oil fields

Stop 5. Gordon Brook Oriskany Gas Field. We are stopping by here briefly only because it is on the way to our last stop.
We are looking at the No. 1 Ramsey, the discovery well of the Gordon Brook Gas Field. The well was originally drilled in 1970 by Professional Petroleum Exploration but is now owned by Vandermark Exploration.

Gas Production is from the Oriskany sandstone at a depth of 3,840 feet. The well came in with a flow of 6 to 7 million cubic feet of gas per day. It was still producing gas up to a few years ago but now makes too much salt water when production is attempted. The well has produced about one-half billion cubic feet of gas. Just beyond the well head can be seen the separator where the brine is separated from the gas stream. The brine is then stored in the nearby tank until it can be disposed of properly. This is an example of Oriskany sandstone gas production from a small structure

which is extensively faulted and where the Oriskany is pinching out. Gas has migrated into an offset in the pinchout line and later into the highly broken up sandstone to high areas along the faults. The field itself has produced about five billion cubic feet of gas and still has three producing wells.

Stop 6. Seneca Oil Spring.

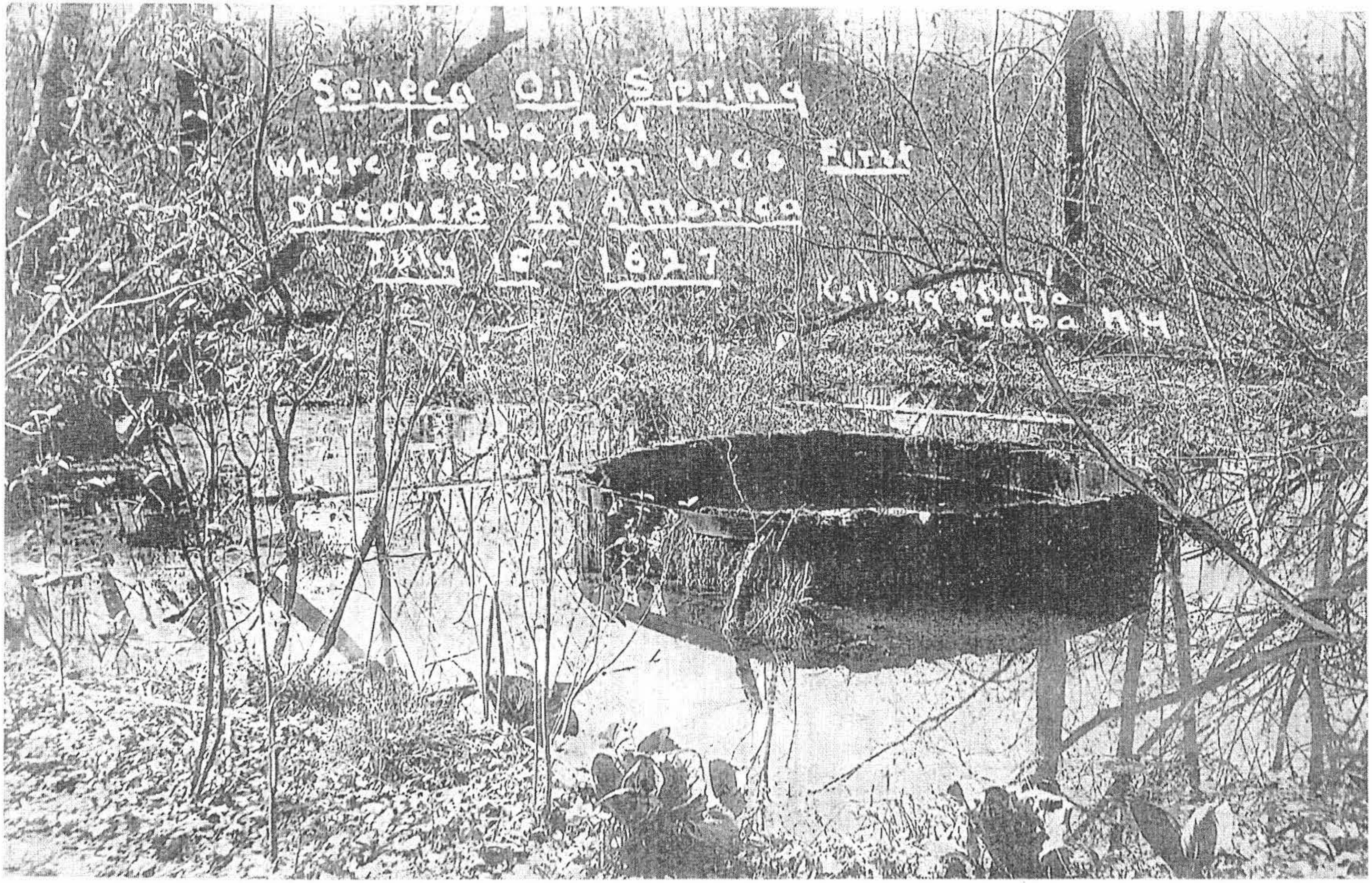
This locale, sometimes known as the Cuba Oil Spring, is located about one and one half miles northwest of the Village of Cuba in central-western Allegany County, NY. This once active oil seep, well known to the local Native Americans, has been reported by many historians of the petroleum industry to have been, in 1627, the site of the first observation by Europeans of petroleum on the North American continent. This was said to have occurred when a Franciscan Recollect Friar, Father Joseph de la Roche Daillon, was taken to see the spring by a group of friendly Native Americans.

Fr. Daillon had been sent to Quebec, Canada by his superiors in France to search for and assist Fr. Nicolas Viel who was working as a missionary to the Hurons. He arrived in Quebec in May of 1625 and soon heard of the death of Fr. Viel in June of 1625. In mid 1626 he journeyed westward to the land of the Hurons. Later, in October of 1626, he journeyed westward again to the land of the Neutrals. While there, on July 18, 1627, Fr. Daillon wrote a letter to a friend in France telling of his experiences with Native Americans. That letter was published in 1636 by Fr. Gabriel Sagard Theodat in his "Histoire du Canada" and became the basis for later interpretors to claim that Fr. Daillon had been taken to see the Cuba Oil Spring.

The Cuba Oil Spring may still be the first place where Europeans saw petroleum in North America but, in the light of further research, we will have to change the date a bit. In a 1962 AAPG Bulletin the late Dr. John Wells, a distinguished Professor of Geology at Cornell University, published a paper in which he disputed the evidence for the Fr. Daillon visit. He quotes the abovementioned Daillon letter as telling of the abundance of vegetables which were available as well as, "some very good oil" (huile). Wells felt that this term refers to an edible oil and believed that Fr. Daillon would have used the word petrole (rock oil) for such an occurrence as the Cuba oil spring because that term had already been applied to oil seeps which had been known in Europe and the middle east for hundreds of years.

He states that the earliest true reference to the Cuba oil spring was made in 1656 by later Jesuit missionaries who described it thusly, "As one approaches nearer to the country of the Cats (Eries) one finds heavy and thick water, which ignites like brandy, and boils up in bubbles of flame when fire is applied to it." This has also been documented in a paper by Thomas, et al given at the Forum on History of Petroleum Geology at the May 1998 AAPG Annual Convention held in Salt Lake City, Utah.

The oil probably came from a subcrop of the Bradford First, and possibly also the underlying Chipmunk sandstone zone under 100 feet or so of valley fill. There is no crude oil visible at the Cuba oil spring now and there has not been any for many years. It


Seneca Oil Spring
Cuba NY
Where Petroleum was First
Discoverd In America
July 18 - 1627
Kellogg Studio
Cuba NY

is probable that the drilling of several shallow wells around the spring in the early 1860s depleted whatever was left of oil which had accumulated under the fill.

By the Treaty of Big Tree in 1797 one square mile around the spring was set aside as an Indian Reservation. The large plaque placed on the boulder of Olean conglomerate and dedicated in 1927 reads as follows:

SENECA OIL SPRING

Its history forms the first chapter in the development of the petroleum industry in America, a gigantic world enterprise transforming modern life.

1627 - Oil on American continent first recorded in this region by the Franciscan Fr. Joseph de la Roche Daillon.
1656 - Spring mentioned by Jesuit Fr. Paul LeJeune.
1721 - Visit by Joncaire.
1767 - Oil from this spring sent to Sir Wm. Johnson as a cure for his wounds.
1797 - Spring permanently reserved by Indians in Treaty of Big Tree.
1833 - Description of spring by Prof. Benjamin Silliman of Yale University.

Erected in 1927 by the University of the State of New York and the New York State Oil Producers Association.

Road Log for Field Trip A-3

Parking lot of Adam's Mark Hotel - Buffalo, NY, Saturday, 10/7/06, 8:30 A.M.
Circle around Hotel and take ramp up to Rt. 190, bear left at first intersection (do not go right onto the Skyway), and drive about 6 miles to Exit 53 and Rt. 90.
Go south (bear right) on Rt. 90 and go four miles to the beginning of Rt. 219, Take Rt. 219 south for 23 miles to the Springville Exit. Here, Rt. 219 becomes a two lane road. Drive south on Rt. 219 for about 27 miles to an exit onto Rt. 417 and 219 at Salamanca. Turn left (Eastward) onto Rt. 417 and Rt. 219 and follow the signs for about 3 miles until directed across the River onto Interstate 86. Go east for 6 miles to Bradford Exit on I86. Get off at Bradford Exit and drive south (turn right) on Rt. 219 again for 11 miles to the Refinery in Bradford, PA. Get off at the Kendall St. Exit to go east to the Refinery gates.

Leaving the Refinery go east on Kendall Ave. to Main St., turn left on Main and go 1½ miles to Foster Brook intersection. Turn right onto PA. Rt. 346 and drive for 2½ miles to PA Rt. 646 intersection. Turn left onto PA Rt. 646 and drive for 5 miles, across the NY-PA State line and onto NY Rt. 16, to Rock City, NY.

Leaving Rock City go north on Rt. 16 for 8 miles into Olean, NY and the Main-State St. intersection just past a small park on the right. This is the Rt. 16-Rt. 417 intersection - go right onto Rt. 417 and drive for 20 miles to the Village of Bolivar. The Pioneer Oil Museum will be on your left in the center of the Village next to a gasoline filling station.

Leaving the Pioneer Oil Museum stay on Rt. 417 and drive a short distance to a traffic light where you should turn right and go out of Bolivar past the school. It will be about 5 miles to the Vosburg lease which will be on the right hand side of Rt. 417. You can pull off onto either one of two roads there.

Leaving the Vosburg lease, continue driving east on Rt. 417, through the hamlet of Allentown, for about 3 miles to an intersection with Knight Creek Rd. (County Rt. 9). NOTE: If you miss the intersection you will be driving up a fairly steep, long hill - you can turn around at the top at a Motel and come back down to the missed intersection. Bear left onto Knight Creek Rd. and drive for 5½ miles to an intersection with Back River Rd. just outside of Scio, NY. Turn left there onto County Rt. 31 and go for 3/4 mile to the Ramsey well on the left.

Leaving the Ramsey well go back onto County Rt. 31, turn left and drive for eight miles to Friendship, NY. Turn left at the intersection with County Rt. 20, just past the P.O. and in the center of the Village. Go through and out of the Village to a major intersection with a blinking light where you will be directed to take a right turn to get onto I 86. Go right for about ½ mile and after going under the highway turn right onto the westbound entry onto I 86. Drive west for 7 miles on I 86 and get off at the Cuba Exit which will bring you down to intersect with Rt. 305. Turn left here, go under I 86, to the traffic light and turn right onto Rt. 446. Drive for 1 3/4 miles to a road to the north (your right) which has a sign for the Cuba Oil Spring. Turn right and go 1¼ miles to the entry road to the Spring - there are marker stone piers here, turn right and go in for a few tenths of a mile to the parking area.

Leaving the Oil Spring, turn left at the entrance and go south to the intersection with Rt. 446 where you turned north previously. Turn right onto Rt. 446 and go 5½ miles to Rt. 16 at Maplehurst. Turn right and take Rt. 16 northward to Yorkshire Corners, a distance of 28 miles. From Yorkshire continue north on Rt. 16 for 14 miles to an intersection with I 400 at South Wales. Take I 400 northwesterly to its blend-in intersection with I 90 a distance of 15 miles. Blend right onto I 90 and drive for about 2 miles to Exit 53. Follow directions to go to Downtown Buffalo at Exit 53 and drive 6 miles to the Church Street Exit and the Hotel.

MIDDLE – UPPER DEVONIAN DEPOSITIONAL AND BIOTIC EVENTS IN WESTERN NEW YORK

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INTRODUCTION

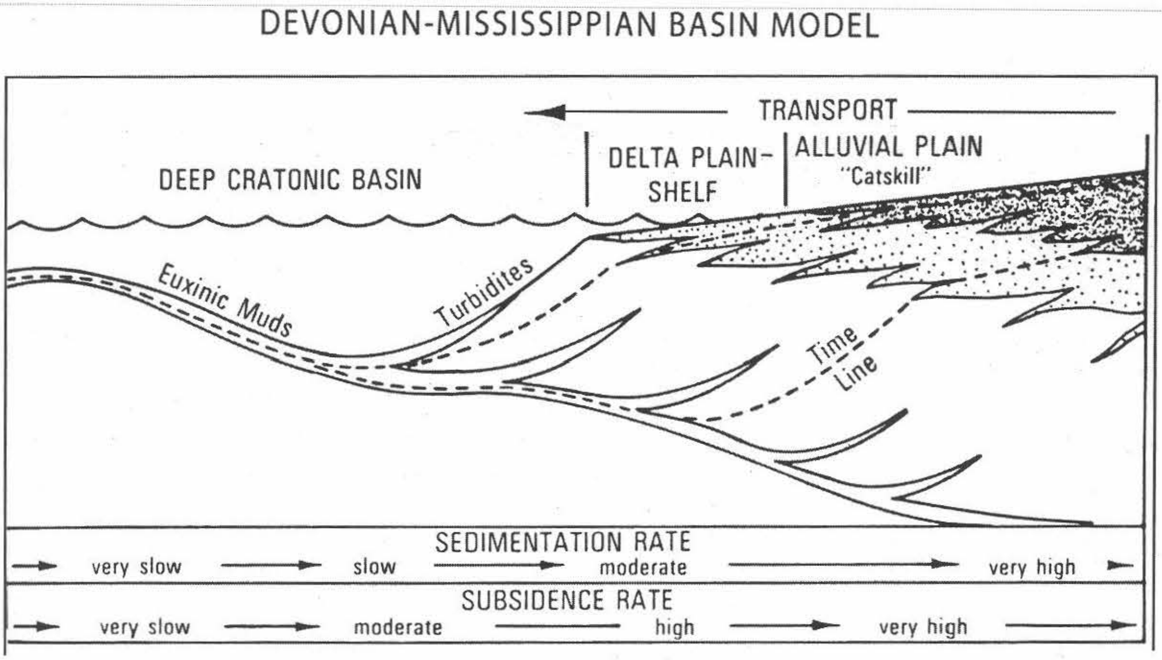
The Middle and Late Devonian succession in the Buffalo area includes numerous dark gray and black shale units recording dysoxic to near anoxic marine substrate conditions near the northern margin of the subsiding Appalachian foreland basin. Contrary to common perception, this basin was often not stagnant; evidence of current activity and episodic oxygenation events are characteristic of many units. In fact, lag deposits of detrital pyrite roofed by black shale, erosional runnels, and cross stratified deposits of tractional styliolinid grainstone present a counter intuitive image of episodic, moderate to high energy events within the basin. We will discuss current-generated features observed at field stops in the context of proposed models for their genesis, and we will also examine several key Late Devonian bioevents recorded in the Upper Devonian stratigraphic succession. In particular, two stops will showcase strata associated with key Late Devonian extinction events including the Frasnian-Famennian global crisis. Key discoveries made in the preparation of this field trip publication, not recorded in earlier literature, include: recognition of wave?-generated bedding and substrate channelization in styliolinid grainstone deposits of the Genundewa Limestone; discovery of a horizon in the lower Rhinestreet Shale with the chronozonally important goniatite cephalopod *Naplesites* and possibly dateable conodonts a few meters above a lag bed with Montagne Noire (MN) Zone 7 conodonts and a K-bentonite horizon presumably older than the Belpre Ash Bed of the southern Appalachian Basin (MN Zone 8); recognition of abundant phosphatic nodules in the Rhinestreet Shale cored by fish bones and scales; and discovery of a possible thin K-bentonite within the Pipe Creek Shale (= Upper Kellwasser Bed marking a global extinction-ecological disturbance event) allowing for potential absolute dating of that event.

This field trip differs from earlier ones in that it does not focus on any particular stratigraphic unit or problem. Rather we will provide somewhat of an ascending "Cook's tour" of units proceeding southwestward across Erie County from the lower medial Hamilton Group (STOP 1) to the Frasnian-Famennian mass extinction (STOP 7). This trip is, however, complimentary to our Sunday field trip (this volume) which focuses in more detail on the origin of the distinctive North Evans Limestone/Genundewa Limestone that we will see at STOP 2. For the benefit of participants, several of the newly-generated schematics are presented both in the Saturday and Sunday fieldtrip papers. In the tradition of NYSGA serving as a teaching forum, this fieldtrip is presented as "work in progress" and is designed to review controversies and ongoing questions. Although material covered is somewhat of an eclectic mix, this should lead to a broader array of issues to be addressed. Moreover, we have discovered several new features in several Late Devonian (Frasnian-Famennian) sections during field trip preparation that demand further investigation. These discoveries, discussed below, include the recognition of small-scale hummocky cross-stratification and substrate channelization in the Genundewa Limestone, the finding of a new K-bentonite bed in the Pipe Creek Shale, the discovery of a geochronologically important occurrence of goniatites in the lower part of the Rhinestreet Formation, and the discovery of abundant phosphatic nodules containing fish bones and spines, also in the lower part of the Rhinestreet Formation. Thematic aspects of this trip will be: firstly, to examine evidence of current activity and bottom erosion in several basinal shale-dominated units intuitively viewed as "deep water" and "basinal," secondly, to look at distinctive facies and event horizons at different levels, and thirdly, to look at levels linked to key Late Devonian bioevents.

GEOLOGIC SETTING

During the late Middle Devonian western New York was located in the southern hemisphere tropics or subtropics and covered by an epicontinental sea (Scotese, 1990). Strata seen on this fieldtrip accumulated on the northern margin of a subsiding foreland basin that periodically expanded and deepened during phases of oblique collisional overthrusting (tectophases) associated with the ongoing Acadian Orogeny (Ettensohn, 1987, 1998; Fig. 1). The most pronounced thrust loading event (tectophase three) coincided with the onset of the deposition of the Genesee Group; this flexural drowning event was also largely coincident with a major rise in sea level (within T-R cycle IIa of Johnson et al. 1985). In west-central New York this deepening is expressed by lithologic change from shelf carbonates of the Tully Formation into basinal black shale deposits of the Geneseo Formation (Heckel, 1973; Baird and Brett, 2003; Baird et al., 2003). In western New York the Tully Formation is absent due to erosional/corrosional processes, and progressively younger divisions of the Genesee Group: Geneseo Formation with Leicester Pyrite at its base, Penn Yan Formation, and condensed North Evans/Genundewa deposits, are observed to successively onlap a major regional disconformity (Taghanic Unconformity) in a westward direction (Fig. 2). This disconformity, separating fossiliferous neritic facies of the late Middle Devonian (Late Givetian) Windom Member of the Hamilton Group from overlying dysoxic, pelagic limestone and lag debris (North Evans/Genundewa deposits), will be seen at STOP 2. In west-central New York localities, the gradational transgressive change from Tully carbonate facies into black shale of the Geneseo Formation coincides with the beginning of the upper Givetian substage of the late Middle Devonian (Huddle, 1981; Kirchgasser et al. 1989; R.T. Becker, personal comm., 2006). Proceeding westward along the Taghanic disconformity, the ages of the onlapping black shale deposits become progressively younger into eastern Erie County; this reflects the regional flexural-eustatic Taghanic event (Kirchgasser et al., 1989; Baird and Brett, 1986). A younger erosion surface, associated with the North Evans Limestone conodont – bone lag below the Genundewa Limestone, oversteps the Taghanic disconformity in Erie County, thus merging the two discontinuities into a composite unconformity (Fig. 2). Hence, at STOP 2, the Late Devonian (early Lower Frasnian) North Evans Limestone rests directly on late Middle Devonian (middle Givetian, *ansatus* Zone) shales of the Windom Member (Moscow Formation; Hamilton Group) with several conodont chronozones missing or whose representatives were reworked and transported. The effective chronostratigraphic (taphonomic) age of the North Evans is early Frasnian upper MN Zone 2 (Figs. 2-3).

Acadian orogenic uplift in New England and the central Atlantic region was associated with progradational developmemt of the Catskill Delta Complex which filled the foreland basin from east to west (Woodrow and Sevon, 1985; see Kirchgasser et al., 1997). Catskill Delta



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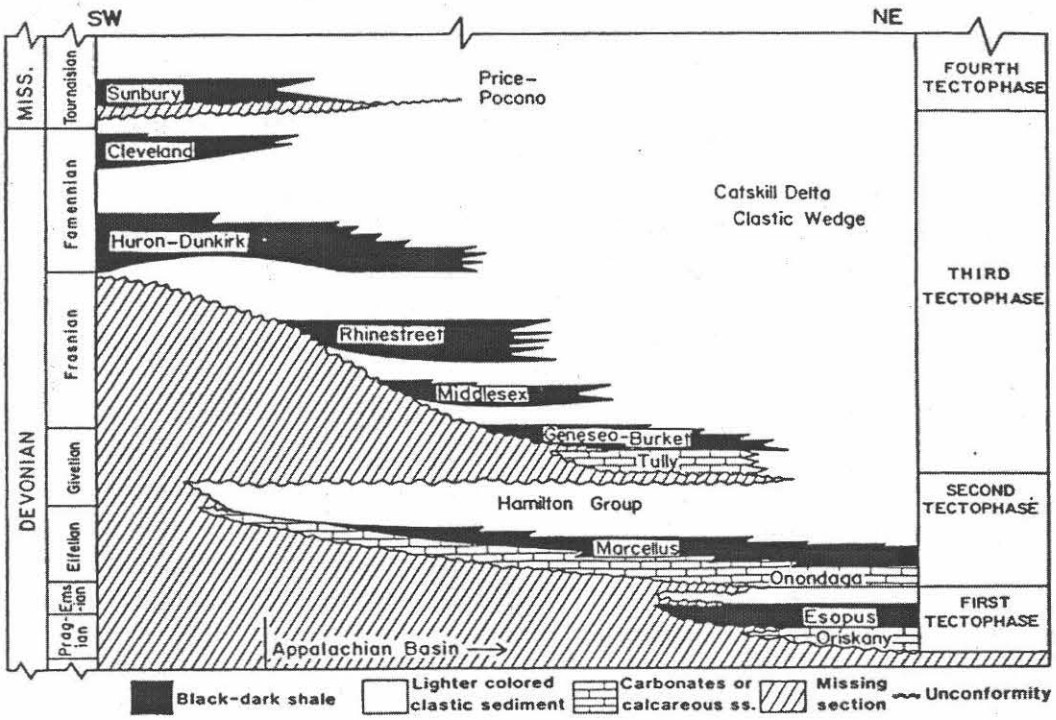

B

Figure 1. A. Idealized depositional model of the Catskill Delta complex (from Broadhead et al., 1982). **B.** Composite stratigraphic section from east-central New York to north-central Ohio in the northern Appalachian Basin showing distribution in time of pre-tectophase unconformities and unconformity-bounded flexural sequences of black shales and coarser clastic sediments attributed to four Acadian tectophases. Note progressive southwestward (cratonward) basin migration of successive black shale strata (from Ettensohn, 1994).

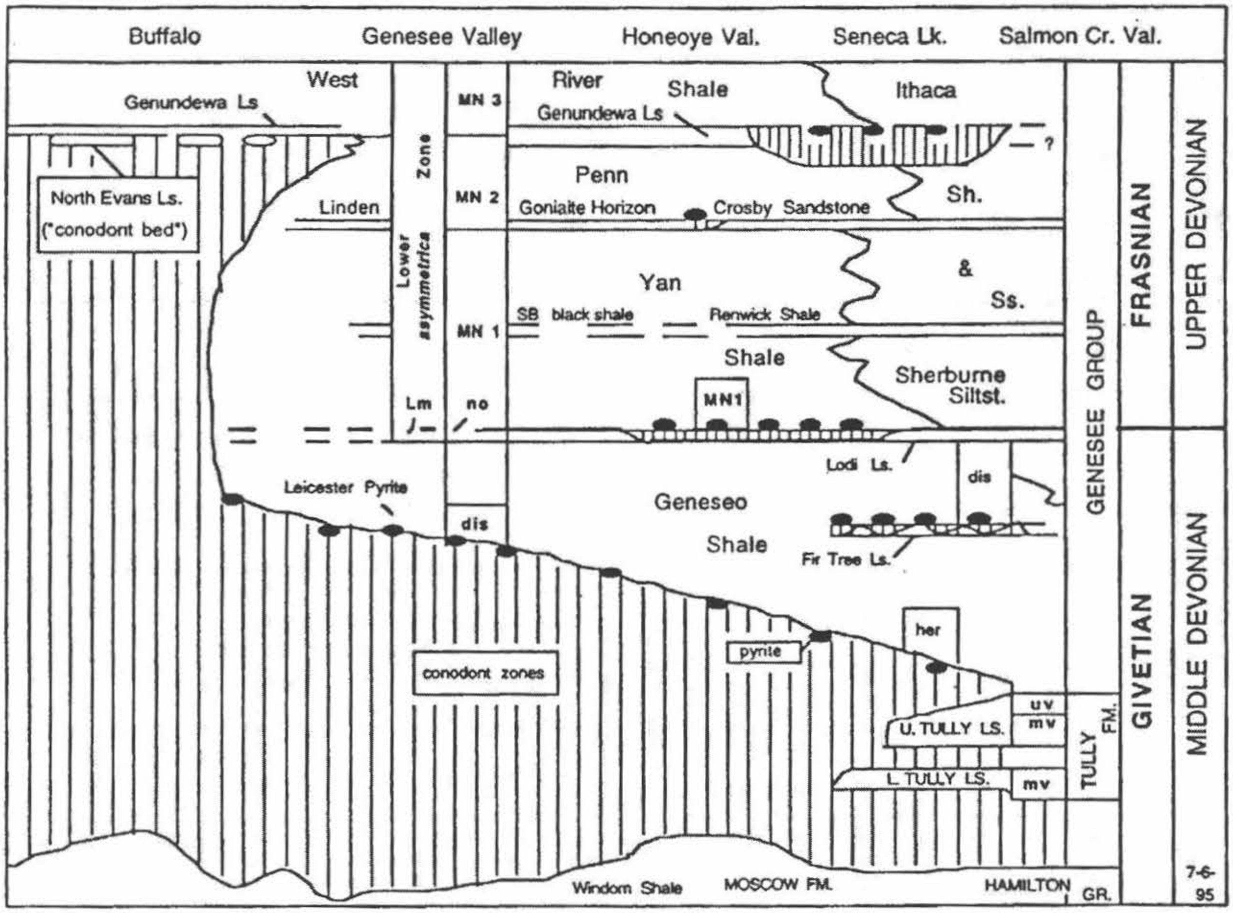

Figure 2. Generalized chronostratigraphic cross section of lower Genesee Group and subjacent Moscow Formation (Hamilton Group, Windom Shale Member). Large hiatus below Genesee Group marks position of compound Taghanic Unconformity, Genesee onlap succession and sub-Genundewa unconformity. The pre-Tully erosion- Hamilton erosion surface marks a major sequence and tectophase (III) boundary. The lenses of detrital Leicester Pyrite are derived from this erosion but were deposited through a long period of diachronous onlap of Genesee black muds from this discontinuity during the Taghanic transgression. The locally beveled beds with pyrite and fish debris include condensed styliolinid limestones and nodules (Fir Tree, Lodi, Abbey, Linden, Crosby, Genundewa) associated with surfaces of maximum sediment starvation formed during pulses of sea level rise. In this report, these horizons have been traced to the most highly condensed westernmost sections. The North Evans Limestone (conodont bed) in the Buffalo area in western Erie County is a lag deposit of crinoid, fish and conodont debris that accumulated in shallow water over the peripheral bulge at the west margin of the basin where of the gap of the compound unconformity is greatest. Lenses of North Evans debris with *ansatus* Zone (Middle *varcus*) to upper MN Zone 2 conodonts and Frasnian goniatites (*Koenenites*) are traceable beneath the sub-upper Genundewa disconformity as far east as the Genesee Valley. (From Kirchgasser, Brett and Baird, 1997, fig. 7). See Fig. 3 for names of conodont zones.

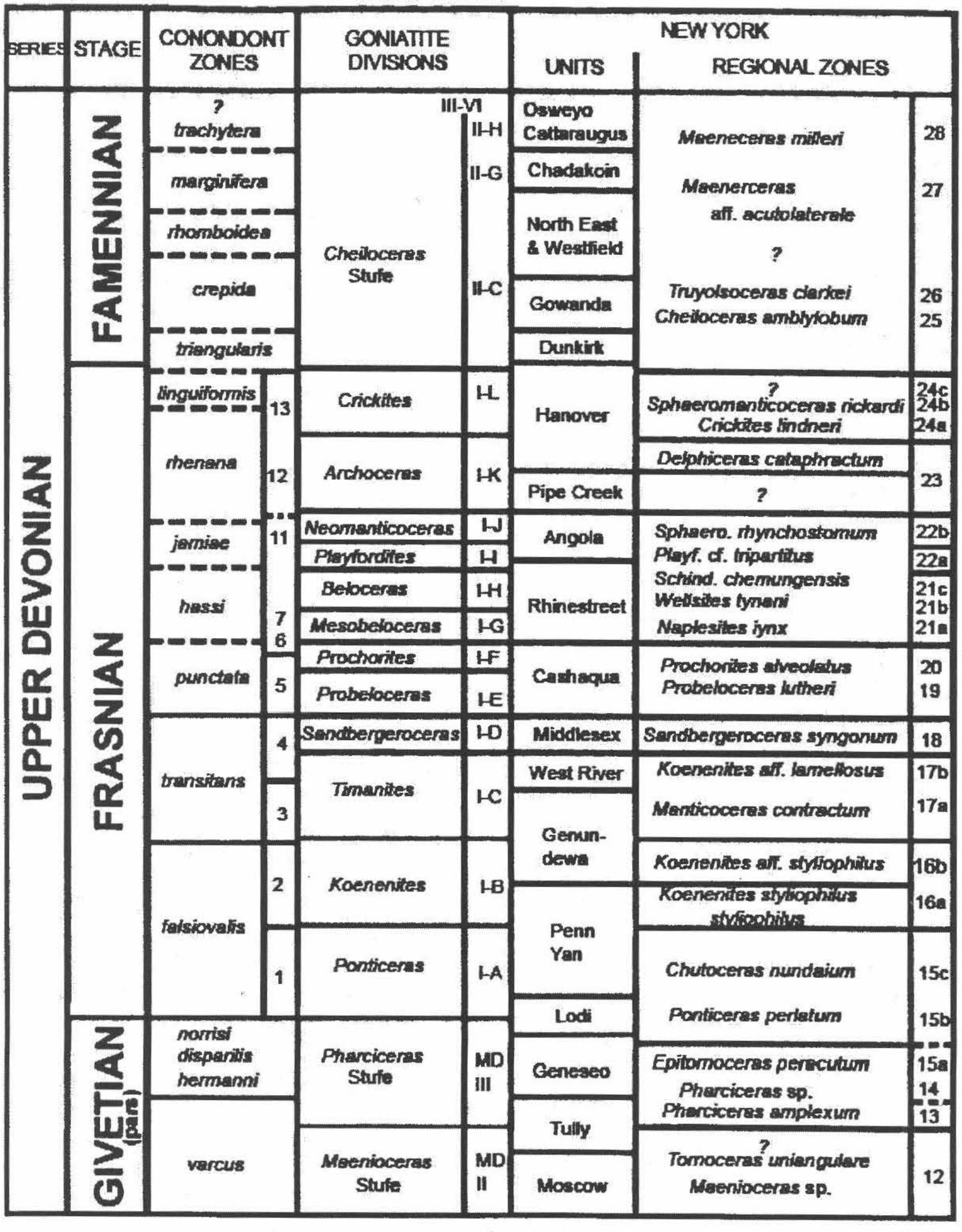

Figure 3. Late Devonian (Givetian, Frasnian, and Famennian) stratigraphic succession in New York State showing alignment to international conodont zones (Standard and Montagne Noire [1-13]) and goniatite cephalopod divisions and New York regional goniatite zones (12 to 28). MN Zone assignments follow Kirchgasser and Klapper (1992), Kirchgasser (1994), and Klapper et al. (1995). Modified from House and Kirchgasser (1993 and in press).

progradation began in earnest during deposition of the Middle Devonian Hamilton Group following the onset of the second collisional tectophase, but accelerated significantly during the third tectophase (Ettensohn, 1998). Not only do strata above the Taghanic disconformity thicken greatly to the east, but they also grade spectrally eastward and shoreward into variably fossiliferous neritic facies which are typically much coarser (Fig. 2).

Generally, the units seen on this field trip represent very fine grained detrital facies of the Catskill Delta representing deposition in deeper water slope and basin settings both at and beyond the delta margin (Fig. 2). Units such as the Rhinestreet Formation of the West Falls Group (Late Devonian, Frasnian) are typically expressed along Lake Erie as organic-rich, fissile to massive, black shale facies recording near-anoxia during phases of transgressive highstand (Rhoads and Morse, 1971; Murphy et al., 2000). However, thinner intervals of gray-green, typically bioturbated shale occur within the Rhinestreet (see STOPS 3-5). Shale of this type thickens greatly eastward toward the depocenter and sediment source. Black shale units, including the Rhinestreet, typically split into eastwardly splaying black shale tongues (Fig. 1). At STOPS 3 and 5 we will see that some of the gray-green shale shows little evidence of bioturbation and may be turbiditic or hemipelagic in origin. The black shale facies is often laminated, but actually, typically displays small flattened burrows, indicating the bottom setting was not exclusively anoxic.

CURRENT ACTIVITY ALONG DEVONIAN SLOPE AND BASIN SUBSTRATES: GENESIS OF BASINAL DISCONTINUITY SURFACES, CHANNELS, AND SMALL-SCALE HUMMOCKY CROSS-STRATIFICATION

Baird and Brett (1986) undertook a regional study of the Leicester Pyrite, a bone/conodont-rich detrital pyrite deposit associated with the Taghanic Unconformity and onlap surface (basal Geneseo black shale contact) from Ontario County westward into eastern Erie County (Fig. 2). Very coarse detrital pyrite was found to occur in laterally disconnected lenses up to 30 cm in thickness along the contact and also slightly above the contact within the black shale. We interpreted the Leicester to be a coarse lag deposit comprised of pyrite derived from the underlying Windom Member and moved by currents along the Taghanic Unconformity surface within erosional channels cut into the underlying shale (Baird and Brett, 1986). The fact that some lenses occur above the contact within the Geneseo Shale demonstrated contemporaneous deposition of the detrital pyrite with the onlapping black muds of the Geneseo; hence, the pyrite and other insolubles were swept downslope over the leading edge of the onlapping organic-rich sediment within the setting of black mud accumulation. This model is further supported by the chronostratigraphic diachroneity of the basal Geneseo Shale (Fig. 2). The development of a channeled regional slope is suggested by the occurrence of the Leicester Pyrite in discrete lenses which could be channel fills. Unfortunately, owing to the direction of our field excursion, we will not encounter true Leicester in our sections. However, at STOP 1, we will see discrete erosional runnels, probable furrows *sensu* Flood (1983), associated with a discontinuity in the lower part of the Levanna Member, Skaneateles Formation within the Middle Devonian Hamilton Group (Baird and Brett, 1991; Fig. 4). The long cutbank section showing this discontinuity shows a series of small shale-filled channels which are crossed transversely (Fig. 5). Some of the larger channels contain lags of small brachiopods and fish bones in their axes.

The channels at STOP 1 are probably smaller versions of those associated with deposition of the Leicester. More recently Schieber (1994; 1998) described numerous occurrences of arcuate erosion surfaces within black basinal facies of the Late Devonian Chattanooga and Ohio shales; these undulating, concave scours reveal truncation of beds both between outcrops and within outcrops. Such contacts may be channeled in a manner similar to the Levanna examples at STOP 1.

Baird and Brett (1986, 1991) discussed a variety of mechanisms to produce coarse tractional lags in black shale settings in the context of a basinal, deeper-water setting interpreted for such facies. Processes including: deep-storm wave impingement, bottom current processes, and internal waves were examined as mechanisms capable of moving coarse particles at depth. We tentatively settled on a model of internal wave-shoaling against a sloped basin substrate as a possible traction mechanism; in this scenario by internal waves generated along the pycnocline within the water column, eventually shoal against the basin margin slope resulting in erosion and sediment traction (Baird and Brett, 1991). This fits into the black shale onlap scenario in that this erosion occurs on the Taghanic Unconformity slope prior to slope burial by black mud; as water deepens, owing to sea level-rise and/or flexural subsidence, the zone of pycnoclinal erosion continually migrates westward in the upslope direction ahead of black mud onlap which takes place within a lower energy, lower dysoxic substrate regime below the pycnocline (Baird and Brett, 1986, 1991). Westward flexural basin expansion during Geneseo Shale deposition would account for east-to-west slope drowning and conveyor belt-type pycnocline migration and subsequent sediment onlap along a 100 km lateral distance across western New York. Calcareous fossils and diagenetic carbonate debris reworked from the underlying Windom Shale on the east-sloping, sediment-starved, Taghanic erosional ramp would start out as calcareous lag material in a shallower water wave-influenced, oxygenated regime. Subsequent slope drowning with consequent overspread of dysoxic water below the pycnocline was believed to explain the dissolution and transformation of the lag material to a residual placer of pyrite and other insolubles. Since the zone of pycnocline impingement was always upslope from the mud onlap limit during Geneseo time, the basal Geneseo lag would always be made up of insoluble material (Baird and Brett, 1986).

It is significant that the Leicester example is not isolated; coarse insoluble lags associated with Devonian black shale-roofed unconformities have been examined elsewhere (see summary in Baird and Brett, 1991; Schieber, 1994, 1998; Brett et al., 2003). Moreover, in the Rhinestreet Formation (STOPS 3-5) and in the upper part of the Hanover Formation (STOP 7), we will observe numerous gray-black shale alternations where thin black layers, some only millimeters-thick, rest sharply on gray shale units. Some of these contacts display thin lags of reworked wire-like, pyritic burrows, flattened goniatites, some with pyritized sutures, and geopetally pyrite-filled, spherical cysts of the algal taxon *Tasmanites* (Schieber and Baird, 2001). Lags flooring these thin black layers are much thinner and finer than those associated with the major contacts.

Juergen Schieber, by contrast, argues for a shallower water origin of these discontinuities and associated black shale facies based on his work on the Ohio, Chattanooga, and New Albany shales (Schieber, 1994, 1998). Coarse tractional siltstones, sandstones, and shell beds within the very condensed black, Chattanooga Shale are interpreted by him as being the result of storm wave impingement. Calculations of orbital wave velocities accounting for coarse sand and

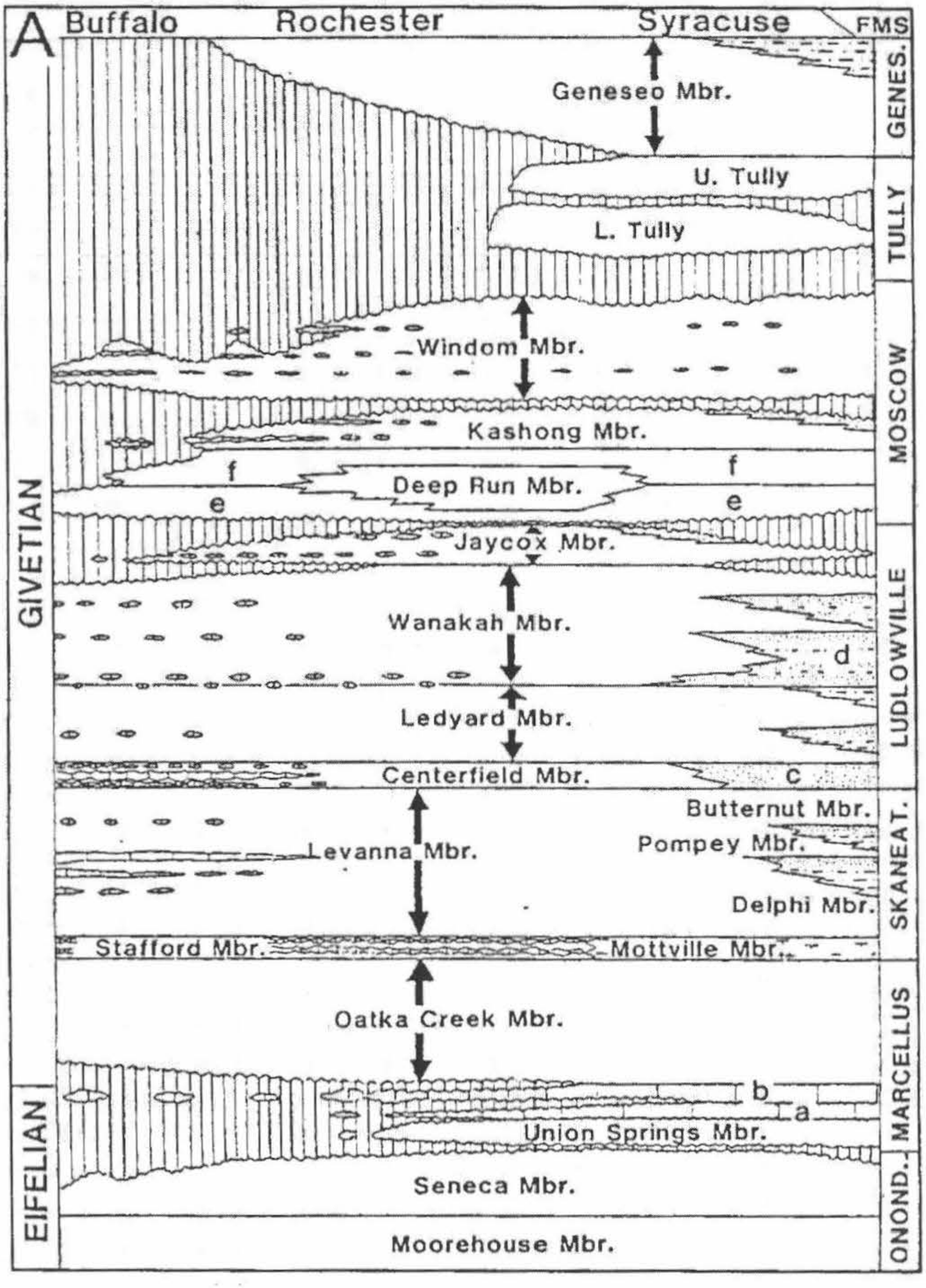

Figure 4. Time-rock diagram of upper Eifelian to upper Givetian succession in central and western New York showing the position and distribution of Ledyard Shale Member of the Ludlowville Formation (Hamilton Group), the unit to be seen at STOP 1. (From Brett and Baird, 1996 after Brett and Baird, 1994; also Kirchgasser, Brett and Baird, 1997, fig. 5).

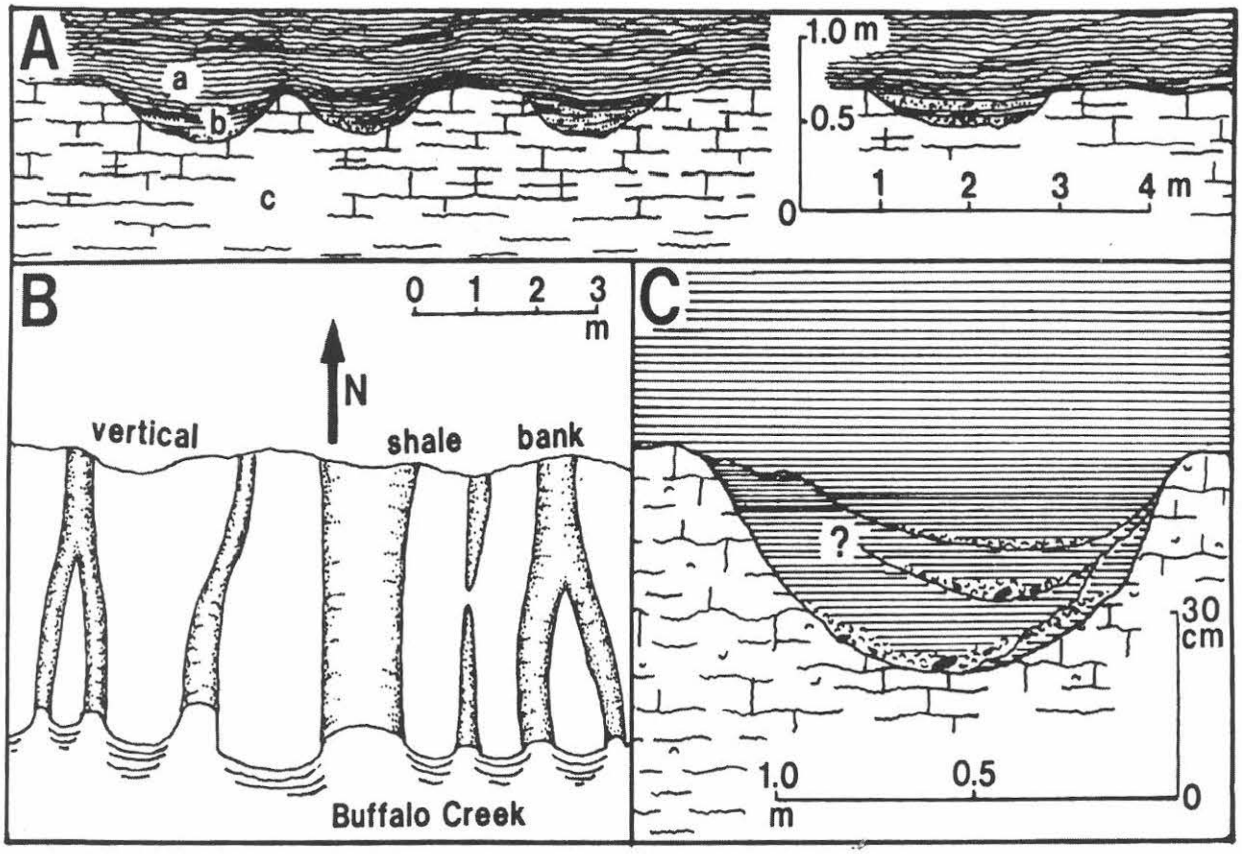

Figure 5. Submarine discontinuity within Levanna Shale Member on Buffalo Creek opposite Charles Burchfield Museum and Nature Center, upstream from (east of) Union Road overpass (see STOP 1). A, Along-bank profile of a series of erosional runnels (troughs) cut into calcareous shale that are filled with 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 filling 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 flattened rhynchonellids, *Styliolina*, and palynomorphs; b, brown-black shale filling in troughs with associated scoured contacts and mixed brachiopod, trilobite, and fish bone lag debris; c, calcareous gray to dark gray blocky to chippy mudstone with *Ambocoelia, Devonochonetes,* and *Eldredgops* (from Baird and Brett, 1991; Baird et al., 1999).

detrital pyrite transport, as well as the scouring of consolidated shale, yielded velocities in excess of 150 cm/sec. suggesting water paleodepths of as little as 10 meters (Schieber, 1994, 1998). These contrasting models revive the long-standing "black shale paleodepth controversy" that has long existed. Very coarse, "grit- grade" pararipples and sheet sands occur within black shale facies of the Dowelltown Member of the Chattanooga Shale west of Nashville, Tennessee [G.C. Baird and A.J. Bartholomew (SUNY-New Paltz), unpublished observations]. Not only does cross laminated, grit-grade, quartz and phosphatic sand rest on the sub-Chattanooga disconformity, but it also occurs at numerous overlying levels within the Dowelltown. Clearly, Schieber's storm model appears to have some credibility in explaining black shale features on the Nashville Dome. The coarse-grained Dowelltown beds, as well as the Leicester and its analogs, pose a key question; how does one reconcile "basinal," widely-distributed, and near-anoxic, organic-rich shale with evidence for high energy current activity? Are these units truly shallow with a pycnocline maintained just below the sea surface, to be disrupted intermittently by storms? Does a shallow-to-deep spectrum of Devonian black shale types exist within the Appalachian Basin and beyond? If shallow Devonian black shales exist, are these maintained by enormous surface productivity or by purely physical mechanisms? Until a good actualistic (modern) example of Leicester-type deposits is found forming, they will remain an intriguing enigma, but also a key insight in our overall understanding of sedimentary processes in the rock record.

The model of regional bed onlap and deeper-water pycnoclinal erosion can also be applied to the younger North Evans bone/conodont lag bed flooring the Genundewa Limestone (see STOP 2; Fig. 6). As will be discussed more fully on the Sunday (B 3) field trip (this volume), the Genundewa Formation of the Genesee Group is a condensed pelagic limestone unit almost entirely composed of the problematic conical microfossil *Styliolina fissurella* [Order Dacryoconarida Fisher, 1962; see Lindemann (2002) for summary of dacryoconarid classification]. It appears to mark a transgression from a eustatic (or tectonically induced) lowstand event recorded by a regional unconformity marking the top of the Penn Yan Shale and an associated coarse lag unit known as the North Evans Limestone bone/conodont bed ("Conodont Bed" of Hinde, 1879; Fig. 2). Recent work (see fieldtrip B3, this volume) shows that dysoxic Genundewa styliolinid carbonate grainstone facies appear to onlap the unconformity surface to the northwest (inferred foreland basin margin) to the point of near bed-extinction in eastern Erie County. The North Evans Limestone, similar to the Leicester Pyrite, is very coarse; it contains reworked fish teeth, bones, spines, scales, abundant pelmatozoan debris, pyritized mollusks, including early whorls of goniatites, and a rich and famous concentration of conodonts, an amalgamation of late Givetian and early Frasnian elements spanning several conodont zones (Baird and Brett, 1982; Brett and Baird, 1990; Bryant, 1921; Huddle, 1974, 1981; Kirchgasser, 1994; Hussakoff and Bryant, 1918; Over et al., 1999). The important difference between the North Evans and Leicester is the dominantly carbonate nature of the former and the overwhelmingly insoluble character of the latter. We believe that the North Evans lag accumulated under conditions that were less dysoxic and, by implications, shallower than those applying to the Leicester. In essence, the North Evans lag is what the Leicester may have looked like at an upslope position on the Taghanic ramp prior to its subsequent dissolution at greater depth (Brett and Baird, 1982; Baird and Brett, 1986). At STOP 2 we will see a variant of the North Evans that more closely resembles the Leicester owing to the presence of a localized black shale unit that closely overlies it at the southwesternmost limit of its outcrop by Lake Erie (Fig. 6). In this area the North Evans lag debris was exposed to more severe dysoxic conditions than elsewhere; it is distinctly more pyrite-rich and is greatly reduced in volume.

Above the North Evans Limestone at STOP 2 is a 25 cm-thick black shale unit that is rich in flattened *Styliolina* shells which, in turn, is overlain by the Genundewa Limestone, a 30 cm-thick layer of dark gray, styliolinid grainstone-packstone carbonate that is regionally widespread (Figs. 1, 6). *Styliolina fissurella* is a problematic 1-2 mm-long calcareous conical shell of uncertain affinities. It was originally described erroneously from flattened material, hence the specific name "*fissurella*" (see Hall, 1843, 1879). Subsequent workers placed these organisms in a the Nashville Dome. The coarse-grained Dowelltown beds, as well as the Leicester and its analogs, pose a key question; how does one reconcile "basinal," widely-distributed, and near-anoxic, organic-rich shale with evidence for high energy current activity? Are these units truly shallow with a pycnocline maintained just below the sea surface, to be disrupted intermittently by storms? Does a shallow-to-deep spectrum of Devonian black shale types exist within the Appalachian Basin and beyond? If shallow Devonian black shales exist, are these maintained by enormous surface productivity or by purely physical mechanisms? Until a good actualistic (modern) example of Leicester-type deposits is found forming, they will remain an intriguing enigma, but also a key insight in our overall understanding of sedimentary processes in the rock record.

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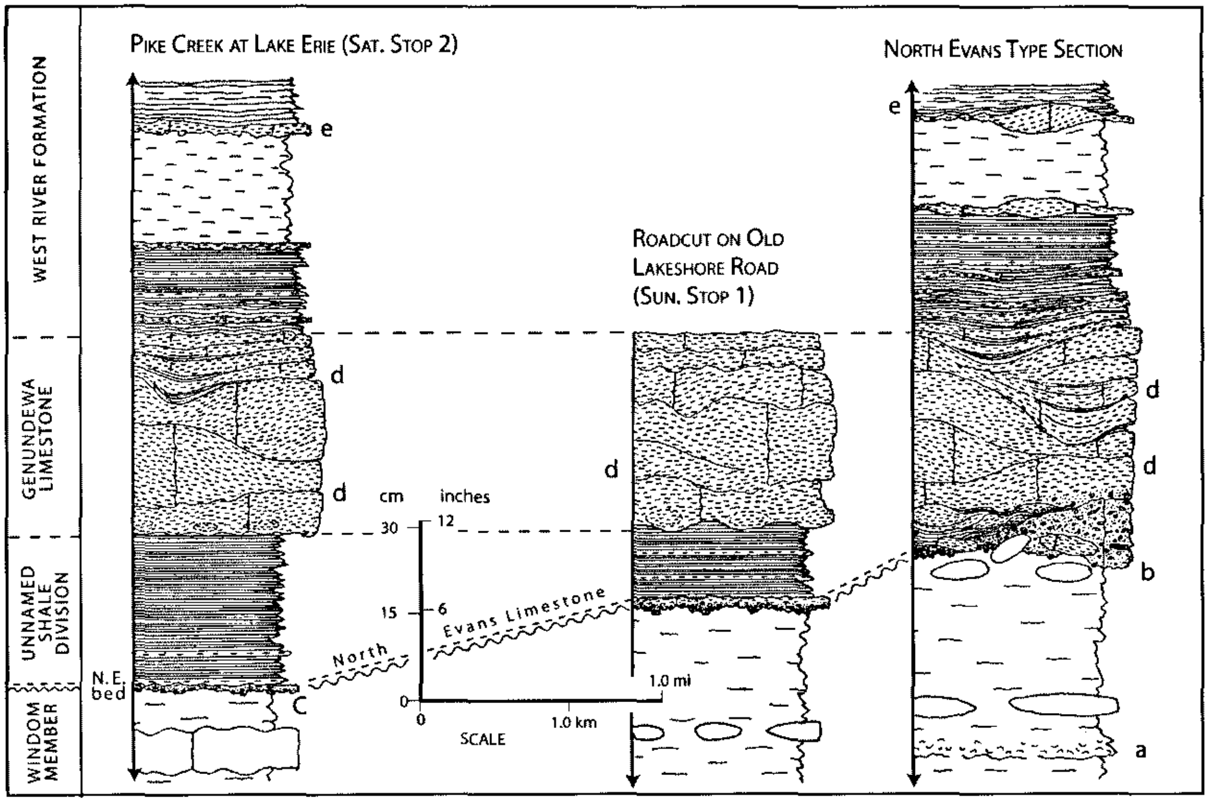

Figure 6. Genundewa Limestone and associated units in the vicinity of Lake Erie. Note prominent westward thinning of the North Evans lag deposit coupled with westward appearance of an unnamed black shale unit that seperates the North Evans from the overlying Genundewa (see discussion in text). Also visible is distinctive pinching and swelling of beds with associated localized channeling within the Genundewa. Lettered units include: a, Amsdell Bed of Windom Member yielding abundant *Emanuella praeumbona*; b, thick, pelmatozoan-rich subfacies of the North Evans Limestone; c, thin, pyrite-rich subfacies of the North Evans Limestone; d, Genundewa styliolinid grainstone-packstone carbonate facies; e, thin, lenticular, styliolinid limestone bed in the West River Formation yielding glauconite and abundant conodonts; this is sample-bed USGS 8122-SD Fall Brook, Geneseo (Livingston County) of Huddle (1981).

Acanthoclymenia, and *Tornoceras* , crinoid ossicles, and wood debris, can be found (see Kirchgasser et al., 1994, fig. 7, for sketches of the goniatites). The biota is of low diversity and suggests a dysoxic stressed environment, particularly, when compared to the rich, high diversity benthic fauna of the Tichenor Limestone, a carbonate unit of comparable thickness 6 meters below the Genundewa at STOP 2. Devonian styliolinid limestone facies is also known from

European and North African sections where it is understood to represent condensed pelagic facies which accumulated in sediment-starved settings on the order of tens to hundreds of meters of water depth (see Tucker and Kendall, 1973; Tucker, 1974; Bandel, 1974). The Genundewa compares most closely to the "cephalopodenkalk" (cephalopod limestone) facies of the German Rhenohercynian region; this carbonate accumulated on structural "highs" (schwellen) where styliolines, goniatites, diminutive bivalves, and ostracodes accumulated in a sediment-starved regime (Tucker, 1974; House and Kirchgasser, 1993). Basins between these swells received contemporaneous accumulation of thick shale units where turbiditic facies yield mainly ostracodes and little else. Compared to descriptions of the Rhenohercynian cephalopodenkalk, the Genundewa notably lacks micrite and is much more nearly a styliolinid grainstone (Fig. 6). However, it is locally packed with goniatite phragmocones in a manner typical of many cephalopodenkalk units (see Sunday Field Trip B3: STOP 4, this volume).

The Genundewa in Erie County is usually massive, but when weathered, the limestone typically splits apart into nodular and flaggy beds (Sass, 1951; Baird and Brett, 1982; Brett and Baird, 1982). Nodules occur as laterally linked to separate zones of sparry styliolinid limestone surrounded by muddy styliolinid partings. Bedding in the Genundewa is usually laminar with some evidence of bioturbation. Preparation for the present field trip led to discovery of cross stratification within the Genundewa along the Lake Erie shore bluffs southwest of Pike Creek (STOP 2) and nearby on Eighteenmile Creek (Fig. 6). Several stacked sets of low angle cross stratified styliolinid grainstone can be seen with distinct thickening and thinning of beds in the cleaner, longer sections (Fig. 6). Locally, beds are distinctly cut out where channelization has occurred. This pattern resembles small-scale hummocky cross-stratification, suggesting the influence of deep-storm wave impingement at the substrate.

LATE DEVONIAN BASINAL FACIES AND EVENT HORIZONS

A sequence of alternating black and sparsely fossiliferous gray shale units (West River Formation-through-Dunkirk Formation) characterize the Late Devonian Frasnian and basal Famennian succession in southern Erie County. Inferred paleoenvironments range from nearly anoxic for portions of black shale units to more broadly dysoxic for the gray facies. No unit in this succession yields a significant benthos, though, as we will see, some strata yield a variety of pelagic taxa.

On the Lake Erie shore southwest of Pike Creek (STOP 3), Sturgeon Point (STOP 4), and on Eighteenmile Creek southeast of North Evans (STOP 5), we will examine the lower part of the black Rhinestreet Formation, a major division of the West Falls Group (Fig. 7). Within the larger black shale interval is a 2.2-2.4 m-thick interval of predominantly gray shale with thin "pinstripe" black shale bands at several levels (Fig. 7). This is well exposed at the lakeshore sections (STOPS 3 and 4) where a basal turbiditic? siltstone bed marks the base of the interval. Although some bioturbation can be seen in the gray shale, much of it is fine grained, conchoidal "satin shale," suggestive of turbiditic or hemipelagic origin. Hence the gray shale complex appears to be the distal "toe" of prodelta sediments within the basin. At STOPS 3 and 5 a 1.0 cm-thick K-bentonite can be seen about one meter below the gray shale unit (Fig. 7). The pyroclastic character of this bed is revealed by an abundance of mica-rich clay which is suffused with diagenetic pyrite. This K-bentonite was earlier reported by Levin and Kirchgasser (1994) to be the Belpre Ash of Tennessee, but it now appears to be one conodont chronozone older (see STOP 3 discussion).

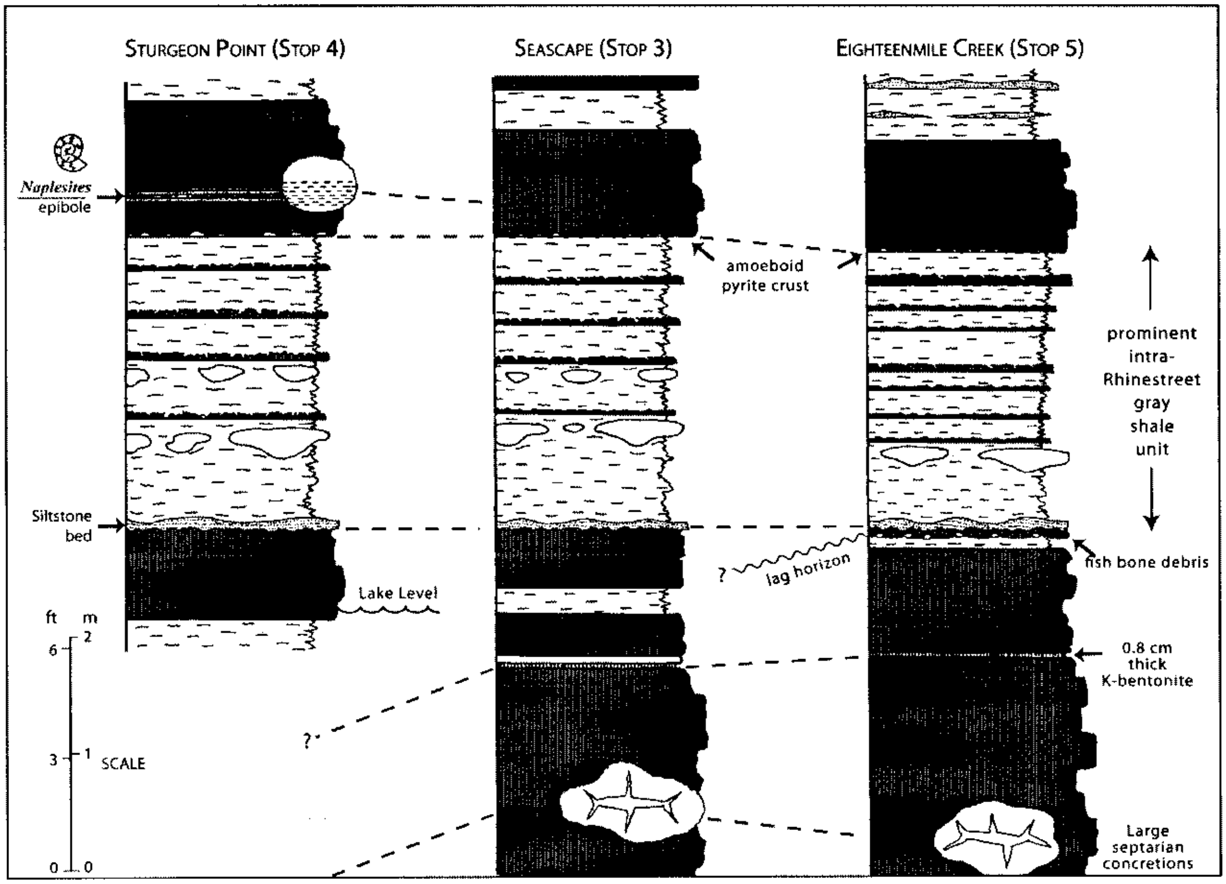

Figure 7. Unnamed gray shale division within the lower part of the Rhinestreet Formation in the vicinity of Lake Erie in southern Erie County. Note the numerous gray-green shale alternations within the gray shale division, the K-bentonite bed below it, and the *Naplesites* goniatite epibole above it (see text).

Some 40 cm above the gray shale interval exposed northeast of Sturgeon Point (STOP 5) is a 4 – 8 cm-thick interval of *Styliolina*–rich black shale that yields the zonally significant goniatite *Naplesites*, previously known in New York from a few specimens (presumably from the Rhinestreet Shale) that were reported by Clarke (1898) from around Naples in Ontario County (see STOP 4 discussion). This anomalous fossil concentration (*epibole*) was fortuitously discovered this past June when we completed a survey of this stop for the field trip. The discovery of conodonts in *Styliolina* concentrations associated with the goniatites, offers an opportunity to link conodont and ammonoid chronostratigraphic information. Moreover, given the possibility that the underlying K-bentonite can be radiometrically dated, a chronostratigraphic key point can be "hammered in" at this section. The goniatites are notable for their poor, ghost-like preservation in the black shale, suggesting that the aragonite of the phragmocone may have dissolved before significant mud compaction had taken place, but that the organic periostracum survived after compaction, leading to the flattened, composite impressions of these fossils; the edges of some of the septa (suture lines) are replaced by pyrite.

The goniatite clusters include a variety of ontogenetic stages from juveniles to adults with shell diameters exceeding 100 cm.

Southwest of the Sturgeon Point marina and car park is a low, resistant bench of Rhinestreet black shale that probably overlies most of the observed bluff section northeast of the point owing to dip effects. At the top of the bench and above a band of massive septarial concretions are two closely-spaced zones of small, 0.3 – 3.5 cm- diameter phosphatic nodules which occur interspersed among larger limestone concretions. Close examination of the phosphatic nodules shows that they selectively grew around fish bones, spines, and scales. The high proportion of fish- nucleated nodules in this location suggests a time of significant water column productivity and/or sediment condensation. This horizon may be one of many similar beds yet to be recognized in the Devonian basinal succession.

The closing of the Late Devonian Frasnian stage was marked by two major episodes of ecological disruption and faunal extinction. The second of these, marking the Frasnian-Famennian boundary and associated mass extinction, was the greater crisis globally. This extinction, in part, probably explains the lower diversity and more generalized ecological character of Famennian neritic faunas seen higher in the Devonian succession in Chautauqua and Cattaraugus counties, south of the field trip area. In Europe, North Africa, and elsewhere the two extinction events are marked by black shale or black limestone beds within slope and basin successions. These are known respectively as the "lower Kellwasser Bed" and "Upper Kellwasser Bed" in the literature (see Over, 2002; Racki, 2005; Schindler, 1993; Schindler and Königshof, 1997). Recently, both the lower and Upper Kellwasser equivalent beds have been found in western New York on the basis of lithology correlated to conodont zonation (Over, 1997, 2002; Day and Over, 2002; Over et al., 1997). The lower Kellwasser event is now linked to the Pipe Creek Formation, marking the base of the Java Group; we will see this unit on the south branch of Eighteenmile Creek (STOP 6; Figs. 8, 9). The Upper Kellwasser event correlates to a black shale bed in the upper part of the Hanover Shale Formation near the top of the Java Group (Over, 1997, 2002); we will see this bed on the south branch of Eighteenmile Creek (STOP 7; Fig. 10).

The Pipe Creek Formation at STOP 6 is a 1.5 meter-thick, very hard, black shale that abruptly overlies the softer, gray Angola Formation (Fig. 8). The laminated microfacies of the Pipe Creek contrasts dramatically with a subjacent zone of gray, pyritic Angola mudstone; this 15 cm-thick mudstone interval is thoroughly penetrated by networks of pyritic burrows which can be dramatically seen through x-ray imaging (Fig. 9). The Pipe Creek can be traced regionally southwestward into Chautauqua County where it thins to approximately 0.7 meters in its westernmost section (Tesmer, 1963). To the east, it thickens to about 6 or 7 meters near Warsaw, then becomes more depositionally complicated and interbedded with turbiditic silts and sands of the underlying Nunda Formation (Baird and Jacobi, 1999). The actual ecological reorganization-extinction event, best seen in neritic facies, is cryptic in Erie County; study of this faunal change will be the domain of work in equivalent silty-sandy facies in central New York. However, tentative discovery of a 1.2 cm-thick K-bentonite bed rich in pyroclastic micas in the Eighteenmile Creek section by the authors this past June offers the possibility that this interval can be dated radiometrically. Given that the Pipe Creek Formation is succeeded by gray, goniatite-bearing nodular shales of the basal Hanover Formation (STOP 6 description; Fig. 8), the combined radiometric date and the goniatite - conodont information will constitute a key point for the global Lower Kellwasser Event.

The Upper Kellwasser Bed is admirably exposed on the south branch of Eighteenmile Creek upstream from STOP 6 (Fig. 10). At this locality it occurs within the upper part of the Hanover Formation 2.4 meters below the base of the black lower Famennian Dunkirk Formation. Generally, the Hanover Formation is predominantly gray mudstone with a few rhythmic bundles of thin black shale units. Above the Pipe Creek and at several higher levels, this unit is spectacularly nodular with repeating bands of calcareous concretions and distinctive beds of irregular, closely crowded, beige nodules resembling calcrete (Fig. 8, unit g). Generally the Hanover yields only a low diversity fauna of ostracodes, small gastropods, bivalves, sparse goniatites, and small rugosans despite its light color, pervasive bioturbation, (including *Zoophycos*), and numerous carbonate layers. It appears to record relative sedimentary condensation under dysoxic conditions. However, compared to dysoxic – minimally oxic units in the Middle Devonian Hamilton Group (Levanna Member, Ledyard Member), the facies is markedly poorer in shelly benthos and richer in bands of small and irregular nodules. This suggests some dynamic geochemical-evolutionary changes across the Givetian and Frasnian that have yet to be identified or quantified.

The Upper Kellwasser Bed at STOP 7 is expressed as a fissile black shale unit with some silty laminations in the upper part (Fig. 10). This bed is herein designated the *Point Gratiot Bed* for an excellent exposure along Lake Erie at Point Gratiot at the southwest edge of Dunkirk, Chautauqua County. This layer, which is 15 cm-thick at Point Gratiot, is traceable eastward to the vicinity of Hornell and Canisteo in Steuben County where it is approximately 2 meters-thick (Over, 1997, 2002). At STOP 7 it displays both conformable, bioturbated lower and upper contacts and is 30 cm-thick (Fig. 10). At Point Gratiot and at Beaver Meadow Creek at Java Village the upper part of this layer has yielded articulated fish remains. At Beaver Meadow Creek, *Spathiocaris*, a probable cephalopod anaptychus, is common. It is important to note that the Point Gratiot Bed does not mark the base of the Dunkirk Formation of the Canadaway Group as was indicated by Baird and Lash (1990) and Baird and Brett (1991); the Point Gratiot Bed actually marks an apparent change to finer grained, more basinal facies *within* the upper part of the Hanover Formation [see revised schematic (Fig. 11) clarifying this relationship]. Between Point Gratiot and Java Village the interval between the Upper Kellwasser Bed and the overlying Dunkirk thickens from 15 cm to 7 meters with addition of numerous alternating black and gray-green shale beds (Fig. 11). The occurrence of reworked pyrite in the form of wire-like detrital burrow fragments at the bases of the black Dunkirk Shale and underlying upper Hanover black bands indicates that these contacts are of erosional character; some of the southwestward thinning of the upper Hanover is apparently due to collective overstep at such contacts (Baird and Lash, 1990).

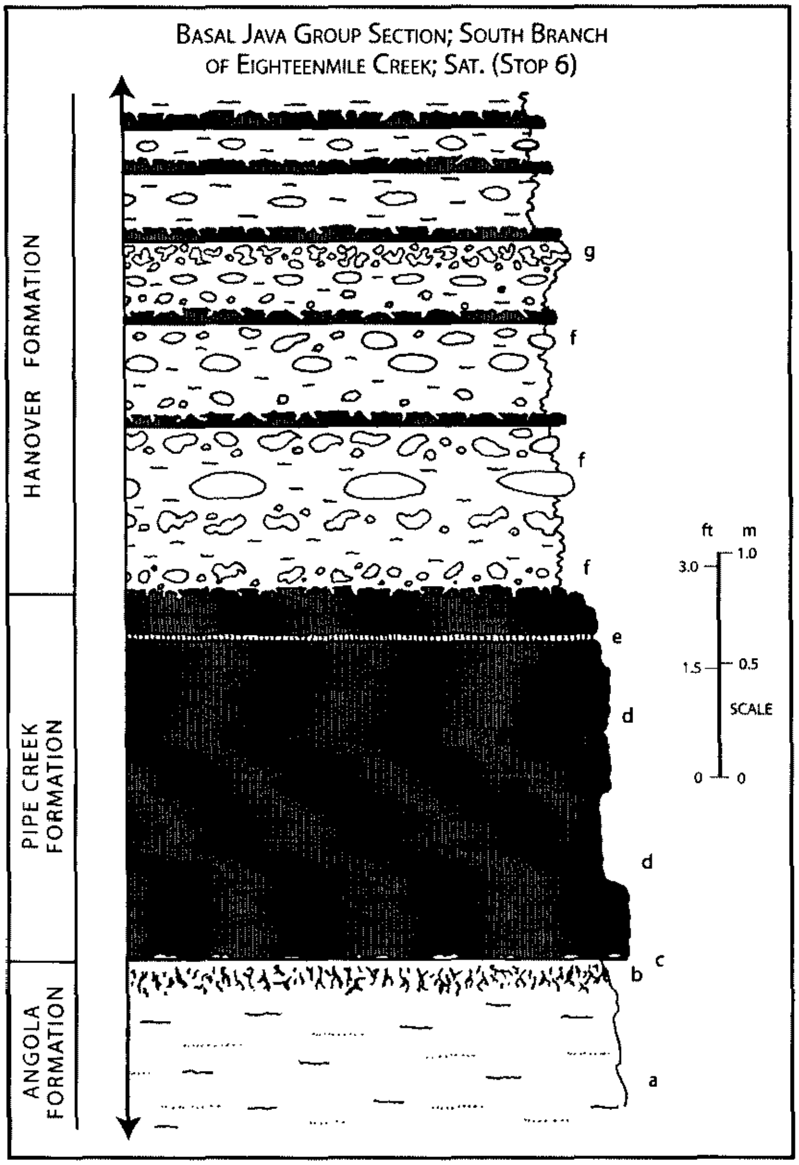

Figure 8. Pipe Creek Formation ("Lower Kellwasser Bed") and synjacent units capping major waterfall on the south branch of Eighteenmile Creek (STOP 6). Lettered units include: a, silty, gray mudstone deposits of topmost Angola Shale; b, 15 cm-thick diagenetic zone immediately beneath Pipe Creek Shale; rims of uncompacted burrows are selectively pyritized with center voids filled with sparry calcite; c, very sharp base of Pipe Creek with patchy, amoeboid, diagenetic pyrite, but no reworked pyrite (see also Figure 9); d, hard, black, well-jointed Pipe Creek Shale; e, 1.0 cm-thick gray clay bed yielding pyroclastic mica; f, basal, gray Hanover Shale characterized by profuse development of calcareous concretions and by the presence of thin black shale bands; g, conspicuous band of closely packed, irregular concretions that has a superficial resemblance to calcrete, but which appears to be a muddy version of basinal, condensed, nodular limestone facies described by others (see text).

The Upper *linguiformis* (MN Zone 13)/Lower *triangularis* chronozonal conodont boundary and inferred Frasnian-Famennian contact (Fig. 3) is crossed near or at the top of the Upper Kellwasser Bed based on work at Point Gratiot in Dunkirk, Irish Gulf, and at Beaver Meadow Creek in Java Village (see Over, 1997, 2002). Again, the major extinction event, observed globally at this level, is cryptic in the black shale facies except for the microfossil changes. However, one of us, Jeff Over, has described a bed of shelly taxa containing earliest Famennian brachiopods and bivalves in a thin, anomalous layer only one meter above the extinction horizon near Java Village (Day and Over, 2002). This "recovery layer" sheds important clues as to the nature of macrofossil changes in western New York following the mass extinction. Moreover, new fieldwork by Over in neritic deposits at this level further east near Hornell, and work by Baird in southern Chautauqua County, is shedding light on the more visceral effects of the extinction on shelly benthos and bioturbators in lower Famennian neritic deposits.

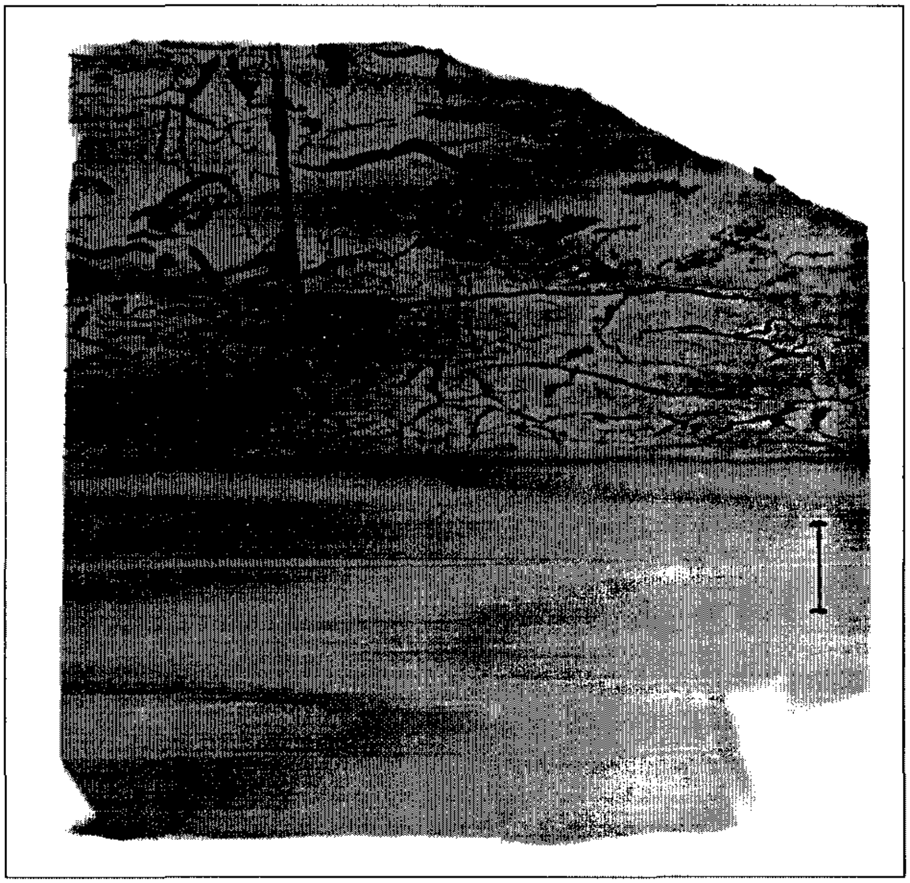

Figure 9. X-radiographic image of the contact between the Angola Shale and the overlying Pipe Creek Formation. Although this specimen is from the type Pipe Creek section near West Falls, Erie County, that contact is essentially identical to the one seen at STOP 6. Note the conspicuous vertical change from the bioturbated gray Angola into the laminated Pipe Creek lithology. Scale is 1.0 centimeter.

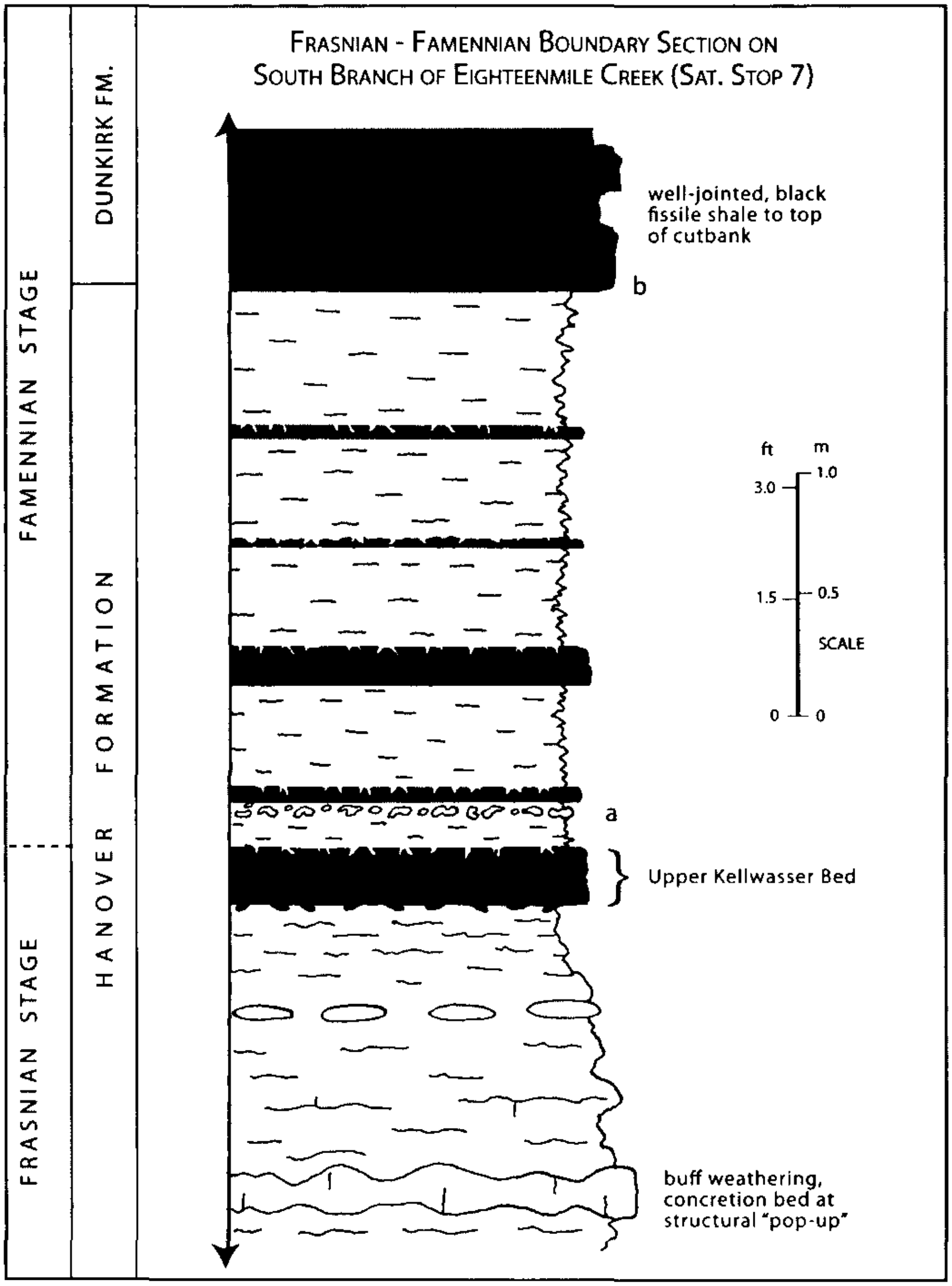

Figure 10. Frasnian-Famennian boundary horizon within the upper part of the Hanover Shale on the south branch of Eighteenmile Creek downstream from the New Oregon Road bridge near Clarksburg, Erie County. Note that the local Frasnian-Famennian boundary unit, corresponding to the global "Upper Kellwasser Bed" of chronostratigraphic literature, is herein designated the *Point Gratiot Bed* (see text). Lettered units include: a, zone of small concretions immediately below a thin, upper Hanover black shale bed; b, base of Dunkirk Shale, usually characterized in this area by a thin lag of reworked, wire-like, pyritic burrows (Baird and Lash, 1990; Baird and Brett, 1991).

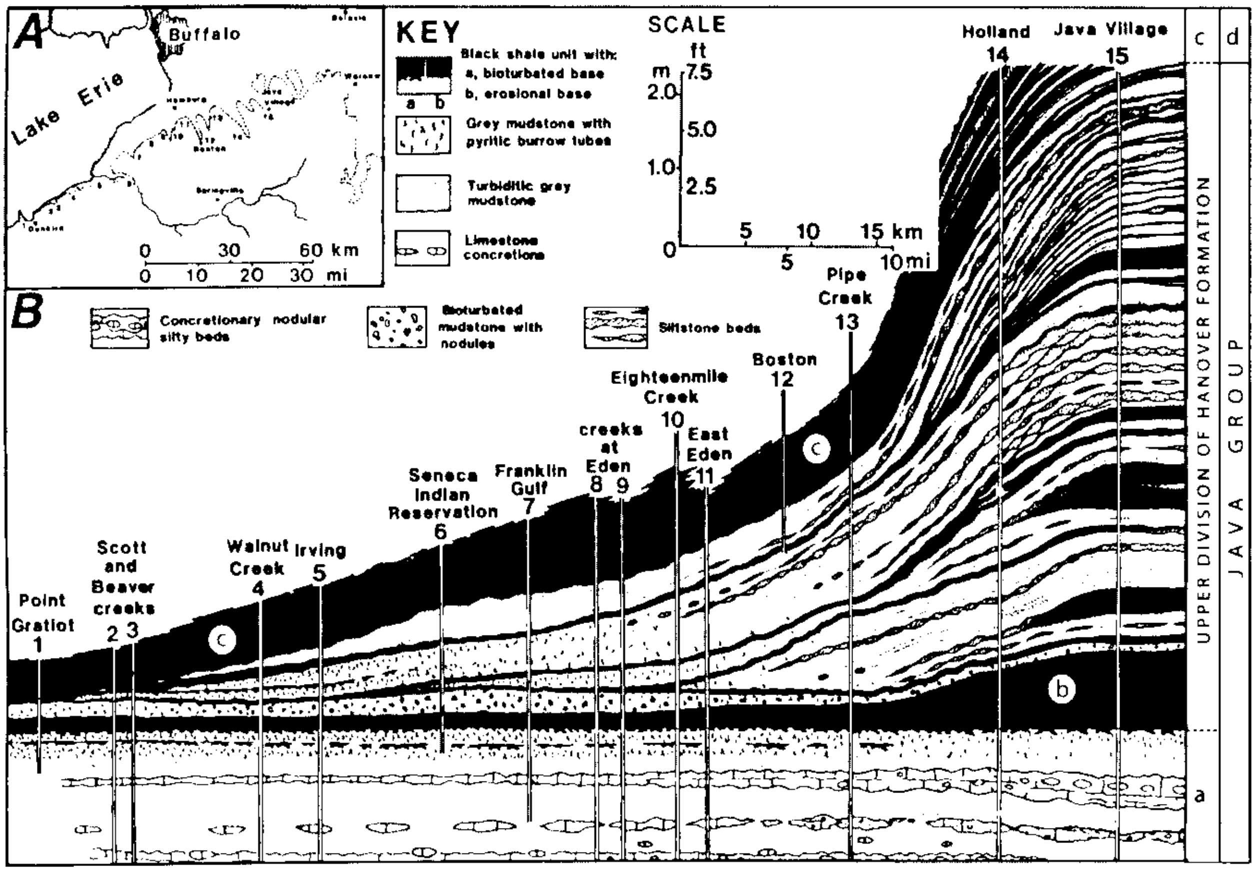

Figure 11: Regional stratigraphy of units within the upper part of the Hanover Formation across Chautauqua, Erie, and southwest Wyoming counties. Note conspicuous eastward thickening of the upper Hanover unnamed division of alternating thin gray and black shale with eastward splaying of units into a distal deltaic wedge and westward erosional overstep of underlying units by Dunkirk and upper Hanover black shale beds. Note also that the upper medial Hanover Formation below the newly named Point Gratiot Bed (= "Upper Kellwasser Bed") is notably more calcareous, bioturbated, and lighter colored than overlying units (see discussion in text). This figure is modified from Baird and Lash (1990) and Baird and Brett (1991) in that the Point Gratiot Bed is shown to be a division *within* the upper Hanover succession rather than the *base* of the Dunkirk Formation as shown in these earlier reports. Lettered units include: a, calcareous bed in upper medial part of Hanover Formation; b, Point Gratiot Bed; c, basal strata of Dunkirk Shale.

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ROADLOG AND STOP DESCRIPTIONS

Leave the Adam's Mark Hotel in downtown Buffalo and head south to the junction with I-190. Enter I-190 eastbound towards the junction of I-190 and the New York State Thruway (I-90). The road log starts at the junction of the Thruway and I-190 where we will then proceed on the Thruway in the southbound direction.

Accumulated Miles	Incremental Miles	Road log description
0.0	0.0	Enter New York State Thruway (I-90) from I-190; proceed south on I-90 towards Erie, Pa.
1.7	1.7	Exit I-90 (to the right)into feeder ramp for Route 400 Expressway. Proceed eastbound on Route 400.
2.15	0.45	Merge onto eastbound NY 400 from feeder.
3.7	1.55	Exit Route 400 expressway onto exit ramp for Union Road.
4.0	0.3	Junction with Union Road. Turn left (north) onto Union Road. Proceed north on Union Road to third light north of junction.
4.8	0.8	Turn right at red light onto Race Street opposite Indian Church.
4.9	0.1	Turn left from Race Street into parking area of Charles E. Burchfield Art Museum and Nature Center. Exit Vehicles and proceed to edge of Buffalo Creek.

STOP 1. Charles E. Burchfield Art Museum and Nature Center, Town of West Seneca (see Figs. 4, 5). When one of us (Baird) originally included this locality as a stop for a 1999 NYSGA fieldtrip (Baird et al., 1999), the present-day park was a vacant lot. The park has improved the access to this classic cutbank section along Buffalo Creek immediately upstream from Union Road. This stop is included, partly due to its relevance to our understanding of basinal submarine erosion processes, but also due to the fact that this stop was skipped owing to lack of time on the earlier fieldtrip. To reach the bank section we must wade across Buffalo Creek to the south-facing shale bank directly opposite the center. In the case of recent rains, the creek may be high, preventing our crossing. However, small, low exposures of the same channels are displayed at creek level on the south bank. These can be accessed in a high water scenario. At the bank, a prominent undulating contact is visible, separating gray to beige, calcareous mudstone below (lower division of the Levanna Member) from fissile, black shale above (medial part of Levanna Member; Figs. 4, 5). The undulating contact is an erosion surface which links eastward to a shell bed division of the Pompey Member of the Skaneateles Formation in central New York sections (see Baird et al., 2000). At this locality, conspicuous channel-like features are developed into the erosion surface which trend north-south, nearly perpendicular to the bank

trend (Fig. 5). The channels range from a few decimeters to 1.5 meters width and up to 0.5 meters in depth. The dark shale fills in channels sometimes show smaller, nested channels within larger ones (Fig. 5c) indicating that channel formation was a continuous process during formation and burial of the contact. Parallel, north-south alignment of channels suggests a prevailing current direction during channel formation (Fig. 5b). Axes of channels sometimes contain concentrations of brachiopod and trilobite debris admixed with fish teeth and bones. The channels are interpreted by Baird and Brett (1991) to be ancient examples of submarine, erosional furrows, *sensu* Flood (1983); furrows form from sustained, unidirectional currents and can form in a variety of aqueous settings. However, they are found to be particularly common at modern continental slope breaks (Flood, 1983). We believe that the lenticular outcrop profiles of Leicester- and other detrital pyrite accumulations are the lateral perspective view of similar, debris-filled channels where surrounding mudrock has dewatered and flattened out channel relief.

5.0	0.1	Return to junction of Race Street and Union Road. Turn left (south) onto Union Road and retrace route back to the NY 400 Expressway.
5.8	0.8	Westbound feeder ramp of the NY 400 Expressway. Enter it and proceed back to the New York State Thruway (I-90).
7.3	1.5	Feeder to I-90 (southbound) to Erie, Pa. Enter it by keeping to the left lane.
7.7	0.4	Merge leftward into main flow of southbound I-90.
8.35	0.65	Cross Cazenovia Creek
10.3	1.95	Lackawanna Toll Plaza of New York State Thruway. Continue straight.
11.3	1.0	Cross south branch of Smoke Creek.
12.05	0.75	Blasdell exit of Thruway. Enter exit feeder on the right.
12.35	0.3	Junction of I-90 exit feeder with Mile Strip Road; turn right (west) onto Mile Strip Road.
13.75	1.4	Mile Strip Road-NY 5 junction; turn right to enter junction rotary which will bear around to the left (south) for entry to NY 5 southbound.
13.95	0.2	Exit rotary to the right onto NY 5 (southbound) entrance ramp.
14.05	0.1	Merge southbound onto NY 5. Ford plant on the left.
14.85	0.8	Major road splits to the left for town of Hamburg. Bear right (straight) and continue on NY 5.

15.55	0.7	Lake Erie visible to the right. Low shale bluffs along shore are exposures of the Ledyard Member of the Ludlowville Formation (Middle Devonian).
15.65	0.1	Enter town of Athol Springs. We will proceed southwestward through several small connected shore communities.
19.35	3.7	NY 5 junction with Lake Shore Road; bear right onto Lakeshore Road.
22.35	3.2	Bridge over Eighteenmile Creek. Upstream cutbank to the left shows the Middle Devonian Hamilton Group divisions unconformably overlain by Late Devonian strata of the Genesee Group.
22.65	0.1	Fishermen parking area below pull-off to the left immediately south of the Lake Shore Road bridge over Eighteenmile Creek and small roadcut on Lake Shore Road to the right. This is STOP 1 of tomorrows B3 Genesee Formation field trip (this volume).
22.8	0.15	Entrance to Frank Lloyd Wright Estate ("Graycliffs") on the right. This newly restored attraction in the Buffalo area was a major access route to the shore cliffs along the lake in the 1980s before the great ladder below the estate fell into disrepair.
24.55	1.75	Turn right into long private driveway immediately to the north of the Lake Shore Road bridge over Pike Creek; continue through the turnaround at the end of the driveway and park facing back toward the entrance. Do not block the driveway. Exit vehicles.

STOP 2. Lake Erie bluffs adjacent to the mouth of Pike Creek near Derby, Erie County (see Fig. 6). Proceed from end of long private driveway north of Pike Creek across yard to beach. Follow the beach southwest past Pike Creek to the beginning of a long bedrock cliff section bordering the beach. This section is in the southwestern part of a nearly continuous outcrop extending from Eighteenmile Creek almost to Sturgeon Point (STOP 4). Visible to the northeast are high cliffs exposing a long section, extending from the Wanakah Shale in the Hamilton Group, upward into the Frasnian Rhinestreet black shale. Units visible in the cliff south of Pike Creek, include: in ascending order, 2 meters of the Middle Devonian (Late Givetian) Windom Member, approximately 45 cm of lower Late Devonian (Frasnian) North Evans-Genundewa strata, 2.5 meters of gray and black shales of the West River Formation, and 2 meters of black Middlesex Formation. At this stop we will focus on the North Evans-Genundewa interval (Fig. 6).

This section shows the Windom Member-North Evans Limestone contact as a knife-sharp boundary separating chippy gray Windom shale rich in the small brachiopod *Emanuella praeumbona,* from dark, dysoxic beds of the Genesee Group (Fig. 6). The North Evans is anomalous in this area in that it is very thin and characterized by more detrital pyrite than typically found in sections closer to Buffalo. Furthermore, above the thin North Evans bed, and separating it from the overlying Genundewa Limestone, is a 25 cm-thick brownish black, flaggy shale unit that is unnamed (Fig. 6). This shale, containing flattened *Styliolina*, *Pterochaenia* , and wood debris, is very localized in distribution. At the North Evans type section on Eighteenmile Creek, adjacent to-, but immediately downstream from, the Amtrack railroad overpass, about four km east of STOP 2, the North Evans Limestone is a 6-12 cm thick lag accumulation dominantly composed of crinoidal debris, but also characterized by abundant conodonts, glauconitic grains, fish debris, and reworked concretions (Fig. 6). At the overpass bridge, the North Evans grades directly into styliolinid grainstone facies of the overlying Genundewa, and no dark shale is present. However, further downstream from that bridge to the northwest, a succession of creek bank sections shows dramatic thinning of the North Evans Limestone with the appearance of the feather edge of the dark shale as a nodular parting above the lag unit. Moreover, the North Evans undergoes a lateral, spectral change from a thick crinoidal unit to a thin layer of mixed carbonate grains and detrital pyrite that is distinctly more "Leicester"-like (Fig. 6). In the next key section (Fig. 6) along the Old Lakeshore Road (to be seen by participants of the Sunday fieldtrip B3: this volume), the intervening shale is 10-11 cm-thick and the North Evans Limestone is only about 2-3 cm-thick. This intervening shale unit is herein interpreted as being a local basinal facies of the Genundewa Limestone in that its upper contact is conformable (non erosional) with the Genundewa Limestone. The westward North Evans allochem transition from dominantly calcareous to largely insoluble grains, accords closely to the appearance of the overlying dark shale, and it suggests that carbonate dissolution of the exposed lag was more intense in the more basinal subenvironment of STOP 2 than at localities further to the northeast.

At STOP 2, the North Evans Limestone rests on the Taghanic Composite Unconformity, a disconformity of large time-magnitude. Based on comparison with more continuous Devonian sections further east, the section at Lake Erie is severely truncated; approximately one third of the Windom Member, the entire overlying Tully Formation, and all of the succeeding Geneseo and Penn Yan formation successions are absent at STOP 2 (Fig. 2). The North Evans Limestone is a complex lag blanket famous for its fish material (Hussakoff and Bryant, 1918, Bryant, 1921, Turner, 1998) and conodonts (Hinde, 1879, Bryant, 1921, Huddle, 1974, 1981). North Evans conodonts are exceedingly abundant and diverse in species and types of elements. Elements of the North Evans fauna, both its fish debris and conodonts, can be traced from Erie County to the Genesee Valley. Preservation of the conodont elements is distinctive and varies from complete and pristine (light amber-colored) to broken, dark (almost black), and degraded (Color Alteration Index is 2 to 3). Colors of some of the fish debris are red, orange, gray and blue. The taphonomic history of the North Evans debris is complex and tests the limits of chronostratigraphic resolution (Kirchgasser and Koslowski, 1996; Kirchgasser and Vargo, 1998; Kirchgasser, 1994, 1998, 2001, 2002, 2004). The final taphonomic (burial age) of the North Evans material correlates to the youngest zone conodont in the mix, which is *Ancyrodella recta* of the upper part of Lower Frasnian MN Zone 2 (Kirchgasser, 1994); early whorls of the Lower Frasnian goniatite *Koenenites* (also MN 2 age) have been recovered from North Evans conodont residues at the type section at Eighteenmile Creek (Fig. 6a) and in lenses within the Genundewa Limestone at Linden in Genesee County (see Sunday illustrations). The North Evans, as with most lag beds, poses a depositional paradox; even though the lag content records an enormous span of time, the actual final depositional event producing the bed may have been geologically instantaneous.

The Genundewa Limestone at STOP 2 is a 30-40 cm-thick ledge composed of styliolinid grainstone-packstone carbonate (Fig. 6). It is typically brownish gray, massive to nodular, limestone which sometimes weathers into thinner, flaggy beds. Aside from *Styliolina* and the small bivalve *Pterochaenia* , fossils are scarce, usually small, and of low diversity. The goniatites *Koenenites, Acanthoclymenia,* and *Tornoceras,* as well as *Manticoceras* at the very top, occur in the Genundewa, but are rare or poorly preserved here; the conodont *Ancyrodella recta* of upper MN Zone 2 occurs in the North Evans Limestone and at the base of the Genundewa Limestone at Linden in Genesee County (see Sunday illustrations). As noted, the Genundewa is a pelagic limestone that probably represents oxygen-stressed, sediment-starved, basin slope conditions. However, at this locality, local erosional channelization is evident locally within the limestone (Fig. 6). These channels, as well as pervasive small-scale hummocky cross-bedding and widespread alignment of *Styliolina* throughout the Genundewa, attest to significant current activity at the substrate, probably caused by deep-storm waves. The top of the Genundewa is gradational with the overlying shaley West River Formation; the topmost Genundewa becomes shaley and flaggy before giving way to the typical alternation of thin black and gray shale normally seen in the West River (Fig. 6). We herein interpret the interval from the sub-North Evans disconformity upward into the lower medial West River Formation is a Transgressive Systems Tract, starting with an erosional lowstand event recorded by the disconformity. The succeeding Genundewa is a condensed interval recording transgressive deepening within the basin and sediment-starvation on slope. A possible Maximum Flooding Surface is represented by a recurrent conodont-glauconite-bearing styliolinid bed within the basal West River Shale (Fig. 6); this conodont-rich bed has been traced westward from the Genesee Valley and is Huddle (1981) sample horizon 8122SD at Fall Brook, Geneseo (Kirchgasser et al., 1994, A-4, STOP 1, fig. 10). Early highstand facies are represented by succeeding alternations of black and gray shale. This overall transgression probably represents a hybrid eustatic and flexural event within the basin.

Board vehicles and drive to driveway entrance immediately to the north of Pike Creek. Turn right onto Lakeshore Road proceeding to the southwest.

24.6	0.05	Cross Pike Creek.
24.8	0.2	Sweetland Road-Lake Shore Road junction; bear right and continue straight on Lake Shore Road.
25.8	1.0	Turn right into private driveway (shore home). Park in available lots and exit vehicles.

STOP 3. Rhinestreet Formation at Lake Erie shore bluffs below at Seascape (private residence), 2 km northeast of Sturgeon Point near Derby, Erie County (see Fig. 7). Near-vertical shore cliffs between STOPS 2 and 3 prevent our seeing units (Middlesex, Cashaqua formations) between the West River shale and the Rhinestreet shale. We will, however, see the top of the Cashaqua at STOP 5. STOPS 3 – 5 are directed to features in the lower part of the Rhinestreet Formation. Proceed across private yard to steps leading down the shore cliff. Turn

right and proceed northward (down-section) along the beach to exposure of basal Rhinestreet strata.

The Rhinestreet Formation is one of the thickest Late Devonian black shale divisions in western New York. Hard, well-jointed, black shale makes up most of this section. Less organic-rich, recessive weathering dark gray shale intervals, as well as, beds of greenish gray shale, can also be seen (Fig. 5). Conspicuous at several levels within the black shale intervals are large, often massive, septarial concretions. After walking northward on the beach, we will stop at a level displaying the largest concretions which occur about two meters below a 2.2 meter-thick gray shale unit (Fig. 7). The black shale facies within the Rhinestreet records intervals of severe dysoxia to near-anoxia within the Devonian basin associated with a broad time interval associated with global sea level highstand (Johnson et al., 1985). This organic-rich lithofacies accumulated in a relatively deep-water, stratified basin setting west of the prograding Catskill Delta in a foreland basin already maintained by collisional thrust loading (Ettensohn, 1998; see Fig. 1). Contemporaneity of Catskill Delta progradation is splendidly shown by the eastward splaying of Rhinestreet black shale divisions into the deltaic clastic wedge and by eastward passage of organic-rich basinal facies into shoreward, coarse, fossiliferous neritic facies and terrestrial red beds (Woodrow and Sevon, 1985; see deWitt, 1960; deWitt and Colton, 1959; Colton and deWitt, 1958; Sutton, 1963, and Kirchgasser et al., (1994) for evolution of correlations and unit terminology pertaining to eastern Rhinestreet and other West Falls Group subdivisions). The 2.3 meter to 2.4 meter-thick gray shale unit above the band of large concretions is the western distal "toe" of a progradational clastic pulse extending westward from the delta complex; this fine grained turbiditic or hemipelagic sediment was probably deposited during a sea level lowstand event which allowed prodelta muds to be exported far into the basin (Fig. 1). A turbiditic origin for part of this interval is suggested by the presence of a 2 – 5 cm-thick siltstone bed at its basal contact at STOPS 3, 4, and 5 (Fig. 7); the basal surface (sole) of this layer displays erosional groove cast impressions and the top of the bed fines upward into featureless gray shale, suggestive of a turbiditic event. At STOP 3 (Seascape section), the basal lag of the siltstone bed yields conodonts of MN Zone 7 (Levin and Kirchgasser, 1994; Kirchgasser et al., 1994; Kirchgasser and Klapper, 1992). About one meter below the green-gray shale unit is a recessive notch that marks the position of a K-bentonite layer (Fig. 7). This altered ash bed is characterized by gray brown clays (kaolinite and mixed layered clays), bleached (pyroclastic?) micas, and secondary pyrite with minor quartz, calcite, plagioclase and ?apatite. The ash bed is graded with clay flakes in the upper part with even stronger parallel orientation than the clays in the overlying black shale. Lenses with the distinctive color and fabric of the ash occur intermittently for a few centimeters above the ash bed. As noted above, this ash bed is apparently not the Belpre Ash Bed of Tennessee (conodont MN Zone 8, Rotondo and Over, 2000) but an earlier event within the interval of MN Zone 7); the Belpre Ash has a radiometric date of 381.1 +/- 3.3 Ma (Tucker et al., 1998; Kaufmann, 2006).

Return to vehicles and exit to driveway junction with Lakeshore Road; turn right onto Lakeshore Road and continue to the southwest.

26.9 1.1 Junction of Lake Shore Road with Sturgeon Point Road; turn right (northwest) onto Sturgeon Point Road.

27.6 0.7 Enter Sturgeon Point Park and boat marina complex. Enter to the left and follow the road around to the northeast end of the complex.

27.95 0.35 Arrive at northeastern limit of parking area. Park and exit vehicles.

STOP 4. Rhinestreet Formation in Lake Erie shore bluff succession, both to the northeast of-, and southwest of, the Sturgeon Point marina and car park complex near Derby, Erie County (see Fig. 7). We will, first, proceed on foot from the northeast end of the car park to the beach and follow the beach and continuous shale cliff section to a position near the outflow pipes from the Derby waterworks. The bluff section is nearly the same as that for STOP 3, except that most of the black shale interval below the gray shale unit is below lake level. The K-bentonite is not accessible here, but the gray shale unit is much easier to examine at this outcrop. Moreover, the black shale interval above the gray unit can be examined here (Fig. 7).

Within the gray shale unit in this locality are very thin, 0.2 – 6 cm-thick, black shale beds that often display sharp contacts and strong visual definition within the thicker gray succession. Buff gray concretions occur at two levels within the gray shale interval; these are concentrated closely below black shale bands and appear to be controlled by the presence of the bands. However, in a few places along the exposure, the thin, overlying black shale bands pinch out over the tops of the subjacent nodules. This suggests that concretion growth may have created differential paleorelief , perhaps due to early dewatering and differential settling of mud. The sharply defined thin, black bands pose an interesting question: Do they represent slow background deposition between turbiditic gray mud pulses, or do the thin black bands, themselves, represent some alternative type of rapid depositional event involving sedimentation, or resedimentation, of organic-rich sediment? Is it possible that the organic-rich sediment, instead, may have been originally pelletal, hence mobile and easily transported on the seafloor? This latter scenario, yet untested, could explain these sharp pinstripe bedforms as rapidly deposited, current-traction-generated features.

To the immediate northeast of the second waterworks outlet is a problematic structural displacement or offset at the level of the gray shale unit. At the lake edge, several fractures in the lower black shale division can be seen that are filled with gray shale that displays soft-sediment shearing and fracturing. Directly across from the area of shore fractures in the cliff face the gray shale unit thins to less than a third of its normal thickness across a distance of about 80 meters. However, debris on the beach conceals the intervening area and the full nature of this structure. The shore cliff and lake edge exposure displays excellent examples of joint networks, particularly for the black shale bands. These joints are believed by Gary Lash to have evolved during thermal-burial maturation of the black shale unit as the black shale units began to function as hydraulic top seals for moving fluids migrating up from below (see Lash, 2006; Lash and Blood, NYSGA Sunday trip B1, this volume). The different orientations of the various joint sets are believed to correlate to a series of far-field stress phases associated with the Allegheny orogeny (Lash and Blood, Sunday trip B1, this volume). This outcrop will be important as a stop on that Sunday trip.

Between 40 and 45 cm above the gray shale interval, within black shale facies, is a styliolinid-rich, hashy layer that is associated with widely-spaced, spheroidal concretions (Fig. 7). Close examination of the layer shows the presence of numerous, flattened goniatites (some partially pyritized), spotty concentrations of *Styliolina*, fish debris, large horizontal (arthropod?) trace fossils and scattered conodonts. The goniatites first identified during a survey of this section this past June, belong to the genus *Naplesites* in the family Beloceratidae. The lineage is characterized by compressed (discoidal), evolute shells with increasingly numerous, distinctly pointed lobes (and saddles) forming chevron-like patterns. The ancestor of the family is *Probeloceras* which in New York occurs below the Rhinestreet in the Cashaqua Shale (Sonyea Group; MN Zone 5); the group culminates with the extremely multilobed *Beloceras*, a genus still unknown in North America (Fig. 3). The discovery of the *Naplesites* horizon at Sturgeon Point (Stop 4) is important in that its presumed position in the lower Rhinestreet (House and Kirchgasser, 1993) is confirmed and the conodonts in the bed may prove to be datable. *Naplesites* is otherwise rare in New York and is known only from a few specimens (two species) described by Clarke (1898) far to the east from unspecified horizons and sections in the shales around Naples in Ontario County (Canandaigua Lake meridian). *Probeloceras* and *Naplesites* (as *Mesobeloceras*) are illustrated in Kirchgasser, Over and Woodrow, 1994, fig. 7).The close stratigraphic coincidence of useful conodonts, *Naplesites*, and the K-bentonite bed is important geochronologically and is the subject of ongoing work.

Return to vehicles and proceed by car around to the southwestern most parking area near the Sturgeon Point pier.

28.35 0.4 Park and exit vehicles.

Proceed on foot to beach below car park and follow beach southwestward for approximately 120 meters to a low outcrop bench of black Rhinestreet shale at the lake edge.

This exposure of black shale is probably stratigraphically higher than the accessible beds northeast of the marina owing to regional dip effects, but the precise match of beds can not yet be made owing to the long covered interval between the two sections. Notable in this outcrop are several bands of concretions including a line of massive septarial concretions below the bench of very resistant black shale. At the top of the black shale bench are two closely-spaced horizons of small spherical to ellipsoidal, 0.3 – 4 cm-diameter, phosphatic nodules. Phosphatic nodules, well known from the upper part of the New Albany Shale and the Cleveland Shale, have not been reported from the Rhinestreet. The occurrence of nodules of this type has been interpreted as evidence of nutrient upwelling and high productivity in surface waters (see Robl and Barron, 1989). What is striking here is that many, if not most, of the nodules are nucleated by fish spines and scales. Small, sand-size grains within some of the nodules may be radiolarian tests.

Return to vehicles and follow one-way road around to exit of Sturgeon Point Marina.

28.7 0.35 Exit Sturgeon Point Marina and proceed southeast on Sturgeon Point Road.

29.4 0.7 Junction of Sturgeon Point Road with Lake Shore Road; continue straight on Sturgeon Point Road.

30.4 1.0 Junction of Sturgeon Point Road with NY 5; continue straight on NY 5.

33.55 1.25 Junction of Sturgeon Point road with US 20. Turn left and proceed to the northeast on US 20.

35.3 1.75 Junction of US 20 with Shadagee Road immediately southwest of US 20 bridge over Eighteenmile Creek. Turn right onto Shadagee Road.

35.35 0.05 Park vehicles by North Evans cemetery on the right. Depart vehicles.

STOP 5. Cashaqua Shale/Rhinestreet Shale contact and overlying Rhinestreet strata exposed on Eighteenmile Creek and on access road leading to that creek, town of North Evans, Erie County (see Fig. 7). Enter access path opposite from North Evans Cemetery. Proceed on foot down sloped path to valley bottom. Cross flat area to edge of Eighteenmile Creek.

The Cashaqua Shale Formation of the Sonyea Group in southern Erie County is characteristically composed of gray fissile shale with a rhythmic succession of discoidal concretion bands (Buehler and Tesmer, 1963; Kirchgasser et al., 1997, fig. 9). The fauna of this unit consists of small mollusks and brachiopods, indicative of a dysoxic setting. Most notable, is the occurrence of numerous, often poorly preserved, goniatites belonging to the genera *Manticoceras* and *Probeloceras* which are flattened outside of the concretions and variably three dimensional inside of them (conodont Zone MN 5). *Probeloceras* is the ancestral genus of the family Beloceratidae and is followed by the genus *Naplesites* in the Rhinestreet Shale seen at Sturgeon Point, Lake Erie (Stop 3; see Kirchgasser, Over and Woodrow, 1994, fig. 7 for illustrations of Sonyea and West Falls Group goniatites). The prominent concretion 3ayer with MN Zone 6 conodonts in the dark shales of the upper Cashaqua Shale is a septarian band with white or pink barite filling the shrinkage cracks and in places east of the Genesee Valley in Livingston County, replacing the shells of a rich molluscan fauna including the goniatites *Manticoceras*, *Prochorites*, *Acanthoclymenia* and *Aulatornoceras*; conodonts in the bed indicate MN Zone 6. The *Prochorites* is a species (*P. alveolatus*) known elsewhere only in Western Australia.

We will see the contact of this unit with the overlying Rhinestreet Formation of the West Falls Group in the bed of Eighteenmile Creek. The base of the Rhinestreet is sharp and marked by a seam of diagenetic pyrite which weathers to a rusty band in bank sections. No detrital pyrite has been found on the contact, but a rich association of conodonts is reported from the basal few centimeters of black Rhinestreet shale (Huddle, 1968; Kirchgasser and Klapper, 1992; Klapper et al., 1995).

Ascending the access path as we return to the vehicles, we pass the same lower Rhinestreet gray shale division that we examined at STOPS 3 and 4. Along the path this section is weathered and slumped, but, on the other side of Eighteenmile Creek, an excellent profile through this interval can be seen on a side tributary waterfall (Fig. 7).

Board vehicles and turn cars around. Return to Shadagee Road/Route 20 intersection.

35.4 0.05 Shadagee Road/ US 20 intersection. Turn left (to southwest) on US 20.

37.15 1.75 Junction of US 20 with Sturgeon Point Road; continue to southwest on US 20.

39.25 2.1 Junction of US 20 with Eden-Evans Center Road; turn left (east) on Eden-Evans Center Road.

39.95 0.7 Eden-Evans Center Road crosses over New York State Thruway (I-90).

40.6 0.65 Entrance to the Thruway on the left; continue straight (east) on the Eden-Evans Center Road.

43.9 3.3 Intersection of Eden-Evans Center Road with NY 62 in Eden; Continue straight (east); the road name changes to Church Street.

45.2 1.3 Intersection of Church Street with Jennings Road; continue straight (east) on Church Street. Church Street soon bends to southeast as the Road descends into the valley of the south branch of Eighteenmile Creek.

46.35 1.15 Church Street bridge over the south branch of Eighteenmile Creek.

46.5 0.15 Turn right onto Old Mill Road (a dead-end lane) and continue to its end.

46.65 0.15 Exit vehicles at end of Old Mill Road.

STOP 6. Section of Pipe Creek Formation ("lower Kellwasser Bed") and synjacent units at waterfall along south branch of Eighteenmile Creek, southeast of Eden, Erie County (see Figs. 8, 9). Proceed on foot from the vehicles to the lip of the large waterfall. The gray, silty, bioturbated unit in the falls face is the top part of the shaley Angola Formation of the West Falls Group. This is abruptly capped by the resistant, black, and well jointed Pipe Creek Formation of the Java group (Fig. 8) that forms the erosional lip of the waterfall. The 1.6 meter-thick Pipe Creek succession is followed upstream in the floor of the creek by softer, nodule-rich, gray and black shale facies of the Hanover Formation (Fig. 8).

The black Pipe Creek shale consists of massive, organic-rich facies that contrasts markedly with both the underlying Angola and overlying Hanover. The base of the Pipe Creek is abrupt on the Angola with the development of abundant diagenetic pyrite in the uppermost Angola. No erosional, detrital pyrite has been found at the contact, but a profound change in microtexture is seen as one passes from the bioturbated, pyrite- suffused, uppermost Angola, into the laminated, black Pipe Creek (Fig. 9). As noted in the text, this unit has now been found to correlate to the "Upper Kellwasser Bed" which marks a major ecological reorganization and extinction event in global sections.

In the process of measuring this section for this field trip, we found an apparent K-bentonite bed 15 cm below the top of the Pipe Creek (Fig. 8, unit e). This layer consists of approximately 1 cm of soft clay rich in pyroclastic mica. The bed is currently being analyzed for its geochronostratigraphic potential (see text).

Above the Pipe Creek shale, there is an abrupt change to softer, gray shale that abounds in small concretionary nodules through an interval of about 1.5 meters (Fig. 8). One bed (Fig. 8, unit g), is packed with very small irregular nodules such that this layer somewhat resembles a muddy version of the classic condensed, gray, nodular limestone facies of European sections (Tucker, 1974; Tucker and Kendall, 1973). Fossils are common in the lower Hanover, but are usually very small. Diminutive gastropods, bivalves, and ostracodes make up most of the faunal trail mix. Goniatites, important as chronostratigraphic markers, occur sparingly in this interval at this section. However, at Java Village in Wyoming County, *Delphiceras* occurs just above the top of the Pipe Creek and somewhat higher in the section at Walnut Creek. At Silver Creek in Chautauqua County *Sphaeomanticoceras* and *Crickites* can be found, the latter genus being the highest zonal indicator of the Frasnian Stage (Fig. 3; House and Kirchgasser, 1993).

Board vehicles and return to junction of Old Mill Road with Church Street.

46.8	0.15	Turn right onto Church Street immediately west of Church Street-NY 75 intersection.
46.85	0.05	Junction of Church Street with NY 75 (Sisson Highway); turn right on NY 75 and continue to the southeast.
47.8	0.95	NY 75 crosses over south branch of Eighteenmile Creek. Gray mudstone deposits of the Hanover Formation are visible in upstream cutbank exposure to the left.
48.15	0.35	Junction of NY 75 with New Oregon Road which splits off to the left; turn left onto New Oregon Road.
48.4	0.25	New Oregon Road bridge over south branch of Eighteenmile Creek; park cars on shoulder near bridge and exit vehicles.

STOP 7. Upper part of Hanover Formation showing "Upper Kellwasser Bed" (Point Gratiot Bed) and basal contact of Dunkirk Formation on south branch of Eighteenmile Creek below New Oregon Road bridge, southeast of Eden, Erie County (see Figs. 10, 11). Exit vehicles and proceed to the downstream end of a long, west-facing cutbank along the south branch of Eighteenmile Creek below the New Oregon Road overpass. At this point gray shale of the upper middle part of the Hanover Formation are abruptly overlain by the black Point Gratiot Bed and succeeding gray and black shale beds of the upper part of the Hanover Formation (Fig. 10).

At this locality the Point Gratiot Bed is a black ledge that correlates to the "Upper Kellwasser Bed" of Late Devonian sections globally (see text). The topmost part of this ledge marks the actual Frasnian-Famennian extinction event of literature based on conodont work in western New York sections (see Over, 1997, 2002; Day and Over, 2002). Although the bed is composed of basinal black shale, fossils such as the probable anaptychus organ *Spathiocaris* , fish material, and wood debris can be found, particularly near the top of the bed. Moreover, elevated levels of Platinum Group elements occur at this level and other horizons in the boundary interval black shales (see Over et al., 1997).

Above the Point Gratiot Bed in this section is a 2.2 meter-thick upper Hanover succession of alternating gray and black shale beds that is capped by the black, well jointed Dunkirk Shale at the top of the bank (Fig. 10). The upper Hanover interval closely resembles the lower Rhinestreet gray shale unit owing to several discrete, thin, black shale bands which contrast sharply with the thicker gray lithology. Reworked pyritic burrow clasts and exhumed geopetally pyritized *Tasmanites* half-spheres (*sensu* Schieber and Baird, 2001) have been found in thin lags at the base of the thin black beds and at the base of the Dunkirk Shale. Immediately above the Point Gratiot Bed in this section is a thin, gray, biotubated shale bed yielding small nodules (Fig. 10). Above this level, gray shale units become less visibly bioturbated and more non calcareous, suggestive of a long-term transgressive trend in the interval. Hanover deposits below the Point Gratiot Bed are markedly more calcareous, lighter colored, and more intensely bioturbated; upward changes across the Point Gratiot Bed probably reflect combined effects of transgressive deepening in the basin and adverse biological effects associated with the Frasnian-Famennian crisis. Participants should look for any sign of the fossiliferous "recovery bed" discovered by Jeff Over approximately one meter above the Point Gratiot Bed near Java Village (Day and Over, 2002). Although it is absent from many area sections, the STOP 7 outcrop has not yet been seriously checked for the presence or absence of this unit.

Board vehicles and retrace route to entrance of New York State Thruway.

55.9 7.5 New York State Thruway (I-90) entrance from Eden-Evans Center Road. Enter onto Thruway and return to Adam's Mark Conference Center.

End of field trip.

Life on the Edge: Death and Transfiguration in Mud

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SUMMARY

The hypothesis of Coordinated Stasis posits that paleocommunities were relatively stable associations of species that tracked together through cyclical changes in their regional environment. When this tracking capacity was exceeded, they experienced relatively abrupt turnover in species composition, leading to a new interval of relative stability. This fieldtrip will focus on changes in community structure within the dysoxic, *Ambocoelia*-chonetid biofacies through the Wanakah and Lower Windom depositional cycles. We will also discuss the implications of the observed patterns for habitat tracking and community stability over longer time scales.

INTRODUCTION

Recent work on the causes underlying patterns of biofacies distribution in time and space have lead to a set of hypotheses referred to as Coordinated Stasis (Brett et al. 1996). Basic postulates of this model are that relative stability in the species composition of biofacies assemblages is a consequence of the ability of species communities to track preferred habitats and thereby avoid selection pressures that would otherwise disrupt community stability. When, on rare occasions, extraordinarily pervasive and abrupt environmental changes overwhelm the ability of species assemblages to track their preferred habitats, communities undergo a cascade of species turnover during which a substantial fraction of the formerly characteristic species go extinct and new species join the survivors to form a new and markedly altered community (Brett and Baird, 1995 and Brett *et. al,* 1996). Habitat tracking is most easily generalized as an effect of sea level change. As sea level fluctuates through transgressive - regressive cycles, it follows that environments will synchronously fluctuate along an onshore - offshore gradient. If communities of organisms that share a preferred environment are to remain in this environment, then they will be forced to migrate up and down depositional slope through time with sea level. A critical issue for the testing of Coordinated Stasis, therefore, is the degree to which we can observe habitat tracking through the fossil record that is preserved within stratigraphic sequences.

The geographic setting of this trip is Western New York, with four outcrops from the Wanakah and Lower Windom Shale Members of the Hamilton Group, northern Appalachian Basin (Figures 1 and 2). The Middle Devonian Hamilton Group (Givetian Stage) is a dominantly fine-grained marine deposit. The sediments originated from the erosion of a mountain belt to the east that formed during the Acadian Orogeny. Most of the Hamilton Group records the interval between the second and third phases of mountain building that occurred from the collision of microcontinents into New England and the central Atlantic United States (Ettensohn, 1985; Ver Straeten and Brett, 1995); more specifically, the oblique convergence between the Laurentian and Avalon terranes (Ettensohn *et al.*, 1988). This well-documented stratigraphic succession provides an ideal setting in which to examine the predictions of Coordinated Stasis. Indeed, it is the setting in which Brett and Baird (1995) developed the hypothesis.

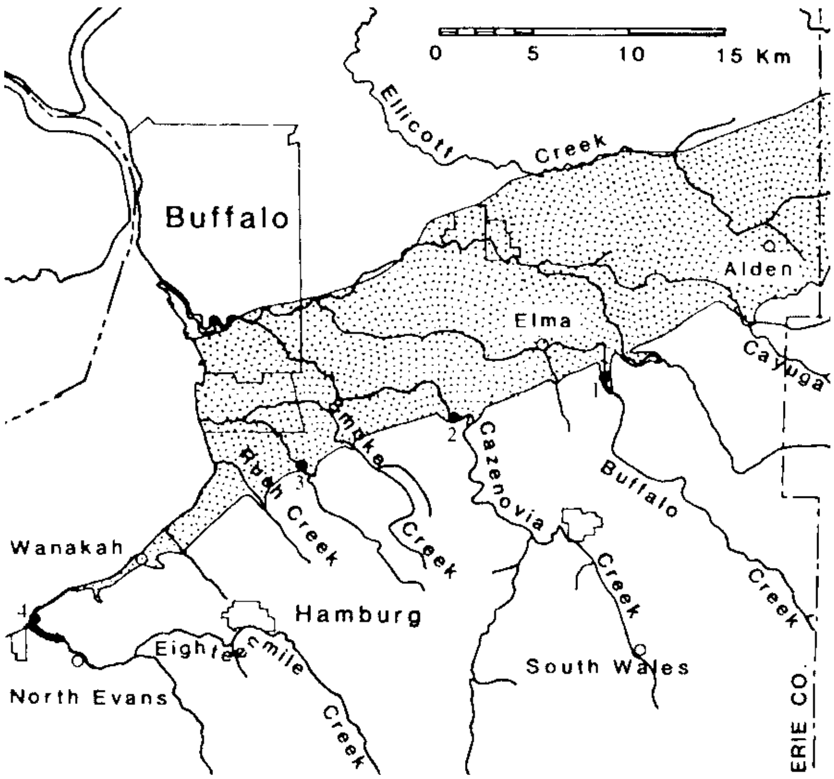

Figure 1: Map of a portion of Erie County, showing geographic location of the localities visited on this fieldtrip. Stippled area is the outcrop belt of the Hamilton Group, Middle Devonian. Locality 1: Buffalo Creek at Bullis Road; 2: Cazenovia Creek at US 20; 3: Smoke Creek (South Branch) at Milestrip Road; 4: Lake Erie Shore at Eighteenmile Creek. (Modified from Goldman and Mitchell, 1990).

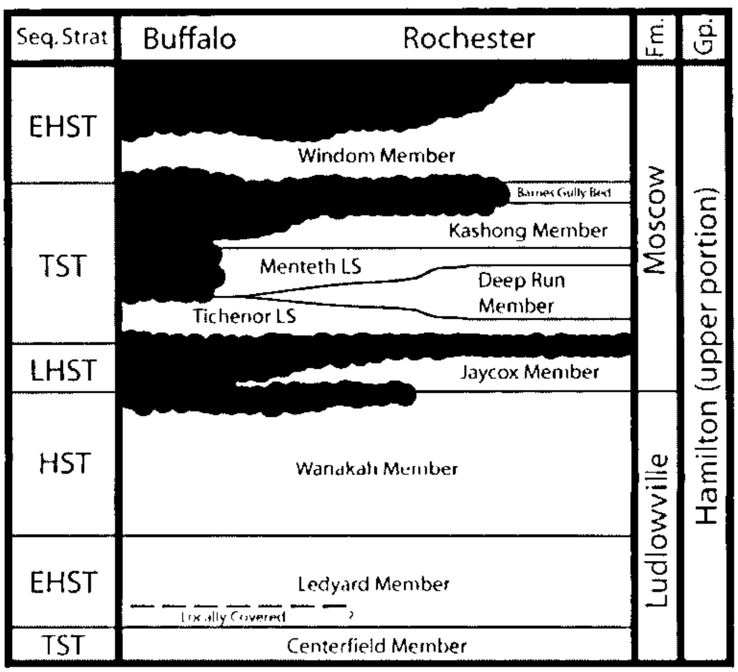

Figure 2: General stratigraphy of the upper portion of the Middle Devonian Hamilton Group in western New York. Column on left represents position within third-order sea level cycles. TST: transgressive systems tract, EHST: early highstand systems tract, HST: highstand systems tract, and LHST: late highstand systems tract. (Modified from Brett and Baird, 1996).

Sea level change can occur at different magnitudes. Fifth- or sixth-order cycles are typically seen as 0.5-meter thick intervals in the study area. These consist of subsymmetrical alternations of gray mudstone with concretionary, calcareous mudstones; an example of the latter is seen as the various trilobite beds of the Windom and Wanakah Shale Members. If these cycles were the product of Milankovitch forcing, they may have had a duration of ~20 to 100 ka each (Brett and Baird, 1996). Fourth-order cycles commonly display a biofacies spectrum (described in more detail below) from a low diversity "*Leiorhynchus*" fauna, through the *Ambocoelia*-chonetid biofacies, to either a diverse brachiopod or coral-rich community (Brett and Baird, 1996) and, correspondingly, may have had an approximate duration of 0.1 to 1.0 m.y. Third-order cycles, representing 0.8 to 2.0 m.y., correspond fairly well to previously described formations in New York State, primarily because Cooper (1930, 1933) used the thin and widespread limestones in New York as the bases of formations. These limestones have subsequently been interpreted as third-order transgressive systems tracts (TST) by Brett and Baird (1996).

Both the Ludlowville and Moscow Formations, therefore, represent distinct third-order depositional sequences. On this trip we will examine the relationship between observed faunal changes and the early highstand systems tract (EHST) and highstand systems tract (HST) portions of these third-order cycles. Previous work suggests that both of these systems tracts are aggradational, whereas the late highstand systems tract (LHST, which we will not examine here) is dominantly progradational (Brett and Baird, 1996). These comparisons provide an opportunity to gage the degree of change in biofacies composition that accompanies sea level changes of various magnitudes. Unlike the case for the third-order cycles, fourth-order systems tracts have not been delineated for all of the intervals studied. Thus, fourth-order cycles are described here simply in terms of relative sea level change; that is, as fourth-order transgressive and regressive facies, rather than in terms of systems tracts. See Brett and Baird (1996) and Batt (1996) for full description of sequences and systems tracts addressed on this fieldtrip.

The *Ambocoelia*-chonetid biofacies is one of numerous recurrent fossil assemblages, or biofacies, described within the Hamilton Group of New York. These biofacies are representative of relatively narrow ranges of benthic environmental conditions (Figure 3). The *Ambocoelia*-chonetid biofacies represents conditions of low oxygen in a relatively quiet offshore setting below maximum storm wave base. Apparent habitat tracking through fourth-order sea level cycles results in symmetrical community replacement from "*Leiorhynchus*" to *Ambocoelia*-chonetid, to *Athyris* biofacies and vice versa (Figure 3). *Ambocoelia umbonata, Crurispina nana, Devonochonetes scitulus, Sinochonetes lepidus, Longispina mucronata, Mucrospirifer mucronatas, Styliolina fissurella, Phacops rana, Greenops boothi* and Auloporid corals are all commonly present in this biofacies along with numerous other, less common taxa (Figure 4). For a more detailed list of taxa found within this biofacies, see Zambito (2006).

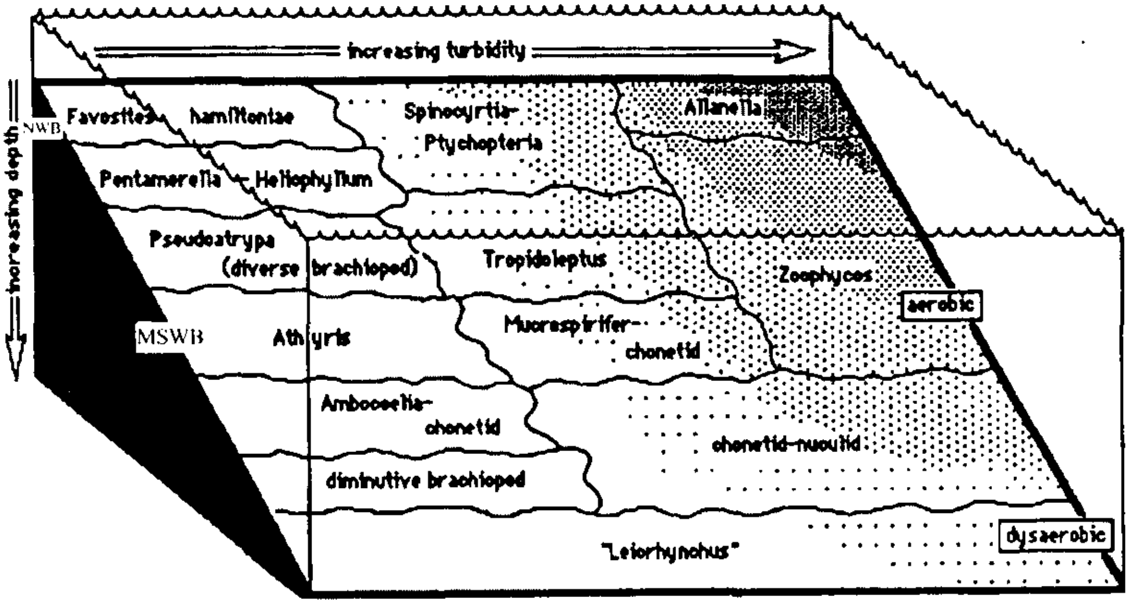

Figure 3: Biofacies model for the Middle Devonian Hamilton Group. Relations to water depth, (and corresponding turbulence and oxygen content) and turbidity (rate of sedimentation) are illustrated. NWB: normal, or fair-weather, wavebase; MSWB: maximum storm wave base. (Taken from Brett *et al.*, 1990).

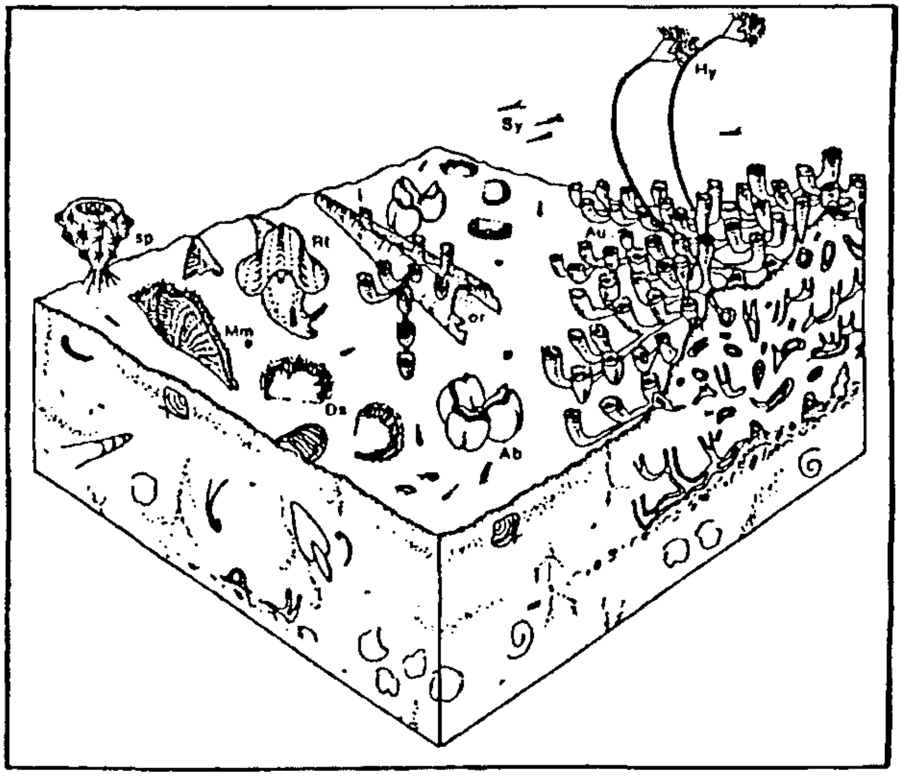

Figure 4: *Ambocoelia*-Chonetid biofacies reconstruction. Ab: *Ambocoelia umbonata*, Ds: *Devonochonetes scitulus*, Mm: *Mucrospirifer mucronatus*, Au: *Aulocystus jacksoni*, Hy: *Hyperoblastus*, Rt: *Retispira leda*; Sy: *Styliolina fissurella*, or: orthoconic nautiloid, sp: sponge. (Taken from Brett *et al.*, 1990).

Complicating any test of ecological stasis is the need to sample similar environments over time. This trip outlines the effects of sea level change, at the scale of both third- and fourth-order cycles, on organisms found within the dysoxic *Ambocoelia*-chonetid biofacies. Differences in community membership are described, and possible explanations for these differences are discussed. The primary application of this trip is a better understanding of considerations to be taken into account for studies that test communities for prolonged stability over time utilizing samples of from relatively small stratigraphic intervals.

STRATIGRAPHIC UNITS EXAMINED

Lower Wanakah Shale Member, Ludlowville Formation—The section of the lower Wanakah Shale Member examined in this study extends from approximately the top of the Fargo Bed (Kloc, 1983) to 1 meter above the Bidwell Bed (Kloc, 1983; Figure 5). Contained within this interval are up to 4 distinct limey bands (Murder Creek and Bidwell beds) that are rich in trilobites, including *Phacops rana* and *Greenops boothi*. The calcareous nature of these beds may reflect carbonate diagenesis below the sediment-water interface during the deposition of the overlying transgressive, sediment-starved, shales (Batt, 1996). This interval contains a number of species of ambocoeliids and chonetids and will be observed at Stop 4, along the Lake Erie shoreline at the mouth of Eighteenmile Creek (Figures 1 and 5). These sediments were deposited during a fourth-order regression within the third-order, Ludlowville HST (see Fig. 2, 6).

Upper Wanakah Shale Member, Ludlowville Formation—The most pronounced feature of the upper Wanakah section, seen during this trip at the stops at Cazenovia Creek and Lake Erie Shore, are the numerous concretionary horizons that can be grouped into two major intervals separated by approximately 2 to 3 meters of shale that contains sporadic concretions. In descending order, these are the Spring Brook Concretion Bed and the Walden Cliffs Concretion Bed of Kloc (1983). The Spring Brook and Walden Cliffs concretionary intervals comprise a series of concretions, which sometimes overlie one another, that reflect diagenetic enhancement of marker beds similar to the trilobite beds described above. The Spring Brook and Walden Cliffs concretionary intervals are similar both sedimentologically and faunally: both contain abundant pyrite, sometimes in the form of pyritized fossils and as pyritized shell hash horizons (Kloc, 1983); and both contain a number of species of ambocoeliids and chonetids. This interval of the upper Wanakah can be observed at stop 2, Cazenovia Creek at US 20 (Figures 1 and 5). The upper Wanakah Shale Member was deposited during a fourth-order transgression within the third-order, Ludlowville HST (see Fig. 2, 6).

Lower Windom Shale Member, Moscow Formation—The portion of the lower Windom Shale Member examined in this trip includes the lowermost *Ambocoelia*-rich shales. This is a medium gray, sometimes concretionary interval representative of an early highstand phase within the Windom Member depositional cycle (Brett and Baird, 1994). Horizontal, and sometimes vertical, pyritized burrows are common, indicating deposition under dysoxic conditions. This shale interval contains up to four species of chonetid brachiopods, only one species of ambocoeliid. We will examine this interval at stops 1 (Buffalo Creek at Bullis Road)

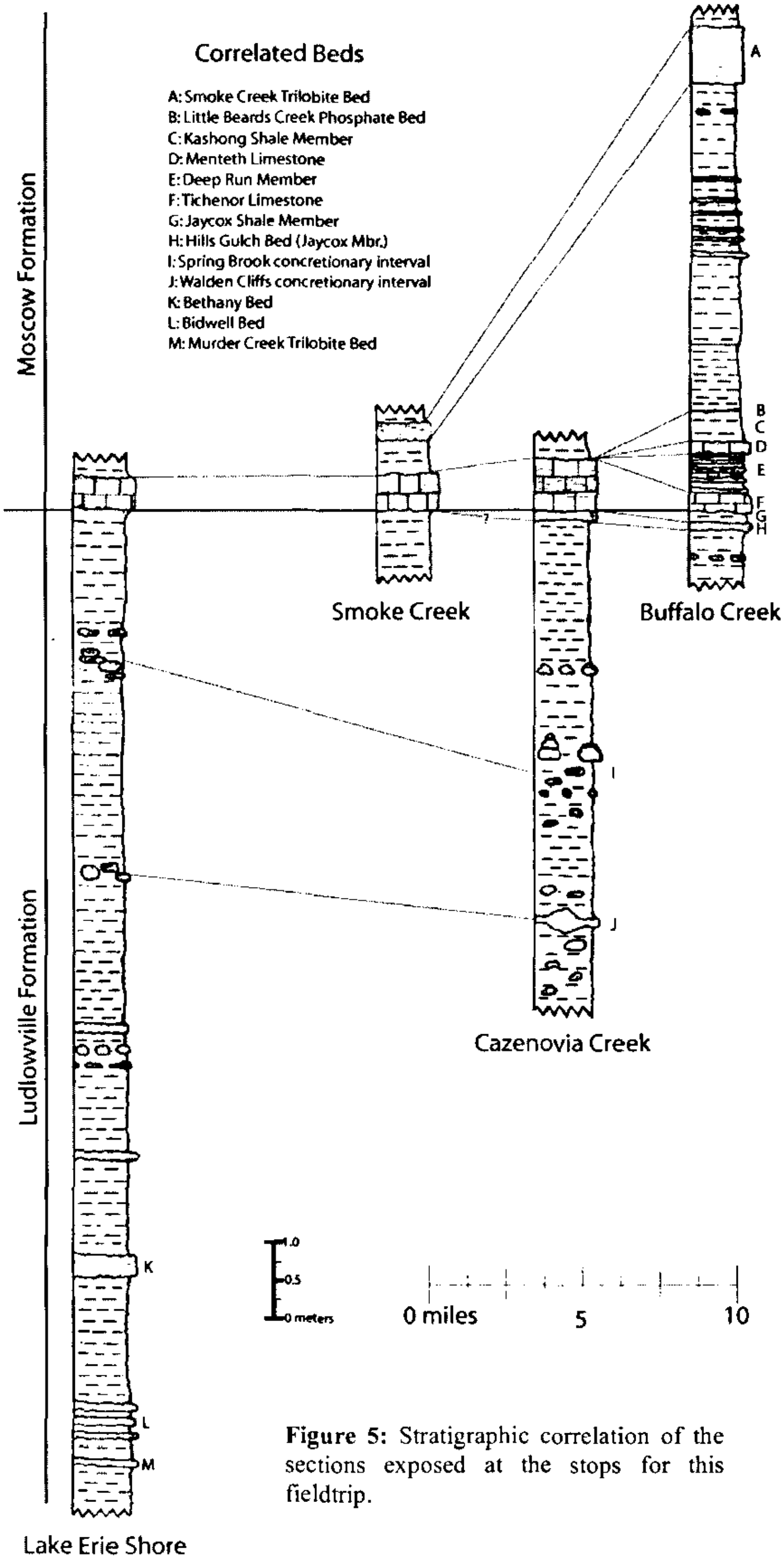

Figure 5: Stratigraphic correlation of the sections exposed at the stops for this fieldtrip.

and 3 (S. Branch of Smoke Creek at Windom, NY) (Figures 1 and 5). The lower Windom Shale Member sediments seen on this trip exhibit a fourth-order regression within the third-order, Moscow Formation EHST (Fig. 2, 6).

SAMPLING BIOFACIES

If we are to make a convincing test of community stability and habitat tracking, it is necessary to sample similar environments over time and to identify this habitat independently of the taxonomic composition of assemblages whose stability we seek to gage — that is, to control as nearly as possible for other sources of variation in community composition. Some environmental parameters of concern include sedimentation rate, redox conditions, and nutrient abundances (through measures of paleoproductivity), which are ultimately related, either directly or indirectly, to sea level fluctuations (Sageman *et. al*, 2003 and Tribovillard *et. al*, 2006). Thus, when testing for Coordinated Stasis it is essential to assess the relative position within sea level cycles at which the studied samples were deposited.

Data on the faunal composition and species abundance were collected through dysoxic intervals in the lower and upper Wanakah Shale Member as well as the lower Windom

Shale Member (Figure 2) to assess the effect that transgressive vs. regressive depositional settings may have on faunal composition. Accordingly, comparison of faunal constituents between these two depositional sequences also allows for insight into similarities and differences between two distinct third-order sea level cycles (Figure 2) within one Ecological-Evolutionary Sub-Unit (EESU; sensu Brett and Baird, 1995), or stable evolutionary block as described by the Coordinated Stasis hypothesis; and from this to gage the degree to which this community is stable in its composition.

The intervals of the Wanakah and Windom Shale Members that will be discussed were bulk-sampled in contiguous, 10-centimeter thick increments. Species abundance was determined by counting the number of specimens of each of the taxa present. Absolute abundances are favored over relative abundances so that a change in one taxon does not affect the abundance of all other taxa (Finnegan and Droser, 2006). Confidence in these results, in particular the concern that rare taxa might be missed and counted as absent is sustained by agreement of presence/absence of taxa sampled in this instance with similar data from Zambito (2006), in which a rigorous sampling effort involving replicate samples and sub-samples was performed on the same intervals within both the lower Windom and upper Wanakah Shale Members. Those samples proved to be statistically indistinguishable estimates of a common underlying species abundance distribution, and that any species that was present at the site in at least a one percent relative abundance was recovered with a 95% confidence. Absolute abundances of the most common taxa, along with and typically including all ambocoeliids and chonetids, are presented in Figure 6 for discussion.

BIOFACIES COMPARISONS AMONG SEA LEVEL CYCLES

The Hamilton biofacies exhibit gradational boundaries between adjacent biofacies. Thus the presence of *Leiorhynchus*, in the upper Wanakah interval indicates that these sedmiments were deposited near the deep water limit of the *Ambocoelia*-chonetid biofacies (see Figure 3), whereas its absence in both the lower Wanakah and the lower Windom intervals suggest that these reflect somewhat shallower environments (Batt, 1996). Thus, the lowest Moscow Formation Kashong, Deep Run, and lower Windom member deposits deepen upward (Brett and Baird, 1994), but the lack of *Leiorhynchus* in the capping lower Windom Shale Member suggests that this deepening did not proceed to the extent needed to introduce this species' preferred habitat. These observations illustrate how the environmental extent within, and the gradational boundaries between, biofacies provide sources of variance in sampled assemblages. This is another aspect to consider when determining intervals for sampling in studies of community stability.

Notable differences in community membership are evident between the Ambocoelia-chonetid communities of the upper Wanakah fourth-order regressive interval and the lower Wanakah transgressive interval, both of which fall within a single third-order cycle. *Crurispina nana* occurs only sporadically in the lower Wanakah Shale, whereas it is present at much higher abundance in the upper Wanakah Shale (Figure 6), indeed, this species makes up nearly 50% of the specimens obtained from some levels in the Wanakah study interval. Conversely, the trilobites *Phacops rana and Greenops boothi* become less common in the upper interval. These dissimilarities in faunal composition imply that environmental differences associated with these transgressive vs. regressive facies affected faunal dominance and composition of this biofacies.

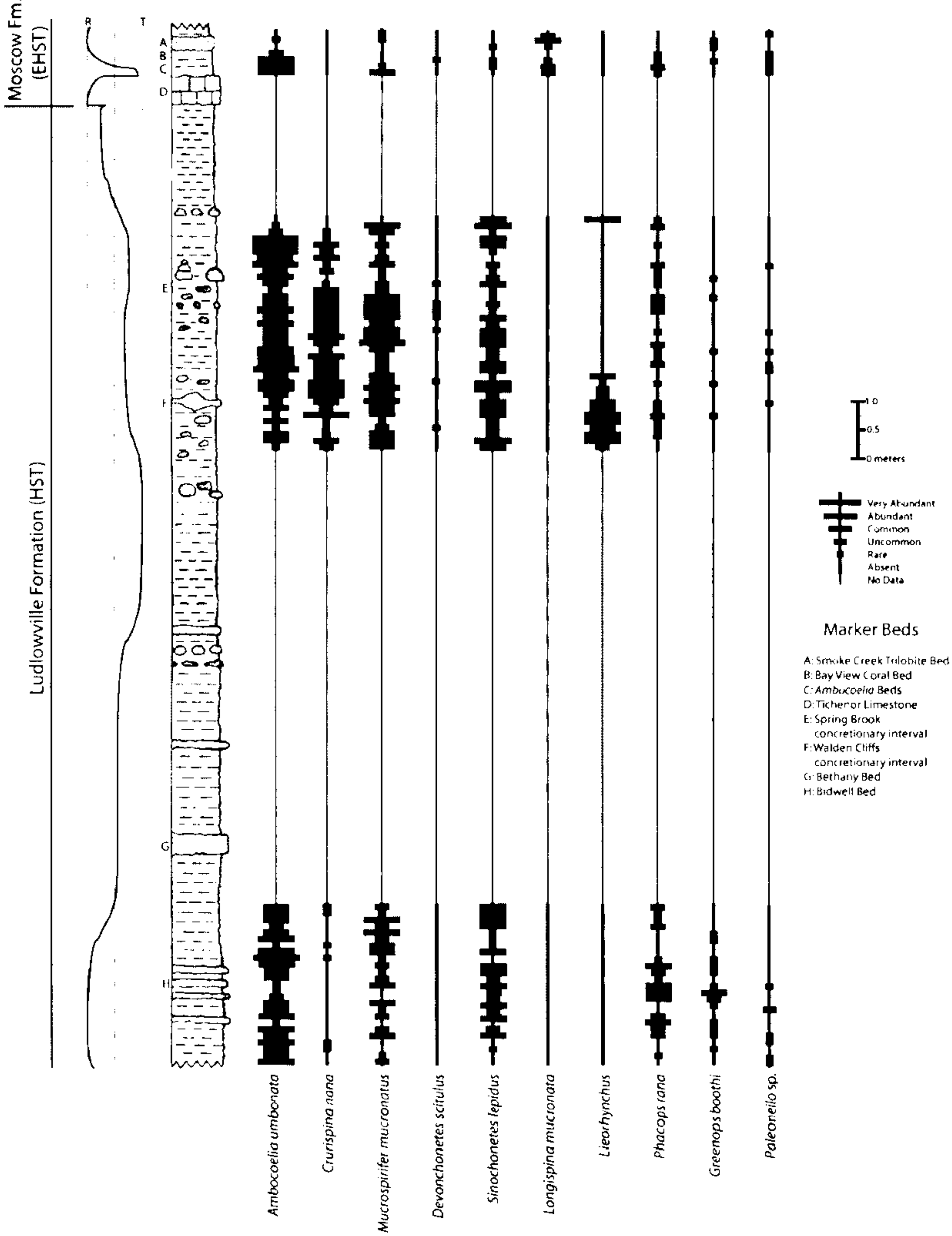

Figure 6: Composite Stratigraphic section for the stops of this fieldtrip, showing corresponding abundance of the main faunal components of the *Ambocoelia*-chonetid biofacies in the form of a spindle diagram. Lower Wanakah Shale Member faunal data collected at Stop 4; Upper Wanakah Shale Member faunal data collected at Stop 2; Lower Windom Shale Member faunal data collected at Stop 3. Column on left represents inferred sea level fluctuation of fourth-order cycles (curve) as well as third-order cycles (EHST vs. HST) (adapted from Batt, 1996 and Brett and Baird, 1994).

We tentatively suggest that *C. nana*, being spinose and therefore possibly specialized for environments with soft substrates, is better suited for regressive facies rather than sediment-starved transgressive facies in which sediment becomes trapped in more proximal environments.

The upper Wanakah Shale and lower Windom Shale members are both fourth-order regressive facies, and therefore, we might expect similarity in faunal composition and abundance. This is not the case, however. *Crurispina nana* again disappears within the lower Windom, *Sinochonetes lepidus* becomes relatively rare, and *Longispina mucronata* is present for the first time among the samples studied (Figure 6). Despite these faunal differences, upper Wanakah and lower Windom sediments are indistinguishable in trace metal studies of paleoenvironmental proxies for sedimentation rate, redox conditions, and paleoproductivity, as well as in measures of the degree of bioturbation (Zambito, 2006). This may imply a third-order sea level cycle control on this biofacies: the upper Wanakah corresponds to a level high in the preceding HST, whereas the lower Windom beds are part of the following EHST deposits and immediately overly the maximum flooding discontinuity in Erie County (Figure 2). Possibly environmental differences associated with these systems tracts may include slight differences in the degree of aggradation versus progradation.

Although not recovered in our lower sample intervals, *Longispina mucronata* does occur at other levels in the Wanakah Shale Member, in particular in the interval up to and just below the Fargo Bed, which occurs slightly below the section present at Lake Erie Shore at the mouth of Eighteenmile Creek (Miller, 1991) — that is, slightly below the interval examined in this study. *Longispina mucronata* has also been observed in the lower Ledyard Shale Member (Savarese *et al.*, 1986; Brett and Baird, 1994). In the Penn Dixie and Buffalo Creek Pyrite beds of the upper Windom Shale in Erie County (see Dick and Brett, 1986), *Crurispina nana* is observed by the present authors to occur in abundances similar to those of the Spring Brook concretionary interval (unpublished data). *Crurispina nana* occurs even more abundantly at certain intervals in the uppermost fourth-order regression of the Levanna Shale Member (Zambito, 2006). These observations demonstrate that a similar representation of the *Ambocoelia*-chonetid biofacies (one that includes common *C. nana*) recurs during three different third-order depositional sequences: the uppermost Levanna, the Spring Brook concretionary interval of the upper Wanakah, and the Windom Pyrite beds. These observations also suggest that fourth-order systems tracts may be necessary to explain biofacies compositions at the thin intervals (10cm) sampled.

Two other major faunal members of this biofacies, *Ambocoelia umbonata* and *Paleoneilo* sp., exhibit only relatively minor differences in abundance and presence/absence in the intervals studied. All other observations, however, suggest a complicated interplay between third- and fourth-order cycles as presently defined for the Middle Devonian Hamilton Group. In some cases, deposits with similar third- and fourth-order cycle positions show some similarities in biofacies composition and abundance, as is the case for the upper Wanakah and the upper Levanna (see Zambito, 2006) (fourth-order regressive facies in a third-order HST), as well as the lower Windom and presumably the lower Ledyard (fourth-order regressive facies in a third-order EHST). Unfortunately, the observations along with the systems tract delineations to date do not form a unifying relationship between third- and fourth-order cycles and biofacies composition and abundance, but rather they emphasize the complexity of obtaining similar samples for tests of faunal stability.

CONCLUSIONS

Of concern in studies of community stability is not only the reproducibility of samples collected, but also the reproducibility of the community to be sampled. Variability due to environmental fluctuations is a reality in any biological system, yet if every effort to sample a given biofacies results in subtle or not so subtle differences in faunal composition which must then be attributed to subtle differences in the sedimentary or environmental regime, then the hypothesis of habitat tracking becomes effectively untestable and degenerates into a tautological truism. This study focused on differences in the *Ambocoelia*-chonetid biofacies and changes in community structure through both transgressive and regressive facies, as well as at the scale of both third- and fourth-order sea level cycles. Differences in community structure were evident between fourth-order transgressive and regressive facies as well as between two fourth-order regressive facies at different positions within third-order cycles despite similarities in geochemical and ichnofabric indicators of environmental conditions. These differences are seen in not only the presence/absence of organisms, but also in their abundance. Furthermore, both third- and fourth-order sea level cycles appear to influence community membership as well as abundance, but a synthesis of the relationships of these two sea level cycle orders is not currently possible because of remaining questions about the differentiation of fourth-order systems tracts for all intervals studied.

Implications of the observed community patterns for the hypothesis of habitat tracking are that there are cases in which it appears that some organisms track a preferred habitat, as is the case with the upper Levanna Shale Member, Spring Brook concretionary interval, and upper Windom Pyrite beds; and the taxonomic composition and abundance of the biofacies is therefore somewhat reproducible. There may be other factors, however, either unresolved environmental (temperature?) or perhaps even biological, that affected the distribution of these organisms since there is no consistent relationship between faunal abundances or presence/absence and position in third- and fourth-order cycles given the currently available systems tract differentiations for the Hamilton Group.

The variability in the *Ambocoelia*-chonetid biofacies observed has important implications for studying community stability over longer time periods. Exactly what is the 'type' *Ambocoelia*-chonetid biofacies? Are tests of community stability more appropriate over longer time periods (member, formation, or group-level), and therefore higher order sea level cycles, rather than samples representing only a relatively small number of bedding planes (see Baugh *et. al*, 2005)? Further work is necessary to fully understand the effect that community variability at different scales may have on tests of faunal stability. The present study suggests that the starting point to any study, however, is to gain an understanding of this variance and the factors responsible prior to selecting limited horizons for samples that represent a variable biofacies.

ACKNOWLEDGMENTS

This research of the *Ambocoelia*-chonetid biofacies has benefited from discussion with Dr. H. David Sheets, Dr. Carl Brett, and in particular Dr. Gordon Baird. Field work was aided by Rob Poczkalski of the Buffalo Geosciences Program. Research was supported by grants from the Geological Society of America, Sigma Xi, and the Mark Diamond Research Fund (University at Buffalo, SUNY).

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ROAD LOG

Mileage		Instructions
0.0	0.0	Leave Adam's Mark Hotel, going west toward Bingham St.
0.1	0.1	Merge onto I-190 S
5.2	5.1	Merge onto I-90 W
6.8	1.6	Take Exit 54: Route 400, West Seneca/East Aurora
7.3	0.5	Merge onto NY Route 400 S
12.2	4.9	Exit US 20 E/NYS Route 78/Transit Rd heading northbound
12.6	0.4	Turn right on Bullis Road
15.5	2.9	Cross Bowen Road
17.0	1.5	Cross Girdle Road
17.2	0.2	Cross Buffalo Creek, take immediate right after bridge, park near barrier
		Proceed down old road toward creek bed

Stop 1: Buffalo Creek

This stop focuses on the lowermost Windom strata, an *Ambocoelia*-rich gray shale. The base of the Windom Shale Member at this locality occurs directly above the Little Beards Creek Phosphate Bed which marks the disconformable contact of the Kashong and Windom Shale Members. This contact and the lower portion of the Ambocoelia-rich interval is accessible on the east-facing bank of Buffalo Creek almost immediately downstream of the new Bullis Road Bridge. The upper portion of the *Ambocoelia*-rich mudstones as well as the Smoke Creek Trilobite Bed are most easily accessed on the west-facing bank immediately upstream of the old Bullis Road Bridge.

The *Ambocoelia*-rich interval exposed at this locality represents the thickest exposure of this lower Windom unit on this trip, possibly representing a sub-basin (Gordon Baird, personal communication). The basal portion of these beds is a fissile and dark gray shale with only rare *Ambocoelia umbonata*, with a thickness of about 1 meter. This grades into a gray mudstone rich in both *Ambocoelia umbonata* as well as chonetids, including *Longispina mucronata*, *Sinochonetes lepidus*, *Devonochonetes scitulus*, and rare *Devonochonetes coronatus*. Near the top of the Ambocoelia-rich shales are at least six concretionary bands rich in styliolinids.

Ambocoelia umbonata becomes rarer above the second concretionary bed. The Smoke Creek Trilobite Bed attains a thickness of almost 1 meter at this locality and is underlain by approximately .5 meters of the Bay View Coral Bed.

The *Ambocoelia*-rich interval was deposited under regressive conditions, evidenced by the fissile dark gray shale below and the Bay View Coral Bed (requiring more oxygenated conditions) above. This interval is unique among those observed on this trip in that while it is representative of the *Ambocoelia*-chonetid biofacies, the absence of *Crurispina nana* and the presence of *Longispina mucronata* distinguishes this occurrence of the *Ambocoelia*-chonetid biofacies from the Wanakah Shale intervals described.

Mileage		Instructions
		Return to vehicles and proceed west on Bullis Road
21.8	4.6	Turn left (southbound) on US 20/NYS Route 78/Transit Road
22.5	0.7	Cross NYS Route 16/Seneca Street
23.2	0.7	Cross over Cazenovia Creek
23.5	0.3	Turn left onto Transit Road and park near Kinsley Road
		Proceed on foot back across US 20 Bridge over Cazenovia Creek
		Enter creek bed via driveway and trail on NW side of creek

Stop 2: Cazenovia Creek

The lower Windom and upper Wanakah Shale Members outcrop on the north-facing bank of Cazenovia Creek, on either side of the US 20 overpass. Only the upper Wanakah is readily accessible at this locality, specifically the strata of and between the Spring Brook and Walden Cliffs concretionary intervals; occurring from the bed of the creek to approximately midway up the outcrop. This medium to dark gray shale unit contains abundant pyrite in the form of burrow steinkerns, pyritized fossils, and also pyritized shell hash horizons, which are commonly encased in concretions. A detailed description of these units, in particular the Spring Brook concretionary interval and its diagenetic history is described by Jordan (1968).

Unlike the lower Windom Shale observed at Stop 1, the Upper Wanakah shale interval contains abundant *Crurispina nana*. The intervals in which there is a presence of *Leiorhynchus* are indicative of a deeper water and lower oxygen biofacies, which corresponds to a slight decrease in the abundance of ambocoeliids.

Mileage		Instructions
23.6	0.1	Return to vehicles and proceed west on US 20
25.9	2.3	Cross Michael Road
27.4	1.5	Turn right on Milestrip Road
28.9	1.5	Turn left on Abbott Road
29.0	0.1	Park behind medical office (first driveway on left)
		Proceed on foot to creek bed

Stop 3: Smoke Creek

The lowest strata exposed at this stop is the uppermost 1 meter of the Wanakah Shale Member. Overlying the Wanakah, where it has not been eroded, are pods of the Hills Gulch Bed of the basal Jaycox Shale Member. The cap of the falls at this locality is the Tichenor Limestone, a crinoidal grainstone containing many abraded fossil fragments indicative of a depositional depth near wave base (Brett and Baird, 1994). Immediately overlying this is the *Ambocoelia*-rich shales observed at Stop 1. No styliolinid-rich concretionary beds are present at this locality, and specimens of *Ambocoelia umbonata* are typically deformed, presumably due to compactional stress. The *Ambocoelia* shale interval is overlain by the Bay View Coral Bed and the Smoke Creek Trilobite Bed. This stop is the type section of the Smoke Creek Trilobite Bed, a well-developed calcareous gray shale interval with abundant *Phacops* and *Greenops*. Similar to the interval exposed at Buffalo Creek, this occurrence of the *Ambocoelia*-chonetid biofacies lacks *Crurispina nana* and contains *Longispina mucronata*.

Mileage		Instructions
		Return to vehicles and turn right on Abbott Road
29.1	0.1	Turn left on Milestrip Road
30.4	1.3	Cross I-90
31.8	1.4	Merge onto NYS Route 5 W
33.2	1.4	Cross Big Tree Road
34.7	1.5	Cross Rogers Road
35.9	1.2	Cross Amsdell Road
36.8	0.9	Turn right on Old Lake Shore Road
38.6	1.8	Cross Lakeview Road
39.9	1.3	Turn left into Eighteenmile Creek DEC pull-off
		Proceed across bridge back over Eighteenmile Creek
		Follow footpath downstream to north bank of creek mouth at Lake Erie

Stop 4: Lake Erie Bluffs

The exposure at the mouth of Eighteenmile Creek spans an almost complete Wanakah Shale section, with both the Tichenor Limestone and the basal Windom Shale Member visible when looking south along the shoreline. The interval examined, and easily accessible, at this section spans approximately 2 meters both above and below the concretionary trilobite beds of the lower Wanakah Shale Member.

Similar to the previous Wanakah Shale Member section examined at Cazenovia Creek, *Longispina mucronata* is absent from this interval. *Crurispina nana*, while present, is extremely rare and has only been observed by the present authors in a few horizons (Figure 6).

End of trip. Return to Adam's Mark Hotel via NYS Route 5 (Main Street) to Church Street.

Famous Hazardous Waste Sites and Fractured Rock Hydrology of the Niagara Falls Area, Niagara County, New York State

Guidebook for the Field Trip Held
October 7, 2006
in conjunction with the
78th Annual Meeting of the
New York State Geological Association
Buffalo, New York

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A. Introduction

More than 200 waste disposal sites have been identified within 5 kilometers of the Niagara River, which connects the largest freshwater system in the world (Yager, 1996). Chemical contaminants are likely to have leaked from one-third of these waste sites. Many of these hazardous waste facilities have discharged waste into the Lockport Group that underlies the region. The Silurian-aged Lockport Group (or Lockport Dolomite according to USGS designation) is fractured along highly continuous bedding planes that bear and conduct water. The hydrogeology of the Niagara Falls Area, and therefore the ground-water contamination, are dominated by fractured rock hydrology.

The objective of this field trip is to provide a first hand exposure to some of the most well-known ground-water contamination sites in the context of the regional hydrogeology of the Niagara Frontier. We will visit four Superfund sites, Love Canal, Bell Textron, 102 Street, and Hyde Park, and then view the stratigraphy of the Lockport group at the Niagara Gorge. Along the way, general principles of fractured rock hydrology will be discussed. Because this trip is held on a Saturday, we will not have direct access to any of the sites themselves, but we should be able to see enough from the road to provide the hydrogeologic setting.

Figure 1. Overview of the Niagara Falls study with trip stops.

Ground-Water Contamination of Niagara Falls, New York

The history of ground-water contamination of the Niagara Falls region is as old as the history of its industrial development. Although the root of the contamination is often blamed upon ignorance of environmental systems, this was not always the case. For example, in a memo to the US Army Corps of Engineers, the superintendent of the Linde Ceramics plant, A. R. Holmes, described alterative options for disposal of radioactive waste generated at the plant (Kelly and Ricciuti, 2006). They might (Plan1) discharge the waste to a storm sewer or (Plan 2) pump the waste into onsite wells. "Plan 1 is objectionable," Holmes wrote, "because of probably future complications in the event of claims of contamination against us. Plan 2 is favored because our law department advises that it is considered impossible to determine the course of subterranean streams and, therefore, the responsibility for contamination could not be fixed." Ultimately, nearly 50 million gallons of liquid wastes were pumped into these shallow wells where they likely discharged to the Niagara River.

The fact that the bedrock ground water and Niagara River are closely linked is certainly recognized now. In 1987, a Declaration of Intent was signed by authorities in both the United States and Canada which included a commitment to reduce the toxic substance loadings to the Niagara River fifty percent (50%) by 1996. In 1989, the U.S. Environmental Protection Agency (EPA) and New York Department of Environmental Conservation (DEC) issued a report identifying 33 site clusters with potential for polluting the Niagara River and proposed a remediation schedule to reduce toxic chemical loadings from these site by 99% by 1996 (**Appendix A1**). This list was later reduced to 26 sites, and it was estimated that a 90% reduction of toxic loadings to the Niagara River had been achieved by the year 2000 (U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2000). Clearly, there has been a significant reduction in toxic loadings to the Niagara River.

This remarkable reduction in toxic loadings to the Niagara River has come at a cost of over $370 million (New York State Department of Environmental Conservation, 1985).

Current schedules call for the remainder of the 26 priority sites to be remediated by 2003, with additional costs of remediation exceeding $261 million. Such enormous expenditures have been justified because the threat to human health was considered critical and immediate. As remediation nears completion at many of these sites, new questions are emerging. If over half-a-billion dollars is a justifiable expenditure to reduce the loadings by 99%, what is a justifiable cost to eliminate the remaining 1%? At most of the 26 hazardous waste sites, the term "remediation" really means containment in perpetuity, with ground-water extraction wells producing millions-of-gallons of water that must be treated. Over what period of time will treatment be cost-effective or even necessary? Once the 26 identified hazardous waste sites have been controlled or remediated, should other sites be revisited and included in the toxic loading calculations? How can the costs of remediated specific hazardous waste sites be compared to the controlling other sources of contamination, say from industrial runoff and other less-obvious non-point sources?

The Hydrogeology of the Niagara Falls, New York

Excellent summaries of the hydrogeology of the Niagara Falls Region have been published and will not be repeated here (Novakowski and Lapcevic, 1988; Novakowski, 1998; Tepper et al., 1991; Yager, 1996, 1998; Yager and Kappel, 1998) . Bedrock geology is dominated by the dolomite of the Niagaran Series (Middle Silurian), which strikes east-west and dips gently to the south at about 4.7 m/km . The Lockport Group crops out along the Niagara Escarpment, where it forms the cap rock of Niagara Falls. The Lockport Group is a petroliferous dolomite that contains gypsum and metal sulfides (Zenger, 1965). Naturally derived hydrocarbons are disseminated throughout the rock matrix in some stratigraphic horizons. Thin layers of bitumin are also present and are commonly associated with layers of gypsum. The stratigraphic nomenclature of the Lockport Group was recently revised by Brett and others, and is shown in **Appendix A2.**

Ground water is thought to flow regionally through the Niagara Falls Area along bedding plane fractures of the Lockport (Novakowski and Lapcevic, 1988). These sub-horizontal bedding plane fractures extend over kilometers. Water bearing fractures identified in Niagara Falls sequences have been identified to be conductive also in Smithville, Ontario, over 40 kilometers to the west (Novakowski et al., 1999). These horizontal bedding plane fractures are thought to be connected by steeply dipping or vertical fractures but such features have never been directly identified in boreholes.

Ground water flows away from a topographic high near the Niagara Escarpment (Figure 2). Northward water discharges to the Escarpment. Southward ground water moves generally toward the southwest where it discharges to the Niagara River. Ground water potential is highly influenced by a number of major man-made structures. The New York Power Authority (NYPA) reservoir recharges the Lockport as does leakage from municipal water supply and storm drains. Excavations act as high-permeability conduits for ground water. The most significant of these is the Falls Street Tunnel, an unlined storm sewer.

Based on information available in 1987, the U.S. identified the Falls Street Tunnel, a major unlined industrial sewer cut into the bedrock under the City of Niagara Falls, as the largest source of toxic pollutants from any of its point sources (U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2005). By the mid-1980s, the Tunnel was only receiving overflows of wastewater from the sewers of a Niagara Falls industrial area, in addition to contaminated groundwater infiltrating from major waste sites via cracks in the Tunnel's bedrock walls. In contrast to flows from other point sources, effluent from the Falls Street Tunnel entered the Niagara River untreated. In 1993, EPA and DEC required the City of Niagara Falls to treat the Falls Street Tunnel discharges during dry weather at the Niagara Falls WWTP. Data gathered by the U.S. indicate that WWTP treatment of the Tunnel's dry weather discharge has reduced mercury loadings by 70% relative to 1980 loads, tetrachloroethylene loadings by 85%, and the loadings of four other priority toxic chemicals by almost 100%

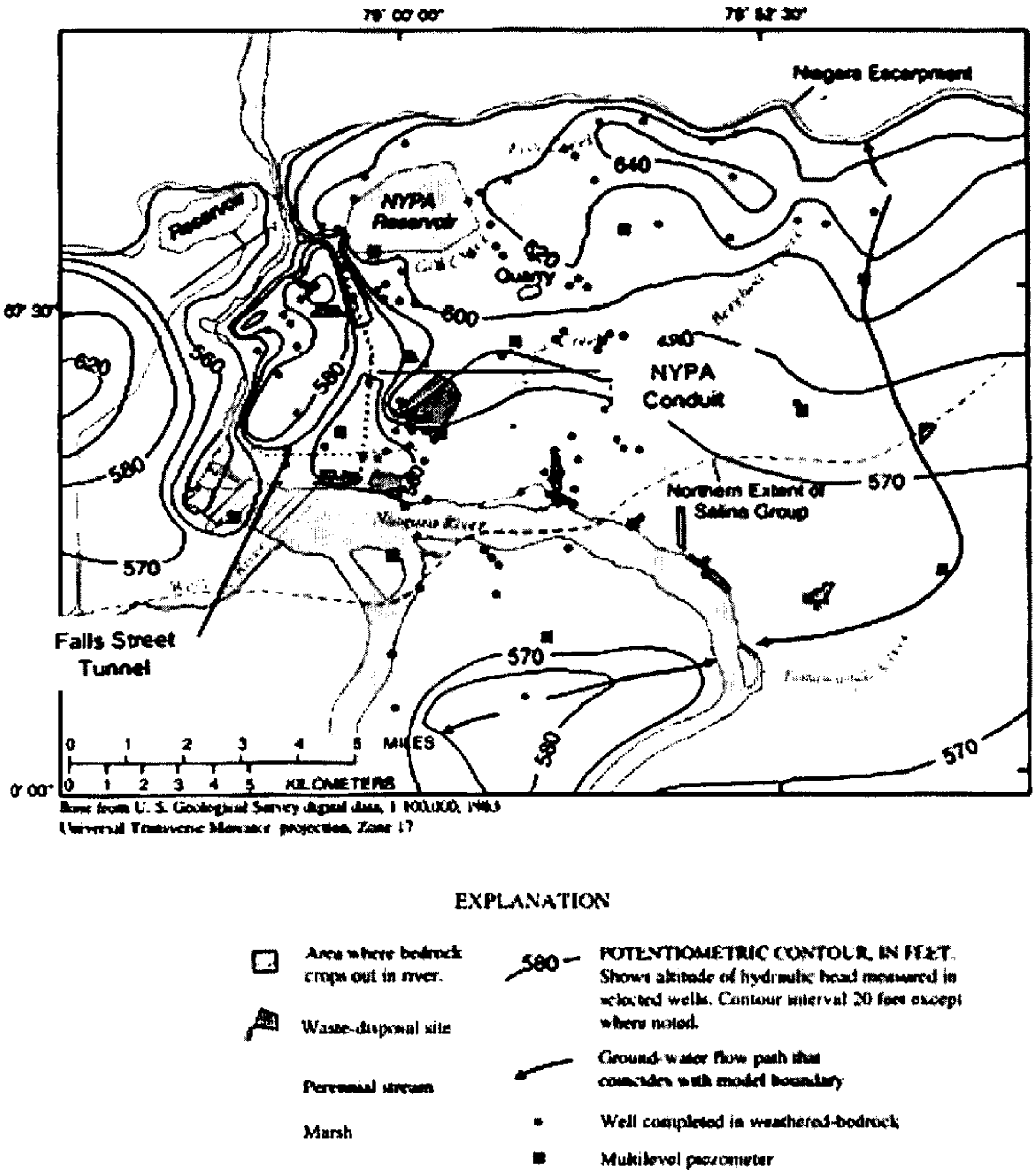


Figure 2. From Figure 3A in Yager (1996). Ground-water potential in the weathered bedrock measured in selected wells. The Falls Street Tunnel and NYPA Conduits are indicated.

The NYPA conduits that transport Niagara River water to the NYPA forebay canal also constitute a major sink and pathway for ground water. A drain system extends the length of the conduits and intersects the entire Lockport Group. A grout curtain surrounding the intakes prevents direct hydraulic communication between the river and the drain system (Yager, 1996).

B. Description of the Stops

Bell Aerospace Textron Site

The Bell Aerospace Textron Site is located in Wheatfield, New York, adjacent to the east site of the City of Niagara Falls. In the 1950's and 1960's TCE (triochloromethane) was used in the production process. Waste TCE was discharged to a shallow "neutralization" pond which infiltrated to the Lockport Group. By 1990, a 280-acre aqueous plume in the Guelph Dolomite (**Figure 3**) that contained TCE and its metabolites (DCE, dichloroethane and VC, vinyl choride) extended about 4,300 ft south of the pond. A 20-acre dense, nonaqueous phase (DNAPL) plume of TCE extended 620 ft south of the pond. A pump-and-treat remediation system consisting of six wells near the pond and five wells 2,900 ft downgradient from the pond began operation in 1993 to decrease the size of the aqueous plume and prevent its further migration.

The study by Yager (2000) illustrates some of the interesting characteristics of contaminant transport in bedrock systems. Flow and transport is through individual fractures with hydraulically estimated apertures of 1 to 1.5 mm. The transmissivity estimated through model calibration is 140 m^2/day. As is often the case in bedrock studies, it is difficult to estimate the effective porosity. Yager (Yager, 2000) modeled effective porosity between 0.3 and 3 percent. The smaller number was consistent with the hydraulic aperture estimates and the larger was more consistent with transport modeling. Ground water velocities are thought the be on the order of meters per day, but are difficult to confirm given the uncertainty in effective porosity. The plume (**Figure 3**) is nearly as wide as it is long, reflecting the similarity in transverse and longitudinal dispersion that is thought to be characteristic of transport in fractured bedrock. Another important consideration in transport is the exchange in contaminant mass between fractures and rock matrix through a process known as "matrix diffusion". Molecular diffusion becomes an important consideration in fate and transport through dual porosity media. Models of plume transport and biodegradation were extremely sensitive to the rate of matrix diffusion assumed. Matrix diffusion is difficult to measure independently for application to field studies.

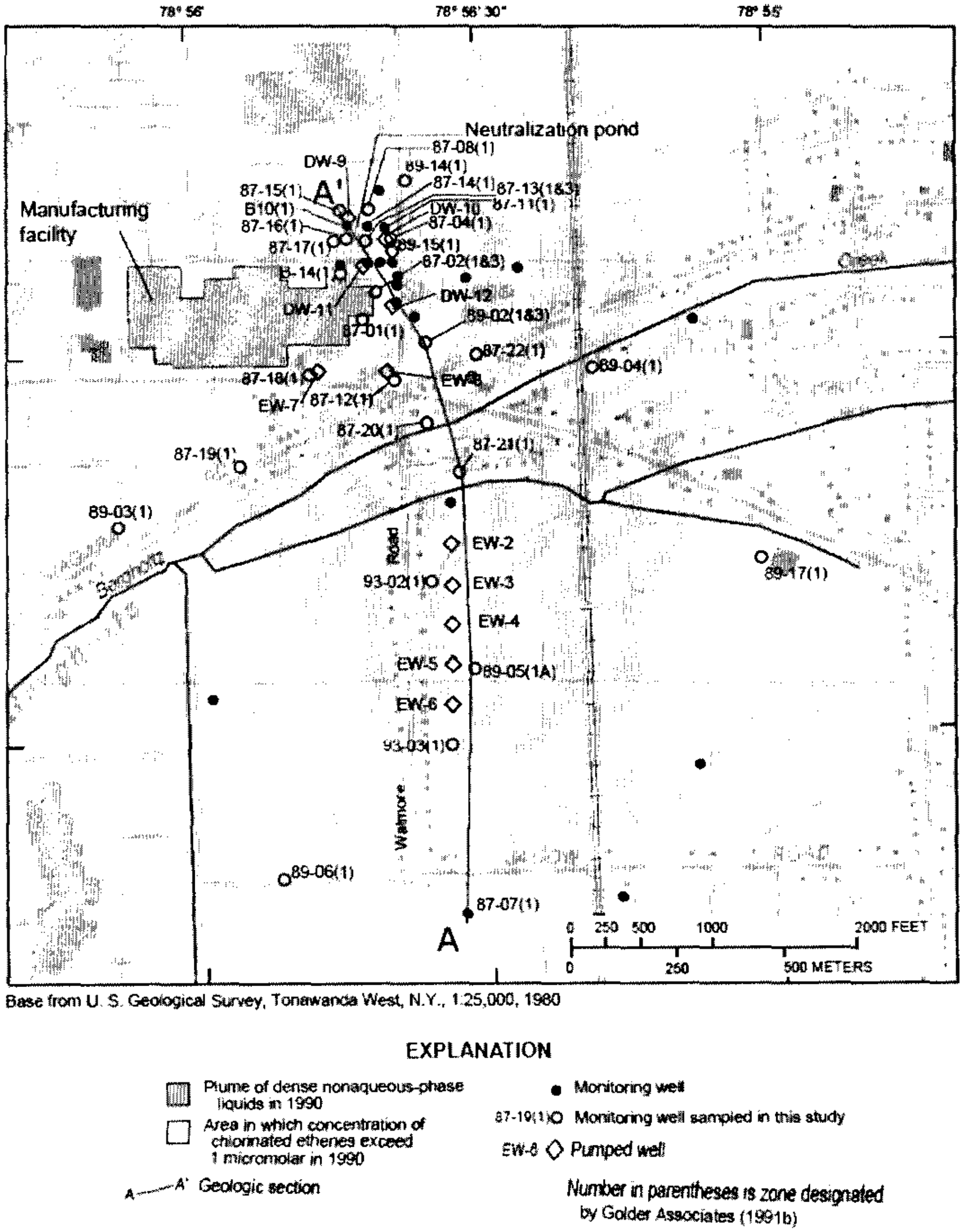


Figure 3. Taken from Yager et al., 2000. The extent of the plume of aqueous phase (APL) and dense non-aqueous phase (DNAPL) chlorinated solvents at the Bell Textron Site.

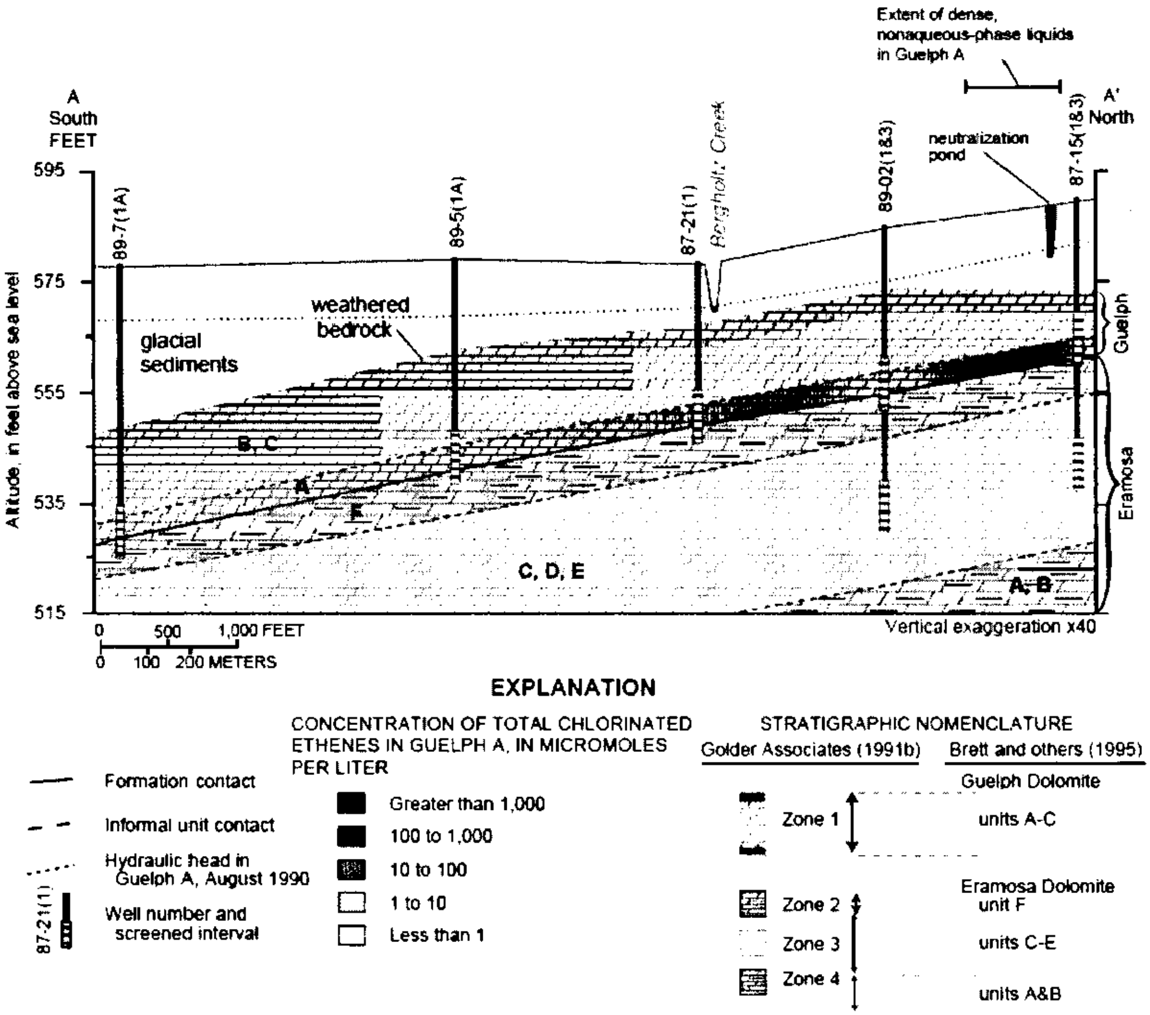


Figure 4. Taken from Yager et al., 2000. Vertical section A-A' through the Bell Textron Site (see Figure 3 for section location).

The plume at the site appears to have reached a dynamic equilibrium between the rate of TCE dissolution and the rate of removal through pumping and biodegradation. The presence of dissolved mass in the rock matrix and non-aqueous phase pools in fractures provides a constant source of TCE. Degradation rates for TCE are 21 to 25 days, DCE are 170 to 230 days, and VC are 18 to 23 days. Consequently, if the source could be remove the plume would disperse rapidly. Removal of all the contamination source in the rock and fractured bedrock is consider impractical given current remediation technology. A summary of the site is provided in **Appendix B1** (U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2005).

Love Canal

It is unlikely that any reader has not at least heard of Love Canal. It is, perhaps, the most famous ground-water contamination site in the country and launched the creation of the Superfund program in the United States. A brief History is presented in **Appendix B1**. The site was removed from the EPA's National Priority Listing (Superfund List) in 2004 and its final remediation actions have been implemented. The hydrogeology underlying Love Canal consists of a shallow system of silts and fine sands, underlain by confining layers of lacustrine clays and glacial till, which are underlain by the Lockport Dolomite. Wastes were emplaced in the canal excavated as part of William T. Love's plan to create a canal that connected the Niagara River to Lake Ontario. The site resembled a geologic bathtub filled with waste and capped with soil. After wet periods, leachate would rise to the surface in swales where humans could come in direct contact with the material. Leachate also seeped into nearby basements.

The 1988 Superfund record of decision for cleanup of the site was "The selected remedial action for this site includes: excavation and solidification/stabilization of 7,500 yd^3 of soil; placement of solidified soil back in excavated location; installation of a RCRA cap; ground water monitoring; and implementation of treatability studies for solidification process." This essentially is the solution that was ultimately implemented at the site. Occidental Chemical owns the site and operates the treatment facility housed on the property. Hazardous materials are stripped from the ground water and sent to an incineration facility in Texas for ultimate disposal.

The hydrogeology underlying Love Canal consists of a shallow system of silts and fine sands, underlain by confining layers of lacustrine clays and glacial till, which are underlain by the Lockport Dolomite. The sediments are believed to form a liner that prevented extensive contamination of the bedrock. Constant pumping within the excavated canal appears to have reversed the ground-water flow at the canal and removed dissolved contamination from both the bedrock and the overburden sediments.

OCC S-Area

The S-Area site is an eight-acre landfill on Occidental Chemical Corporation's (OCC) Buffalo Avenue Plant. The site is located approximately 200 yards north of the Niagara. The site was used primarily from 1947 to 1961 for the disposal of approximately 63,000 tons of organic and inorganic chemicals. Chemicals deposited at the site included chlorobenzenes, organic phosphates, acid chlorides, phenol tars, thionyl chloride, chlorendic acid, trichlorophenol, benzoyl chloride, liquid and chlorotoluene-based disulfides, metal chlorides, thiodan, and miscellaneous chlorinated hydrocarbons. The EPA Fact Sheet for this site is provided in **Appendix B3.**

The S-Area Landfill is historically significant because it was at this site that the term "non-aqueous-phase-liquid" (NAPL) was first used (Pankow and Cherry, 1996). The landfill is located immediately adjacent to the form City of Niagara Falls Water Treatment plant. The plant drew water from the Niagara River via a bedrock tunnel. Contaminants from the S-Area landfill leaked into this tunnel, contaminating water supplies. The entire treatment plant was abandoned and a new plant constructed in 1997. The site of the former water treatment is clearly visible between the S-Area and new treatment plant.

The contaminated ground water flowed toward the treatment plant intakes and the Niagara River prior to remediation. Ground-water flow was through three zones: (1) overburden sediments, (2) shallow weather bedrock, and (3) deeper bedrock (**Figure 5**). Due to the sites close proximity to the Niagara River, contaminated ground water discharged to the Niagara River or to bedrock beneath the river. The remediation strategy is containment. A combination of pumping, drains, and slurry walls are used to create an inward hydraulic gradient at the site. Effluent is treated onsite and then discharged to the Niagara River.

Unlike the Bell Aerospace Textron site, little is known about the natural attenuation of chlorinated solvents. This is essentially because the plume extends immediately to the Niagara River making water quality observation difficult.

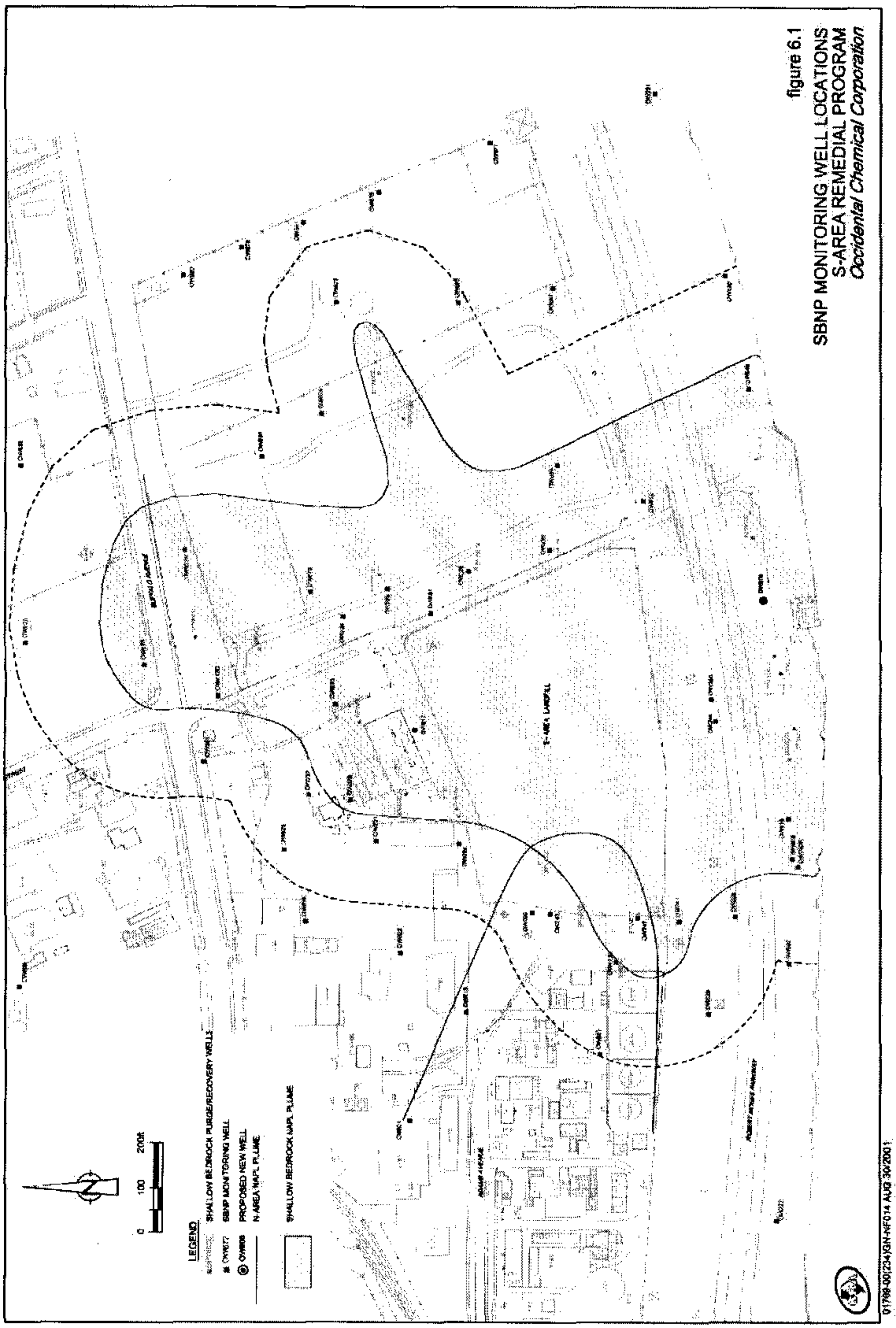


Figure 5a. Extent of the S-Area contamination in the shallow bedrock (compliments of Martin Derby).

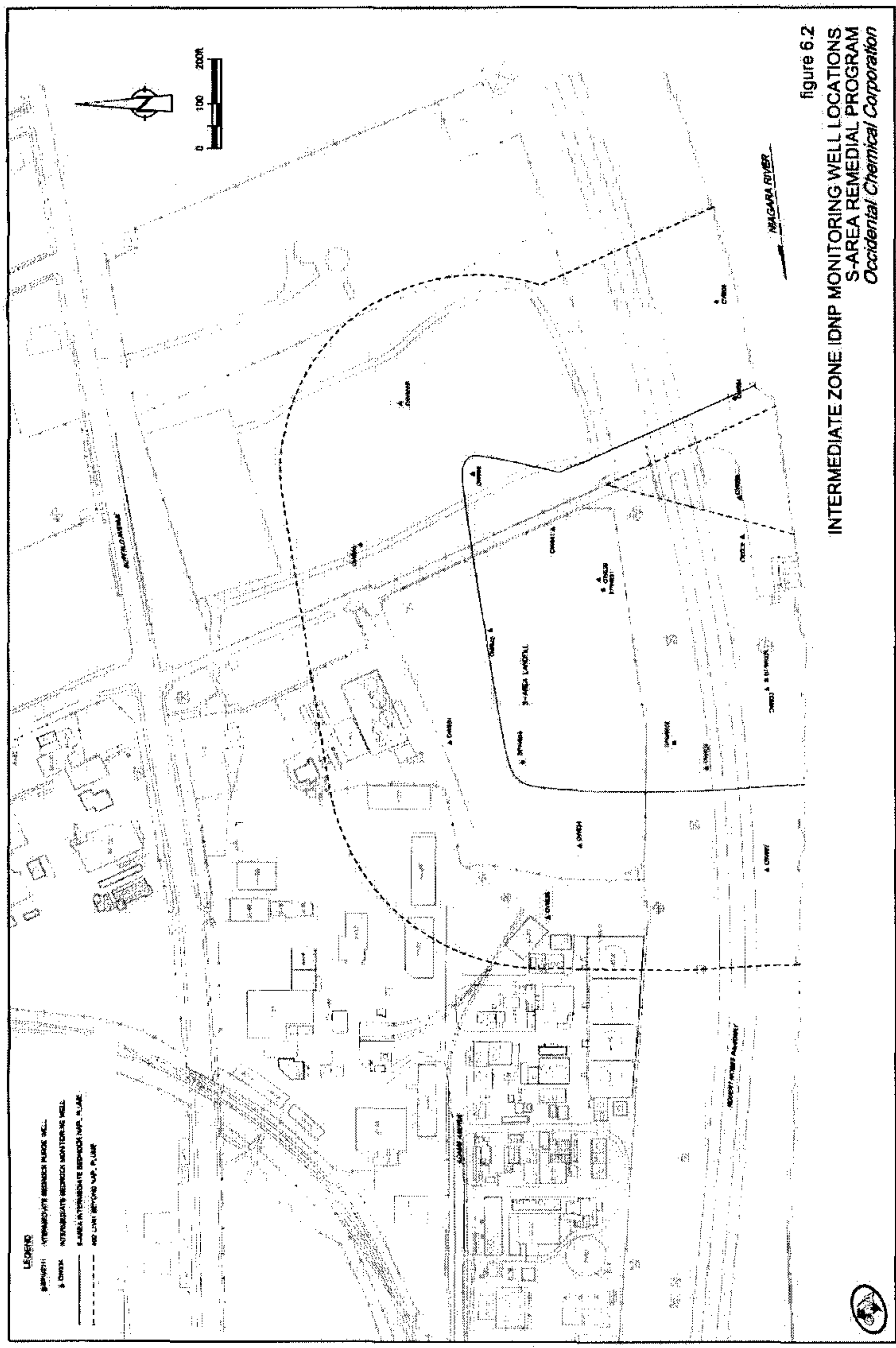


Figure 5b. Extent of the S-Area contamination in the intermediate bedrock zones (compliments of Martin Derby).

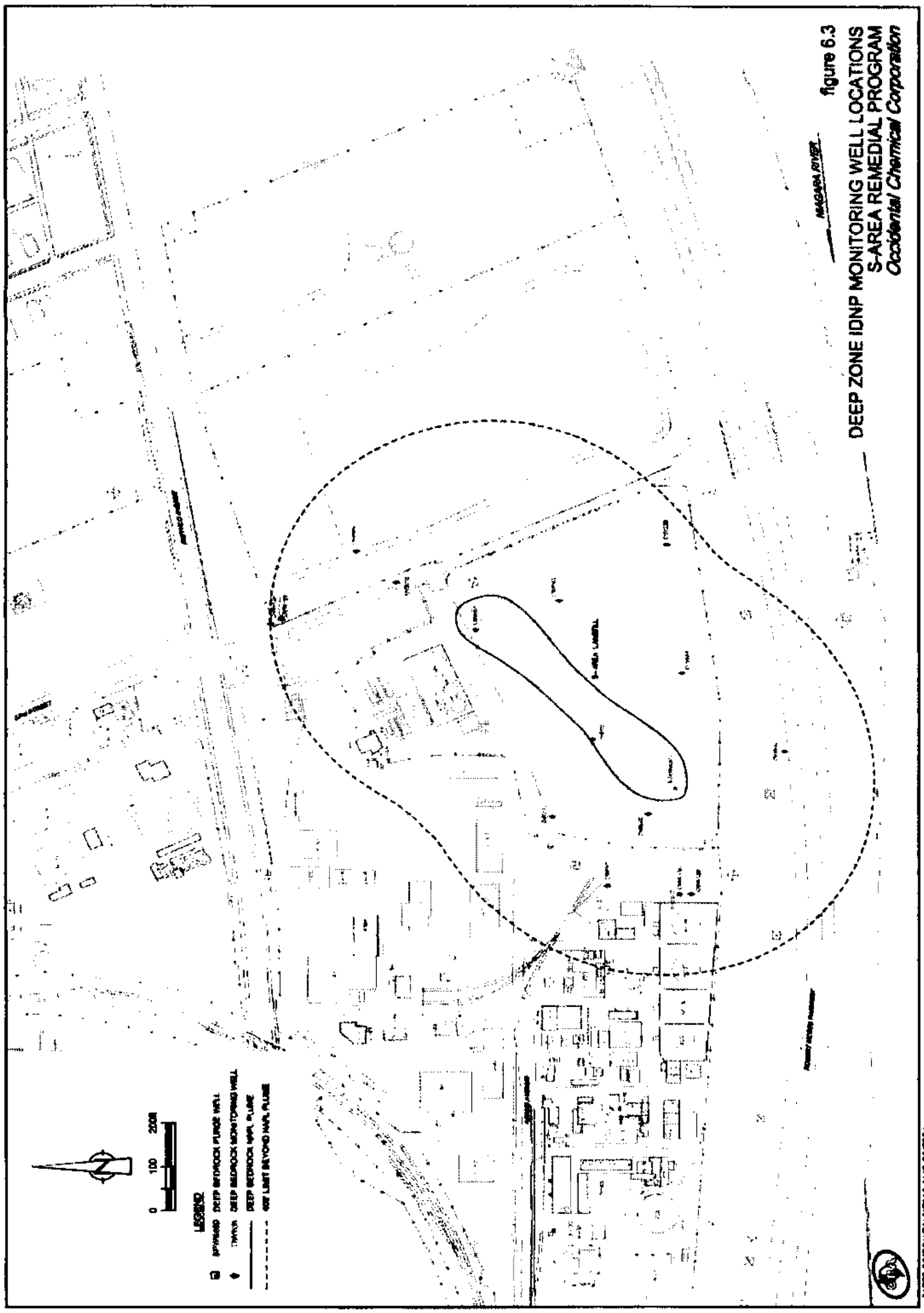


Figure 5c. Extent of the S-Area contamination in the deep bedrock zones (compliments of Martin Derby).

Hooker (OCC) Hyde Park

Occidental Chemical Corporation's (OCC) Hyde Park site is a 15-acre landfill in northwest Niagara Falls, less than one-half mile from the Niagara River. From 1953 to 1975, the company (then Hooker Chemicals and Plastics) deposited approximately 80,000 tons of chemical wastes at the site. This is arguably the largest mass of DNAPL contamination site in the United States. The hazardous materials disposed on site included 3,300 tons of 2,4,5-trichlorophenol (TCP) wastes, which are known to contain significant amounts of 2,3,7,8-tetrachlorodibenzo-p-dioxin (TCDD); approximately 0.7 - 1.6 tons of dioxin are believed to be associated with the TCP. Chlorinated organic wastes, including hexachloropentadiene derivatives, chlorendic acid, chlorinated toluenes, benzenes and phenols, predominate at the site. The former drainage stream of the landfill, Bloody Run, which flows into the Niagara River, was historically contaminated with organic chemicals, including dioxin. A clay cap and a shallow leachate collection system were installed at the site in 1979. A summary of remediation activities can be found in **Appendix B4**.

The site is underlain by Pleistocene overburden deposits that overlie the Lockport bedrock. Overburden sediments are glacial till, lake deposits, and some localized sand and gravel deposits. The principal water bearing zones are in the Lockport Group are the weathered bedrock surface and horizontal-fracture zones that coincide with stratigraphic contacts (Yager, 1996). The underlying Rochester Shale is thought to have much lower hydraulic conductivity than the Lockport and therefore constitutes the lower limit of possible contamination below the Hyde Park Site (S.S. Papadopulos & Associates, 2001). Historically, contaminated ground water seeped at the Niagara Gorge face east of the power dam (**Figure 6**).

The Hyde Park Site is another excellent illustration of the difficulties faced when characterizing and remediating ground water in fractured bedrock. As in the case of Bell Textron and S-Area, DNAPL has seeped into the bedrock and cannot be removed entirely. The remediation strategy is, therefore, perpetual containment. At Hyde Park, containment is achieved by the placement of pumping wells that are designed to assure a constant inward hydraulic gradient at the site boundary. In bedrock, pumping is carried out at three depths that coincide with water bearing fractures identified by local testing and regional modeling (**Figure 7**). In spite of the use of 15 purge wells over three zones, it has been difficult to capture the dissolved phase plume. A recent modeling evaluation of purge well effectiveness showed that only the southern portion of the plume was being captured in the upper bedrock zone. Most of the plume was being captured, however, in the middle and lower zones.

Actually monitoring the ground water velocity vectors has proved challenging at Hyde Park. Monitoring wells were installed radially away from the center of the contaminated area with the intention of ensuring inward hydraulic gradients. Local heterogeneities in transmissivity make such monitoring difficult. The local changes in head due to local changes in transmissivity serve to confuse the hydraulic data. The actual ground-water divide induced by on-site pumping cannot be determined with any certainty. Although the MODFLOW model calibrated to monitoring data indicates that the most of the contaminant plume is currently captured(S.S. Papadopulos & Associates, 2001), this cannot be confirmed with head monitoring data.

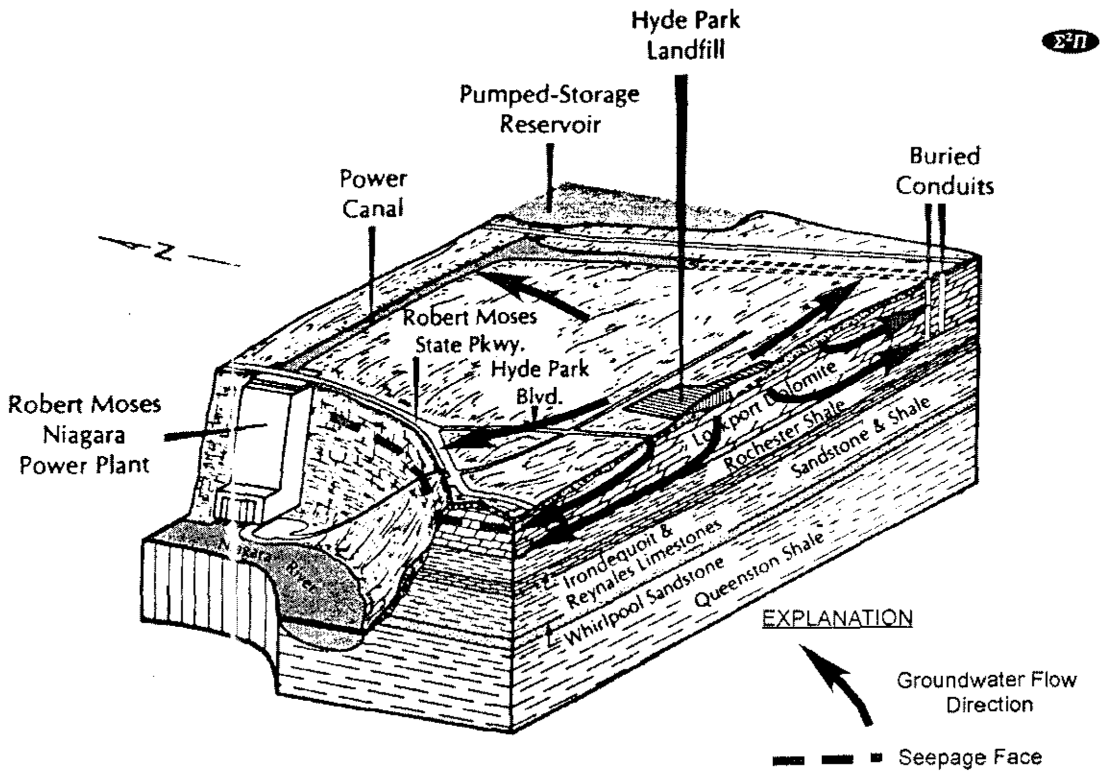
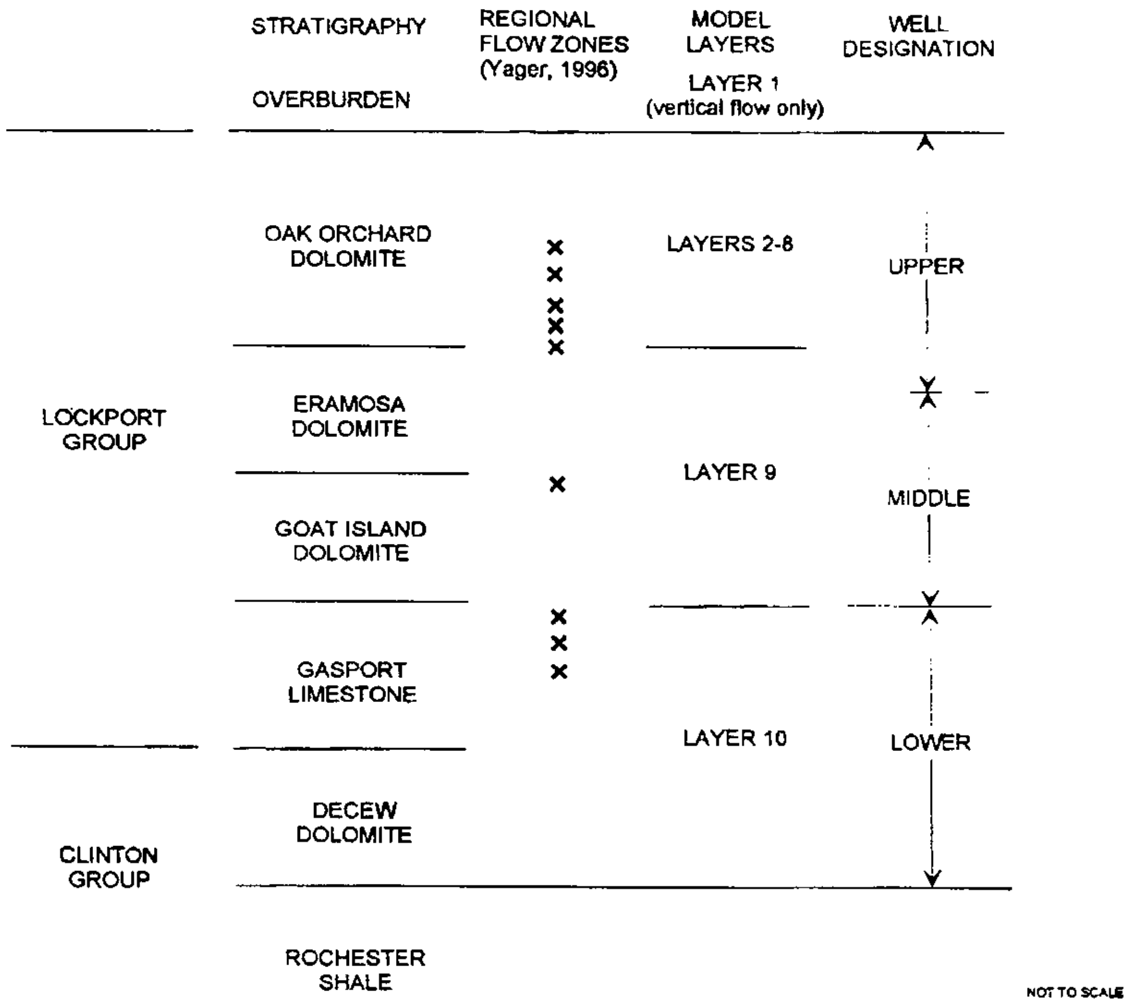


Figure 6. Conceptual sketch of ground-water flow patterns from the Hyde Park Landfill (S.S. Papadopulos & Associates, 2001)



NOTE: Layer 2 is a top-of-rock layer. Therefore, in Site area, layer 2 may correspond to the Oak Orchard Dolomite, or the Eramosa Dolomite (where Oak Orchard Dolomite is not present), or the Goat Island Dolomite (where the Oak Orchard Dolomite and the Eramosa Dolomite are not present).

Layer 3 correpsonds to Vernon Shale and is not present at the Site.
Layers 4 and 5 are the upper portion of the Oak Orchard formation, and are not present at the Site.

Figure 7. Hydrostratigraphic zones used at Hyde Park Landfill. Note that the stratigraphic terminology is that of Zenger (1965) rather than the revised nomenclature of Brett and others (1995).

NYPA Access Road (Hall Road)

This road leads to the base of the NYPA dam where a fishing platform is located. An excellent section of the Lockport and Clinton Group is exposed. Recent installation of fencing has obscured the view somewhat. About half way down the road an ephemeral seep is fenced to keep visitors away from contaminated ground water derived from Hyde Park. In spite of the hydraulic controls at Hyde Park, these seep is occasionally active. The revised stratigraphy of the Lockport and Upper Clinton are provided here (**Figures 8 and 9**).

Table 1. Bedrock stratigraphy of the Niagara Falls area

[Modified from Miller and Kappel, 1987, with additional data from Fisher and Brett, 1981; Brett and Calkin, 1987, Brett and others, 1995.]

System	Series	Group	Formation	Average thickness (feet)	Description
Silurian	Cayugan	Salina	Vernon Shale	57 (in study area)	Green and red shale.
	Niagaran	Lockport	Guelph Dolomite	33	Brownish-gray to dark gray, fine to medium, thick-bedded dolomite, with some argillaceous dolomicrite, particularly near contact with the Vernon Shale.
			Eramosa Dolomite	52	Brownish-gray, biostromal, bituminous, medium- to massive-bedded dolomite, with some argillaceous dolomicrite.
			Goat Island Dolomite	41	Light olive-gray to brownish gray, fine to medium crystalline, thick- to massive-bedded saccharoidal, cherty dolomite, with argillaceous dolomicrite near top of formation.
			Gasport Limestone	33	Basal unit is dolomitic, crinoidal grainstone, overlain by argillaceous limestone.
		Clinton	DeCew Dolomite	10	Very finely crystalline dolomite, medium to dark gray, thin to medium bedded.
			Rochester Shale	60	Dark-gray calcareous shale weathering to light gray to olive
			Irondequoit Limestone	12	Light-gray to pinkish-white coarse-grained limestone.
			Reynales Limestone	10	White to yellowish-gray shaly limestone and dolomite
			Neahga Shale	5	Greenish-gray soft fissile shale.
		Medina[1]	Thorold Sandstone	8	Greenish-gray shaly sandstone
			Grimsby Sandstone	45	Reddish-brown to greenish-gray cross-bedded sandstone interbedded with red to greenish-gray shale.
			Power Glen Shale	40	Gray to greenish-gray shale interbedded with light-gray sandstone.
			Whirlpool Sandstone	20	White, quartzitic sandstone.
Ordovician	Upper	Richmond	Queenston Shale	1,200	Brick-red sandy to argillaceous shale.

[1] Designated Albion Group by the U.S. Geological Survey

Figure 8. From Yager (1996). Detalied bedrock stratigraphy of the Niagara Falls area.

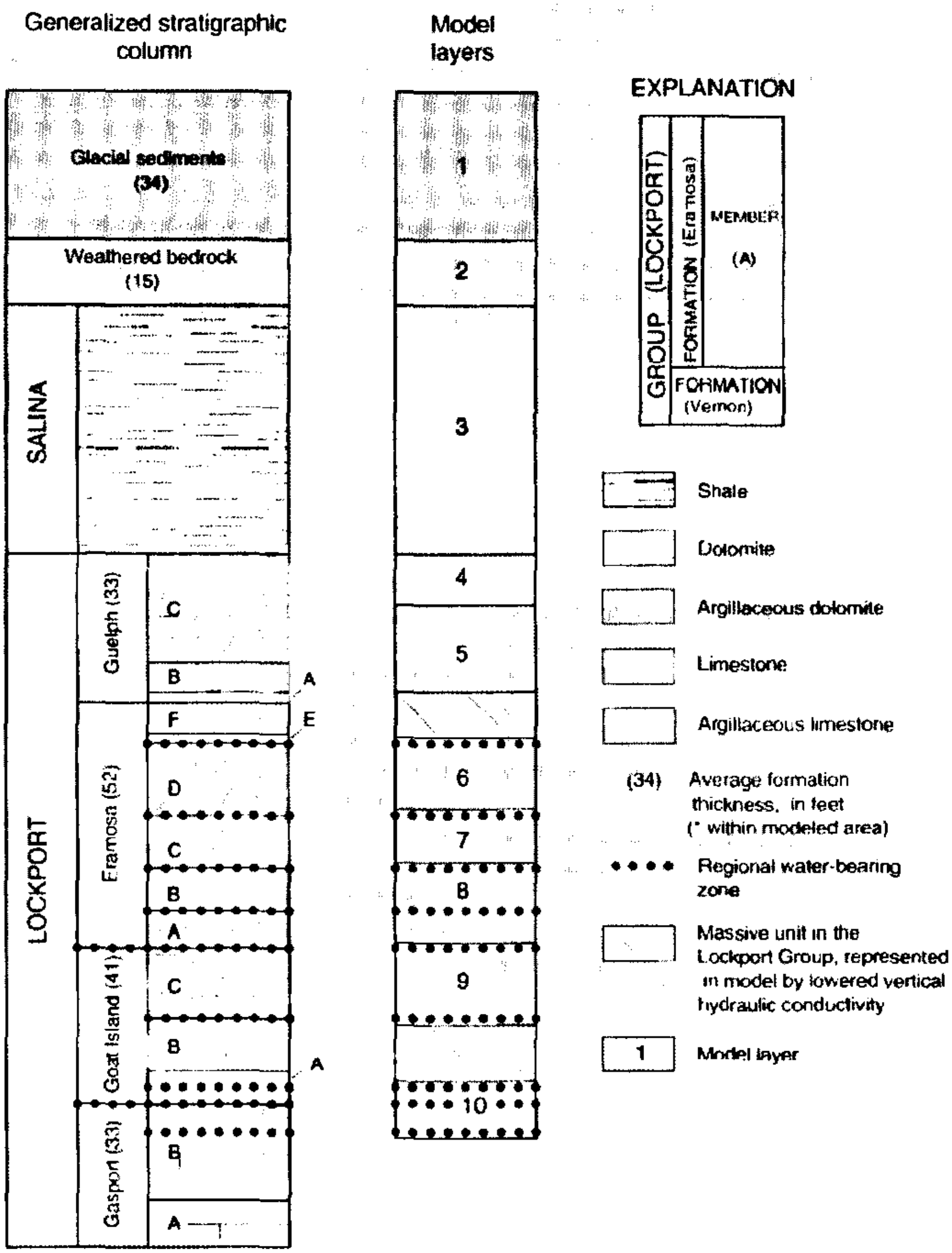


Figure 9. From Yager (1996). Bedrock stratigraphy of the Niagara Falls area showing major water bearing units. Also shown is the construction of a MODFLOW model used to represent regional ground water flow.

NYPA Power Vista Center

We will stop at the Power Vista Visitor's Center (http://www.nypa.gov/vc/niagara.htm) to get a view of the forebay canal. When the Niagara project produced its first power in 1961, it was the largest hydropower facility in the Western world at the time. The Niagara project, located about 4 1/2 miles downstream from the Falls, consists of two main facilities: the Robert Moses Niagara Power Plant, with 13 turbines, and the Lewiston Pump-Generating Plant, with 12 pump-turbines. In between the two plants is a forebay capable of holding about 740 million gallons of water; behind the Lewiston plant, a 1,900-acre reservoir holds additional water.

The excavation of the forebay into the Lockport provides a unique view of how water moves through fractured bedrock. Vegetation and water staining, along with ice formation in the winter, designate seeps along the canal walls. Notice that these seeps occur along specific bedding plane contacts, but only in specific locations. This "flow channeling" is characteristics of water flow through fractured bedrock. It implies that (1) effective porosity may be much smaller than would be estimated from fracture occurrence and (2) monitoring of contamination in bedrock may be partly a matter of luck. When effective porosity is overestimated, ground-water velocity is underestimated, based upon hydraulic arguments. This is due to the expectation in Darcy's Law that effective porosity (n_{eff}) relates specific discharge (q) and average linear velocity of a contaminant (v); $v = q/n_{eff}$.

Devils Hole State Park

If time allows we will hike down the trail that leaves from Devils Hole State Park. This trail affords an opportunity to seem more Lockport exposure up close as well as a natural cave. There are few examples of karst morphology in the Lockport Dolomite. It is unlikely that karst effects ground-water flow in the region. This cave was probably formed during the carving of the gorge by the Niagara River. It does provide an up close and personal look at the Lockport Dolomite.

C. Directions

Start at Bell Textron Site	**On Walmore Road North of Niagara Falls Blvd. in Wheatfield, NY 43.102137, -78.927063**	
	Head **south** from **Walmore Rd** - go **0.2 mi**	0.2 mi
	Turn **right** at **Cayuga Dr** - go **0.8 mi**	**0.8 mi** 1 min
	Turn **left** at **Williams Rd** - go **0.6 mi**	**0.6 mi** 1 min
	Turn **right** at **Colvin Blvd** - go **0.7 mi**	**0.7 mi** 2 mins
	Turn **left** at **95th St** - go **0.4 mi**	**0.4 mi** 1 min
Arrive at Love Canal		
	Head **south** from **95th St** - go **0.2 mi**	**0.2 mi**
	Turn **right** at **Frontier Ave** - go **0.3 mi**	**0.3 mi** 1 min
	Bear **right** at **S Military Rd** - go **0.2 mi**	**0.2 mi**
	Turn **left** at **Cayuga Dr** - go **0.1 mi**	**0.1 mi**
	Turn **right** into the **LaSalle Expwy West** entry ramp - go **0.3 mi**	**0.3 mi**
	Merge into **LaSalle Expy W** - go **1.1 mi**	**1.1 mi** 1 min
	Take the **I-190 S** ramp - go **0.6 mi**	**0.6 mi** 1 min
	Take the **RT-384** exit **21** - go **0.2 mi**	**0.2 mi**
	Turn **right** at **Buffalo Ave** - go **0.6 mi**	**0.6 mi** 2 mins
Arrive at S-Area	**43.078085, -79.003627**	
	Head **west** from **Buffalo Ave** - go **0.8 mi**	**0.8 mi** 2 mins
	Turn **right** at **Hyde Park Blvd** - go **3.6 mi**	**3.6 mi** 6 mins
	Turn **right** at **Power Auth Service Dr** - go **0.1 mi**	**0.1 mi**

Arrive at OCC Hyde Park	**43.132229, -79.038333**	
	Continue along NYPA Service Drive	**0.5 mi**
Arrive at NPYA Access Road	**43.135535, -79.043717**	
	Return up Access road and turn left at Hyde Park Blvd (Rt 61).	0.4 mi
	Take Rt 61 to Rt 104,	0.2 mi
	Turn Right at Rt 104 go 0.4 mi	0.4 mi
Arrive at Power Vista	**43.141132, -79.040698**	
	Exit Parking lot, turn left at blinking light onto rt 104 West – go 2.0 mi	2.0 mi
	Turn Rt onto Findlay Drive – go 0.2 mi	0.2 mi
	Turn Rt onto Robert Moses Parkway – go 1.5 mi	1.5 mi
Arrive at Devil's Hole State Park	Arrive at Devil's Holes State Park	

D. Appendices

Appendix A1: Niagara Falls Hazardous Waste Sites

From (U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2005)

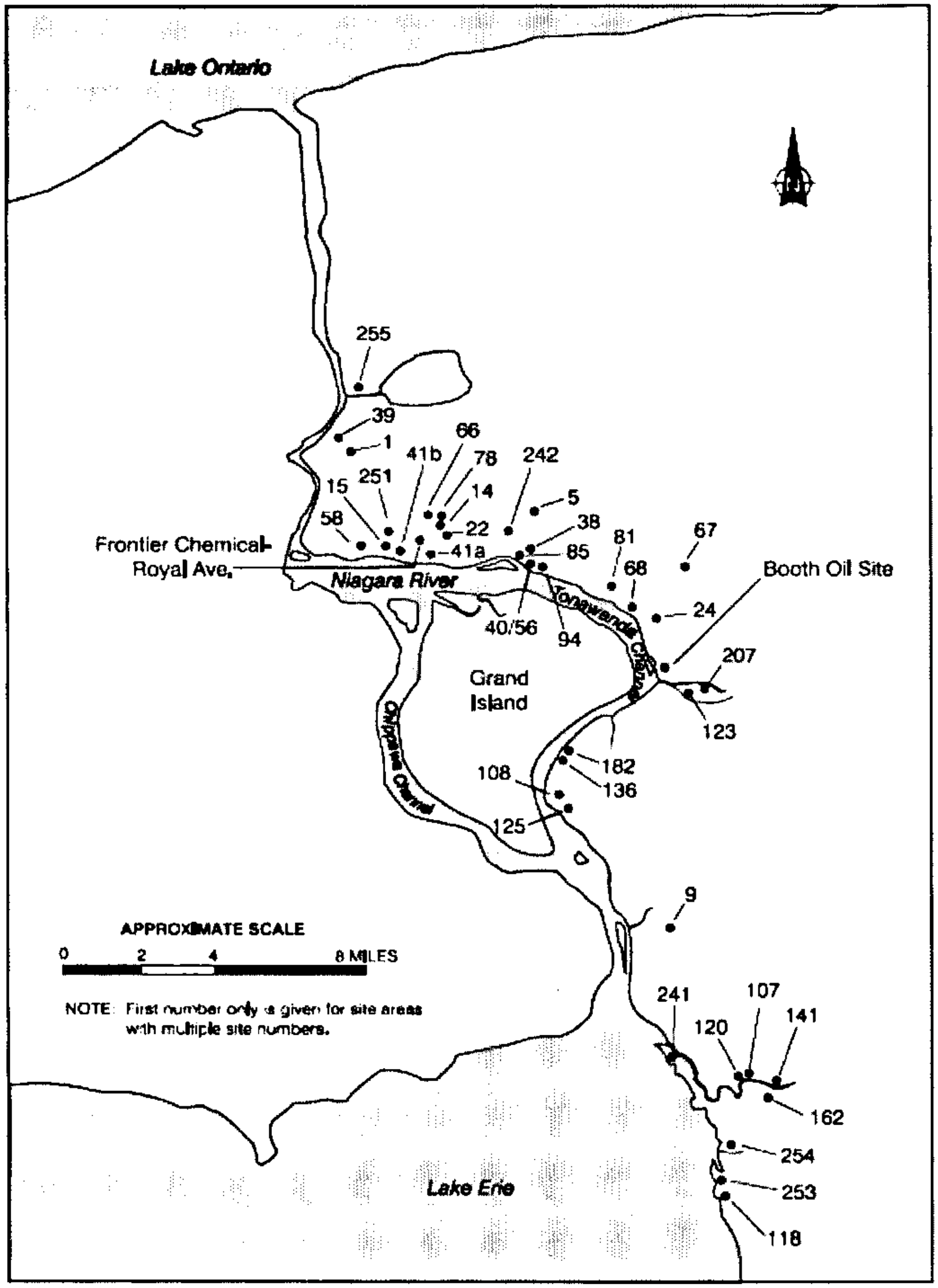


Figure 1: LOCATION OF SIGNIFICANT NIAGARA RIVER HAZARDOUS WASTE SITES

From (U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2005)

Figure 1: LEGEND

USGS SITE NUMBERS	SITE NAME
41b-49	Occidental Chemical Corp. (OCC), Buffalo Ave. Avenue
81	Niagara County Refuse Disposal
14	DuPont Necco Park
78a,b	CECOS International/Niagara Recycling
39	OCC, Hyde Park
40,56,85,94*	102nd Street
5	Bell Aerospace Textron
66	Durez Corporation, Packard Road Facility (formally OCC, Durez Division)
41a	OCC, S-Area
255	Stauffer Plant (PASNY)
251	Solvent Chemical
1	Vanadium Corp. (formerly SKW Alloys)
58,59,248	Olin, Buffalo Avenue
15-19,250	DuPont, Buffalo Avenue Plant
254	Buffalo Harbor Containment
120-122	Buffalo Color Corporation, including Area D
118	Bethlehem Steel Corporation
136	River Road (INS Equipment)
67	Frontier Chemical, Pendleton
24-37	OCC, Durez, North Tonawanda
253	Small Boat Harbor Containment
68	Gratwick Riverside Park
141	Mobil Oil
162	Alltift Realty
242	Charles Gibson
22	Great Lakes Carbon
182	Niagara Mohawk Cherry Farm
241	Times Beach Containment
108	Tonawanda Coke
107	Allied Chemical
207	Tonawanda Landfill
125-127	Dunlop Tire and Rubber
123	Columbus-McKinnon
38	Love Canal
9-15-141	Iroquois Gas/Westwood Pharmaceutical

* Occidental 102nd Street site (#40), Olin 102nd Street site (#56), Griffon Park (#85), and Niagara River Belden site (#94)

Appendix A2: Lockport Group Stratigraphy

Revised Stratigraphy of the Lockport Group according to Brett and others (Brett et al., 1995).

REVISED STRATIGRAPHY AND CORRELATIONS

<table>
<tr><th colspan="10">AUTHOR OR SOURCE</th></tr>
<tr><th colspan="3">BOLTON (1957)</th><th colspan="2">ZENGER (1965)</th><th colspan="2">RICKARD (1975)</th><th colspan="3">THIS REPORT</th></tr>
<tr><td rowspan="8">ALBEMARLE GROUP</td><td colspan="2" rowspan="2">GUELPH FORMATION</td><td rowspan="8">LOCKPORT FORMATION</td><td rowspan="3">OAK ORCHARD MEMBER</td><td rowspan="8">LOCKPORT GROUP</td><td rowspan="3">GUELPH DOLOMITE</td><td rowspan="8">LOCKPORT GROUP</td><td colspan="2">GUELPH DOLOMITE</td></tr>
<tr><td colspan="2" rowspan="2">ERAMOSA DOLOMITE</td></tr>
<tr><td rowspan="6">LOCKPORT FORMATION</td><td rowspan="2">ERAMOSA MEMBER</td></tr>
<tr><td>ERAMOSA MEMBER</td><td>ERAMOSA FORMATION</td><td rowspan="3">GOAT ISLAND DOLOMITE</td><td>VINEMOUNT MEMBER</td></tr>
<tr><td rowspan="2">GOAT ISLAND MEMBER</td><td rowspan="2">GOAT ISLAND MEMBER</td><td rowspan="2">GOAT ISLAND FORMATION</td><td>ANCASTER MEMBER</td></tr>
<tr><td>NIAGARA FALLS MEMBER</td></tr>
<tr><td rowspan="2">GASPORT MEMBER</td><td rowspan="2">GASPORT MEMBER</td><td rowspan="2">GASPORT FORMATION</td><td rowspan="2">GASPORT DOLOMITE</td><td>PEKIN MEMBER</td></tr>
<tr><td>GOTHIC HILL MEMBER</td></tr>
</table>

Figure 21. Historical summary of Lockport Group nomenclature in the Niagara region.

Appendix B1: Summary of Bell Aerospace Textron Site

(U.S. Environmental Protection Agency and New York State Department of Environmental Conservation, 2000)

Site Program: RCRA (State and Federal)
Summary Prepared by: EPA and DEC

Site Description

The Bell Aerospace Textron plant is located approximately 2.5 miles north of the Niagara River, adjacent to the Niagara Falls International Airport.

Between 1950 and 1980, the company used an unlined 60' X 100' surface impoundment to collect wash water from rocket engine test firings, storm run-off, and solvent drippings from cleaning, degreasing, and anodizing operations. Hazardous waste and constituents of concern include trichloroethylene and dichloroethylene. The wastes were discharged to a sanitary sewer after pH adjustment.

Beneath the site lies one overburden and two bedrock aquifers. Groundwater flow through the overburden aquifer is primarily to the south-southeast. There is a potential vertical flow between the overburden and the upper bedrock aquifer, and at least some of the groundwater from the overburden discharges to Bergholtz Creek. The upper bedrock aquifer flows primarily in a southeasterly direction and in the lower bedrock aquifer groundwater flow is generally to the south. The down-gradient extent of groundwater contamination in each of the three aquifers has been well defined, and, as of this update, no contaminated groundwater appears to be discharging directly to the Niagara River.

Remedial Actions

Bell Aerospace Textron is an RCRA site with a closed surface impoundment. The company excavated 1225 tons of contaminated soil and capped the area in 1987.

All of the remedial actions that were required here have been accomplished on schedule.

Since the initial 1989 hazardous waste site report, an RCRA Facility Investigation (RFI) has determined the extent of contaminant migration and a Corrective Measures Study (CMS) has addressed on- and off-site groundwater contamination. A State Part 373 post-closure permit was issued to Bell Aerospace in September 1992, which will expire in September 2003. The permit required final Corrective Measures Implementation (CMI), consisting of groundwater pump-and-treat programs for on- and off-site contamination. In addition, in October 2001 the facility has installed (on a voluntary basis) monitoring wells through the cap of the Neutralization Pond as part of an ongoing investigation of the natural degradation of groundwater contamination at the facility.

The overall remedial program is designed to intercept the bedrock groundwater that is migrating off-site toward the Niagara River. It consists of the installation of 11 groundwater extraction wells.

The off-site remedial system was started up in April 1993. It is achieving its designed objective. The capture zone associated with the system covers the area of groundwater contamination, and the areal extent of the contamination is diminishing. Five extraction wells have been installed to contain the off-site groundwater. However, as the off-site plume has become smaller, four extraction wells were determined to be optimal for pumping. The extracted groundwater contamination is discharged into the publicly owned treatment works (POTW) of the Town of Wheatfield. The off-site system is designed to recover two pounds of volatile compounds daily. The performance of the off-site remedial system is considered acceptable.

The on-site remedial system began the start-up operating period in April 1995. Several technical problems prevented the on-site system from attaining all of its design objectives. The remedial system was redesigned to address these problems, and the following two modifications were made:

- the installation of a 900 foot-long pipeline to divert the cooling water discharge from a rocket testing facility operating at the site to the storm drainage system; and,
- the installation of a slurry wall barrier along the main sewer line on Walmore Road to prevent the water migration from the sewer line to the on-site system.

However, even after these modifications, the on-site system was still not attaining satisfactory hydraulic containment. To address this, an additional extraction well was installed along the southern boundary of the site. This well was installed in July 1998, and is currently in operation. The operation of this well has increased the groundwater capture zone along the southern edge of the facility, but the capture zone was not consistently continuous from two of the five extraction wells. A higher capacity pump has been in operation on the new well since August 20, 1999, thus increasing the groundwater pumping rate.

With the above modifications, the on-site system is achieving its design goals. The on-site system has been effective in creating a groundwater capture zone over the DNAPL plume, therefore, all contaminated groundwater is being intercepted and treated on-site, so that no loading is migrating from the site. Six extraction wells are currently operating in the on-site system. The operation of the higher capacity pump has maintained a continuous capture zone. Monitoring data of 2002-2003 indicates a complete capture zone has been obtained along the southern boundary. The on-site system is designed to recover four pounds of volatile compounds daily.

Appendix B2a: Love Canal History

ecumenical task force
of the niagara frontier

Love Canal Collection

See: http://ublib.buffalo.edu/libraries/projects/lovecanal/

Background on the Love Canal

Introduction

During the summer of 1978, the Love Canal first came to international attention. On August 7, 1978, United States President Jimmy Carter declared a federal emergency at the Love Canal, a former chemical landfill which became a 15-acre neighborhood of the City of Niagara Falls, New York.

The Love Canal became the first man-made disaster to receive such a designation based on a variety of environmental and health related studies. As a result of grass roots interest and media attention, the Love Canal provided an impetus for dramatic interest in and changes to environmental concerns worldwide.

History of the Love Canal: 1892 -1978

The Love Canal, a neighborhood in the southeast LaSalle district of the City of Niagara Falls, New York, takes its name from the failed plan of nineteenth century entrepreneur, William T. Love. Approximately four miles upstream of Niagara Falls, Love saw an ideal location to harness water to generate power to the burgeoning industries developing along the seven mile stretch of the River to the mouth of Lake Ontario. In 1892, the canal was his solution to provide ships a route to bypass the Falls.

A few years later Love's dream of the navigable waterway evaporated. A nationwide economic depression, loss of financial backing, and the invention of alternating electrical current forced Love to abandoned his project. Only one mile of the canal had been dug.

U.S. Geological Aerial Photographs taken in 1927 clearly show an open body of water sixty feet wide and three thousand feet long at the otherwise undeveloped edge of the City. The Love Canal remained as a recreational area for swimming and boating well into the early 20th century.

By 1920, Love's land was sold at public auction and quickly became a municipal and chemical disposal site. From 1942 through 1953, the Love Canal Landfill was used principally by Hooker Chemical, one of the many chemical plants located along the

Niagara River. Nearly 21,000 tons (42 million pounds) of what would later be identified by independent scientists as "toxic chemicals" were dumped at the site.

In 1953, with the landfill at maximum capacity, Hooker filled the site with layers of dirt. As the post-war housing and baby boom spread to the southeast section of the City; the Niagara Falls Board of Education purchased the Love Canal land from Hooker Chemical for one dollar. Included in the deed transfer was a "warning" of the chemical wastes buried on the property and a disclaimer absolving Hooker of any further liability.

Single-family housing surrounded the Love Canal site. As the population grew, the 99th Street School was built directly on the former landfill. At the time, homeowners were not warned or provided information of potential hazards associated with locating close to the former landfill site.

According to residents who lived in the area, from the late 1950s through the early 1970s repeated complaints of odors and "substances" surfacing in their yards brought City officials to visit the neighborhood. The City assisted by covering the "substances" with dirt or clay, including those found on the playground at the 99th Street School. Faced with continuing complaints, the City, along with Niagara County hired Calspan Corporation as a consultant to investigate. A report was filed indicating presence of toxic chemical residue in the air and in the sump pumps of residents in living at the southern end of the canal. Also discovered were 50 gallons drums just below the surface of the canal cap and high levels of PCB's (polycholorinated biphenyls) in the storm sewer system. Remedial recommendations included covering the canal with a clay cap, sealing home sump pumps and a tile drainage system to control migration of wastes. No action was taken.

By 1978, the Love Canal neighborhood included approximately 800 private, single-family homes, 240 low-income apartments, and the 99th Street Elementary School - located near the center of the landfill. Two other schools, 93rd Street School and 95th Street School - were also considered to be part of this neighborhood comprised of working class families.

In April 1978, Michael Brown, a reporter for the Niagara Gazette newspaper, wrote a series of articles on hazardous waste problems in the Niagara Falls area, including the Love Canal dumpsite. In response to the articles, Love Canal residents once more began calling on City and County officials to investigate their complaints. By this time, many residents were beginning to question health risks and noting already existing inexplicable health problems.

At the same time, the New York State Department of Health (NYSDOH) began collecting air and soil tests in basements and conducting health studies of the 239 families immediately surrounding the canal. On April 25, 1978, the New York State Commissioner of Health, Dr. Robert Whalen issued a determination of public health hazard existing in the Love Canal Community. He ordered the Niagara County

Health Department to remove exposed chemicals from the site and install a protective fence around the area.

Once the report was public, Lois M. Gibbs, a resident and mother of two small children, canvassed the neighborhood to petition the closure of the 99th Street School where her son attended kindergarten.

Throughout the spring and summer of 1978, New York State Health Department, City of Niagara Falls and County of Niagara Falls officials, and Love Canal residents met to discuss the growing health hazard.

On August 2, 1978, the New York State Commissioner of Health, Robert M. Whalen, M.D. declared a medical State of Emergency at Love Canal and ordered the immediate closure of the 99th Street School. Immediate cleanup plans were initiated and recommendations to move were made for pregnant women and children under two who lived in the immediate surrounding area of the Love Canal.

The President of the United States Jimmy Carter declared the Love Canal area a federal emergency on August 7, 1978. This declaration would provide funds to permanently relocate the 239 families living in the first two rows of homes encircling the landfill. The remaining 10 block area of the Love Canal, including the home of Lois Gibbs, were not included in the declaration.

Appendix B2b: EPA Love Canal Fact Sheet

Love Canal

New York

EPA ID#: NYD000606947

EPA REGION 2

Congressional District(s): 29

Niagara

Niagara Falls

NPL LISTING HISTORY

Proposed Date: 10/1/1981

Final Date: 9/1/1983

Deletion Date: 9/30/2004

Site Description

The fenced 70-acre Love Canal site (Site) encompasses the original 16-acre hazardous waste landfill with a 40-acre clay/synthetic liner cap. Also, a barrier drainage system and leachate collection and treatment system is in place and operating. The Site includes the "original" canal that was excavated by Mr. William T. Love in the 1890's for a proposed hydroelectric power project but was never implemented. Beginning in 1942, the landfill was used by Hooker Chemicals and Plastics (now Occidental Chemical Corporation (OCC)) for the disposal of over 21,000 tons of various chemical wastes, including halogenated organics, pesticides, chlororbenzenes and dioxin. Dumping ceased in 1952, and, in 1953, the landfill was covered and deeded to the Niagara Falls Board of Education (NFBE). Subsequently, the area near the covered landfill was extensively developed, including the construction of an elementary school and numerous homes. Problems with odors and residues, first reported in the 1960's, increased during the 1970's, as the water table rose, bringing contaminated groundwater to the surface. Studies indicated that numerous toxic chemicals had migrated into the surrounding area directly adjacent to the original landfill disposal site. Runoff drained into the Niagara River, approximately three miles upstream of the intake tunnels for the Niagara Falls water treatment plant. Dioxin and other contaminants migrated from the landfill to the existing sewers, which had outfalls into nearby creeks. In 1978 and 1980, President Carter issued two environmental emergencies for the Love Canal area. As a result, approximately 950 families were evacuated from a 10-square-block area surrounding the landfill. The Federal Emergency Management Agency (FEMA) was directly involved in property purchase and residential relocation activities. In 1980, the neighborhoods adjacent to the Site were identified as the Emergency Declaration Area (EDA), which is approximately 350 acres and is divided into seven separate areas of concern. Approximately 10,000 people are located within one mile of the Site; 70,000 people live within three miles. The Love Canal area is served by a public water supply system; the City of Niagara Falls water treatment plant serves 77,000 people. The Site is 1/4 mile north of the Niagara River. The contamination problem discovered at the Site ultimately led to the passage of Federal legislation, governing abandoned hazardous waste sites

On December 21, 1995, a consent decree, as a cost recovery settlement between the United States and OCC, was lodged with the United States District Court. As part of the settlement, OCC and the United States Army have agreed to reimburse the Federal government's past response costs, related directly to response actions taken at the Site. The primary portion of OCC's reimbursement is $129 million; OCC has also agreed to reimburse certain other Federal costs, including oversight costs, and to make payments in satisfaction of natural resource damages claims. In a second part of this decree, the United States Army agreed to reimburse $8 million of the Federal government's past response costs; these funds have now been directed specifically into EPA Superfund and FEMA accounts.

Also, $3 million of the settlement funds will be directed to the Agency for Toxic Substances and Disease Registry (ATSDR) for the development of a comprehensive health study using the Love Canal Health Registry. ATSDR has awarded a grant to the New York State Department of Health (NYSDOH) to conduct this study which is the final stages of completion.

Site Responsibility: This Site is being addressed through Federal, State and potentially responsible party actions.

Threat and Contaminants

As a result of the landfill containment, the leachate collection and treatment system, the groundwater monitoring program and the removal of contaminated creek and sediments and other clean up efforts, the Site does not present a threat to human health and the environment.

Cleanup Approach

This Site has been addressed in seven stages: initial actions and six major long-term remedial action phases, focusing on 1) landfill containment with leachate collection, treatment and disposal; 2) excavation and interim storage of the sewer and creek sediments; 3) final treatment and disposal of the sewer and creek sediments and other Love Canal wastes; 4) remediation of the 93rd Street School soils; 5) EDA home maintenance and technical assistance by the Love Canal Area Revitalization Agency (LCARA), the agency implementing the Love Canal Land Use Master Plan; and, 6) buyout of homes and other properties in the EDA by LCARA.

Three other short-term remedial actions: a) the Frontier Avenue Sewer remediation, b) the EDA 4 soil removal, and c) the repair of a portion of the Love Canal cap, were completed in 1993 and are discussed below.

Response Action Status

Initial Actions: In 1978, New York State Department of Environmental Conservation (NYSDEC) installed a system to collect leachate from the Site. The landfill area was covered and fenced and a leachate treatment plant was constructed. In 1981, EPA erected a fence around Black Creek and conducted environmental studies.

Landfill Containment: In 1982, EPA selected a remedy to contain the landfill by constructing a barrier drain and a leachate collection system; covering the temporary clay cap with a synthetic material to prevent rain from coming into contact with the buried wastes; demolishing the contaminated houses adjacent to the landfill and nearby school; conducting studies to determine the best way to proceed with further site cleanup; and, monitoring to ensure the cleanup activities are effective. In 1985, NYSDEC installed the 40-acre cap and improved the leachate collection and treatment system, including the construction of a new leachate treatment facility.

Sewers, Creeks, and Berms: In May1985, as identified in a Record of Decision (ROD), EPA implemented a remedy to remediate the sewers and the creeks which included 1) hydraulically cleaning the sewers; 2) removal and disposal of the contaminated sediments; 3) inspecting the sewers for defects that could allow contaminants to migrate; 4) limiting access, dredging and hydraulically cleaning the Black Creek culverts; and, 5) removing and storing Black and Bergholtz creeks' contaminated sediments. [The remediation of the 102nd Street outfall area, as originally proposed in the 1985 ROD, has been addressed under the completed remedial action for the 102nd Street Landfill Superfund site.] The State cleaned 62,000 linear feet of storm and sanitary sewers in 1986. An additional 6,000 feet were cleaned in 1987. In 1989, Black and Bergholtz creeks were dredged of approximately 14,000 cubic yards of sediments. Clean riprap was placed in the creek beds, and the banks were replanted with grass. Prior to final disposal, the sewer and creek sediments and other wastes [33,500 cubic yards] were stored at OCC's Niagara Falls RCRA-permitted facilities.

Thermal Treatment of Sewers and Creeks Sediments: In October 1987, as identified in a second ROD, EPA selected a remedy to address the destruction and disposal of the dioxin-contaminated sediments from the sewers and creeks: 1) construction of an on-site facility to dewater and contain the sediments; 2) construction of a separate facility to treat the dewatered contaminants through high temperature thermal destruction; 3) thermal treatment of the residuals stored at the Site from the leachate treatment facility and other associated Love Canal waste materials; and, 4) on-site disposal of any non-hazardous residuals from the thermal treatment or incineration process. In 1989, OCC, the United States and the State of New York, entered into a partial consent decree (PCD) to address some of the required remedial actions, i.e ., the processing, bagging and storage of the creek sediments, as well as other Love Canal wastes, including the sewer sediments. Also, in 1989, EPA published an Explanation of Significant Differences (ESD), which provided for these sediments and other remedial wastes to be thermally treated at OCC's facilities rather than at the Site. In November 1996, a second ESD was issued to address a further modification of the 1987 ROD to include off-site EPA-approved thermal treatment and/or land disposal of the stored Love Canal waste materials. In December 1998, a third ESD was issued to announce a 10 ppb treatability variance for dioxin for the stored Love Canal waste materials. The sewer and creek sediments and other waste materials were subsequently shipped off-site for final disposal; this remedial action was deemed complete in March 2000.

93rd Street School: The 1988 ROD selected remedy for the 93rd Street School property included the excavation of approximately 7500 cubic yards of contaminated soil adjacent to the school followed by on-site solidification and stabilization. This remedy was re-evaluated as a result of concerns raised by the NFBE, regarding the future reuse of the property. An amendment to the original 1988 ROD was issued in May 1991; the subsequent selected remedy was excavation and off-site disposal of the contaminated soils. This remedial action was completed in September 1992. Subsequently, LCARA purchased the 93 rd Street School property from the NFBE and demolished the building in order to return the resulting vacant land to its best use.

Home Maintenance: As a result of the contamination at the Site, the Federal government and the State of New York purchased the affected properties in the EDA. LCARA is the coordinating New York State agency in charge of maintaining, rehabilitating and selling the affected properties. Pursuant to Section 312 of CERCLA, as amended, EPA provided funds to LCARA for the maintenance of those properties in the EDA and for the technical assistance during the rehabilitation of the EDA. EPA awarded these funds to LCARA directly through an EPA cooperative agreement for home maintenance and technical assistance. The rehabilitation and sale of these homes is complete. Since the rehabilitation program began, approximately 260 homes were sold. Also, a new senior citizen housing development has been constructed on vacant property in the habitable portion of the EDA. In 2000, EPA closed out this cooperative agreement with LCARA.

Property Acquisition: Section 312 of CERCLA, as amended, also provided $2.5M in EPA funds for the purchase of properties (businesses, rental properties, vacant lots, etc.) which were not eligible to be purchased under the earlier FEMA loan/grant. EPA awarded these funds to LCARA through a second EPA cooperative agreement. In 2000, EPA closed out this cooperative agreement with LCARA. LCARA was dissolved by NYS statute in August 2003.

Short-Term Remedial Actions: 1) The Frontier Avenue Sewer Project required excavation and disposal of contaminated pipe bedding and replacement with new pipe and bedding--excavated materials have been transported for off-site thermal treatment and/or land disposal. 2)The EDA 4 Project required the excavation and disposal of a hot spot of pesticide contaminated soils in the EDA and backfill with clean soils; excavated materials were disposed of off-site. 3) The Love Canal Cap Repair required the liner replacement and regrading of a portion of the cap. These short-term remedial actions were completed in September 1993.

Cleanup Progress

In 1988, EPA issued the Love Canal EDA Habitability Study (LCHS), a comprehensive sampling study of the EDA to evaluate the risk posed by the Site. Subsequent to the issuance of the final LCHS, NYSDOH issued a Decision on Habitability, based on the LCHS's findings. This Habitability Decision concluded that: 1) Areas 1-3 of the EDA are not suitable for habitation without remediation but may be used for commercial and/or industrial purposes and 2) Areas 4-7 of the EDA may be used for residential purposes, i.e., rehabitation.

In 1998, the wastewater discharge permit issued to OCC was modified to include the treatment of the leachate water from the 102nd Street Landfill site. In March 1999, the Love Canal leachate collection and treatment facility (LCTF) began receiving the 102 nd Street leachate water for treatment. The latest estimates represent the make up of the various Love Canal waste materials:

Sewer and Creek Sediment Wastes - 38,900 cubic yards @ 1.6 tons/cubic yard = 62,240 tons Collected LCTF DNAPL - 6000 pounds Collected 102nd Street DNAPL - 14,400 pounds Spent Carbon Filter Wastes - 40,380 pounds Treated LCTF Leachate - 4.35 MG Treated 102nd Street Landfill Treated Leachate - 0.58 MG

OCC is responsible for the continued operation and maintenance of the LCTF and groundwater monitoring. The Site is monitored on a continual basis through the numerous monitoring wells which are installed throughout the area. The yearly monitoring results show that the Site containment and the LCTF are operating as designed.

As shown above, numerous cleanup activities, including landfill containment, leachate collection and treatment and the removal and ultimate disposition of the contaminated sewer and creek sediments and other wastes, have been completed at the Site. These completed actions have eliminated the significant contamination exposure pathways at the Site, making the Site safe for nearby residents and the environment.

As a result of the revitalization efforts of LCARA, new homeowners have repopulated the habitable areas of the Love Canal EDA. More than 260 formerly-abandoned homes in the EDA were rehabilitated and sold to new residents, thus creating a viable new neighborhood. The vacant property in the EDA is currently being developed, according to the zoning and deed restrictions that are in place.

The Site was deemed construction complete on September 29, 1999. In September 2003, EPA issued a Five-Year Review Report that showed that the remedies implemented at the Site adequately control exposures of Site contaminants to human and environmental receptors to the extent necessary for the protection of human health and the environment. The next Five-Year review is scheduled for September 2008.

The Site was deleted from the National Priorities List on September 30, 2004.

Site Repositories

EPA Western New York Public Information Office @ (716) 551-4410, 186 Exchange Street, Buffalo, New York 14204.

Appendix B3: OCC S-Area EPA Fact Sheet

HOOKER CHEMICAL S-AREA
NEW YORK
EPA ID# NYD980651087
EPA REGION 2
CONGRESSIONAL DIST. 29
NIAGARA COUNTY
ALONG THE NIAGARA RIVER

Site Description
The Hooker Chemical SArea site is an 8acre industrial landfill owned by the Occidental Chemical Corporation. It is located at the southeast corner of OCC's Buffalo Avenue chemical plant in Niagara Falls, New York, along the Niagara River. Adjacent to the landfill is the City of Niagara Falls (City) drinking water treatment plant (DWTP). The Province of Ontario, Canada, is located across the Niagara River, a distance of approximately two miles. The landfill lies atop approximately 30 feet of soil, clay, till, and manmade fill on an area reclaimed from the Niagara River. Beneath these materials is fractured bedrock. OCC disposed of approximately 63,000 tons of chemical processing wastes into the landfill from 1947 to 1961. The landfill also was used by OCC for disposal of other wastes and debris, a practice that ended in 1975. Two lagoons for nonhazardous waste from plant operations were located on top of the landfill and were operated under New York State permits until 1989, when OCC discontinued operating these lagoons. During an inspection of the DWTP in 1969, chemicals were found in the bedrock water intake structures. In 1978, sampling of the structures and bedrock water intake tunnel revealed chemical contamination. The site is located in a heavily industrialized area of Niagara Falls. There is a residential community of approximately 700 people within 1/4 mile northeast of the site. The DWTP serves an estimated 70,000 people.

Site Responsibility:
This site is being addressed through Federal and potentially responsible parties' actions.

NPL LISTING HISTORY
Proposed Date: 12/01/82
Final Date: 09/01/83

Threats and Contaminants
On and offsite ground water and soil are contaminated with toxic chemicals occurring as both aqueous (water soluble) phase liquids (APLs) and nonaqueous (immiscible) phase liquids (NAPLs). These chemicals include primarily chlorinated benzenes. Dioxin is also present in ground water at trace levels. The main health threat to people is the risk from eating fish from the lower Niagara River/Lake Ontario Basin. Consumption of drinking water from the City's DWTP is not presenting health risks at present. However, the site,

because of its proximity to the DWTP, presents a potential public health threat to the consumers of drinking water from the plant.

Cleanup Approach

The site is being addressed in three phases: immediate actions and two longterm remedial phases focusing on cleanup of the entire site and construction of a municipal drinking water treatment plant.

Response Action Status

Immediate Actions: The City closed the contaminated main intake tunnel at the DWTP and put an emergency tunnel into service to alleviate the threat of contaminating drinking water.

Entire Site: EPA selected a containment remedy to prevent further chemical migration from the landfill toward the DWTP and into and under the Niagara River. The remedy includes: (1) a slurry cut-off wall (barrier wall) to encompass the landfill and offsite areas contaminated with chemicals in overburden soils, (2) an overburden collection system located within the barrier wall and comprised of horizontal drains and groundwater extraction wells to contain and collect both APL and NAPL chemicals, (3) a bedrock remedial system consisting of groundwater extraction wells and NAPL recovery wells; (4) an onsite leachate storage facility for separating and storing APL and NAPL chemicals prior to treatment; (5) a carbon adsorption facility for treating APL chemicals; (6) incineration of NAPL chemicals; (7) a final cap; and (8) monitoring programs to determine the effectiveness of the remedy. All components of the remedy selected for the landfill, with the exception of the final cap and monitoring programs, have been constructed. Operational startup of the remedial systems began in 1996. An evaluation of the remedial systems performances is ongoing.

The evaluation of the overburden drain collection system revealed that it was not operating or functioning as designed. Upon further inspection, the horizontal drain pipe was found to be crushed at several locations. The damaged drain collection system was replaced in 1999. The final cap, once scheduled for completion in 1999, will be installed in the year 2000.

City of Niagara Falls Drinking Water Treatment Plant: The remedy selected to address contamination at the DWTP includes the construction of a new plant at a new location and demolition and cleanup of the old plant property. The new plant was built and on-line by the end of March 1997. Demolition of the old plant was completed in late 1997. The remedy selected for the old plant property includes (1) a slurry cut-off wall (an extension of the S-Area barrier wall) to contain NAPL and APL chemicals, (2) a drain collection system to prevent APL chemicals in overburden soils from migrating to the Niagara River, (3) grouting of the old bedrock raw-water intake tunnel, and (4) capping. Engineering designs were completed in 1997. The slurry cut-off wall, drain collection system and cap were constructed in 1998. The tunnel grouting project is scheduled for 2000.

Site Facts: In 1979, the U.S. Department of Justice, acting on behalf of the EPA, filed a complaint against the parties potentially responsible for the site contamination. The State of New York joined in the suit and a Settlement Agreement was signed by the parties in

January 1984. It was approved and entered by the District Court of Western New York in April 1985. The Agreement called for a potentially responsible party to conduct an investigation at the site, to recommend cleanup standards for the site, and to conduct site cleanup activities. A second agreement was signed by the parties in September 1990 and approved by the Court in April 1991. This Agreement, which amended the original 1985 Settlement Agreement, included an expanded cleanup program to address offsite areas and the construction of a new DWTP.

Cleanup Progress
The construction of a new $70 million DWTP at a new location addresses the threat to the drinking water supply from S-Area. The new plant replaces the old facility, which supplied drinking water to city residents for the past 83 years. The S-Area barrier wall and remedial systems provide physical and hydraulic containment of the 63,000 tons of chemical waste buried in the landfill. Their operations have also reduced the loadings of toxic chemicals to the Niagara River. Approximately 320,000 gallons of contaminated ground water are treated per day, with the treated effluent discharged to the Niagara River via a permitted outfall. Since the startup of the S-Area remedial systems in 1996, approximately 350 million gallons of contaminated ground water have been treated. Approximately 65,000 gallons of NAPL have been collected for incineration.

Site Repository
USEPA Public Information Office, Carborundum Center, Suite 530, 345 Third Street, Niagara Falls, New York, 14303

Appendix B4: Hooker Hyde Park EPA Fact Sheet

Hooker - Hyde Park

New York

EPA ID#: NYD000831644

EPA REGION 2

Congressional District(s): 29

Niagara

Northwest of the City of Niagara Falls

NPL LISTING HISTORY

Proposed Date: 12/1/1982

Final Date: 9/1/1983

Site Description

Hooker-Hyde Park is a 15-acre site that was used to dispose of approximately 80,000 tons of waste, some of it hazardous material, from 1953 to 1975. The landfill is immediately surrounded by several industrial facilities and property owned by the New York Power Authority. The Niagara River, which flows into Lake Ontario, is located 2,000 feet northwest of the site. Bloody Run Creek, the drainage basin for the landfill area, flows from the northwestern corner of the landfill. The creek eventually flows into storm sewers and down the Niagara Gorge Face into the Niagara River. The site is located a few blocks east of a 500-home residential community. Approximately 3,000 people are employed by the industries near the site. All of the industries and most of the residences are connected to a municipal water supply system.

Site Responsibility This site has been addressed through Federal and potentially responsible parties' actions.

Threat and Contaminants

The ground water is contaminated with volatile organic compounds (VOCs) and dioxin from former disposal activities. Bloody Run Creek sediments were contaminated with VOCs until their removal in 1993 and surface water of the Niagara Gorge Face was contaminated with VOCs. Potential health threats include the consumption of contaminated fish from Lake Ontario. Although groundwater is contaminated, there are no known uses of groundwater within the area, so it is unlikely that people would be exposed to groundwater contaminants. Access to the landfill is restricted by a fence and a 24-hour guard.

Cleanup Approach

The site is being addressed in a single long-term remedial phase focusing on cleanup of the entire site.

Response Action Status

Entire Site: Remedial Construction has been completed at this site.

In 1985, EPA selected cleanup remedies which include the following: (1) a source control extraction well system to remove non-aqueous phase liquids (NAPL) from the overburden in the landfill; (2) an overburden drain system surrounding the landfill; (3) a bedrock remedial system to prevent the migration of leachates comprised of (a) a NAPL plume containment system and (b) an aqueous phase liquid (APL or contaminated leachate) plume containment system; (4) a shallow and deep groundwater study; (5) a Niagara Gorge seep program; and, (6) the treatment of leachates. The potentially responsible party, Occidental Chemical Corporation (OCC), has implemented these remedies since 1985. To date, OCC has completed the following remedies. Two source control wells were pump tested in 1993 and are operating. Four additional source control wells were installed in 1994 and are also operating. The Overburden Barrier Collection System, a drain surrounding the landfill to collect and contain leachate, was completed in 1990. This drain system prevents leachate from migrating outwardly through the overburden from the landfill. The bedrock NAPL containment system is a system of extraction wells that will recover NAPL and APL from the bedrock. These wells are placed in three discrete bedrock zones. Pumping these wells will create an inward hydraulic gradient (ground-water flow) towards the landfill which will prevent the outward migration of leachate in the bedrock, while collecting the leachate for treatment. The bedrock NAPL containment system is being installed in phases since not enough is known of the hydrogeology in fractured bedrock to design a final system. Phase I wells were completed in 1993 and are operating. Phase II wells were completed in late 1993 and are operating. Three additional extraction wells (Phase III) were installed in 1997. Two wells were installed in 1998 and connected via a force main to the on-site treatment facility. OCC installed two new extraction wells and the associated monitoring wells during 1999. Currently, the bedrock NAPL containment system consists of a total of twelve extraction wells operating around the site. The APL plume containment system consists of two extraction wells placed near the Niagara Gorge that recover APL and prevent it from reaching the Niagara River. These wells were completed in 1994. The construction of the on-site leachate storage, handling, and treatment facility was completed in 1989. APL is treated on-site with activated carbon. NAPL is collected at this facility and transferred to OCC's Main Plant in Niagara Falls for incineration. The Niagara Gorge Face seeps have been remediated. Contaminated sediment was removed and some water diverted into a culvert so that people no longer have access to these seeps. In addition to these remedial measures, an Industrial Protection Program to protect nearby workers from contaminants has been completed. The draft Lake Ontario Dioxin Bioaccumulation Study was completed in 1989, distributed for scientific review and was available to the public in September 1992. Fish and sediment samples from Lake Ontario were collected and analyzed, and laboratory studies were conducted. The community monitoring program, consisting of monitoring wells placed within the community and sampled quarterly to provide early warning of contamination from Hyde Park indicator chemicals, is ongoing. An assessment was completed in March 1992 to determine the risk of excavating Bloody Run sediments. The risks from excavation, EPA's preferred alternative, were found acceptable and the decision made to excavate the Bloody Run. Excavation was completed in February 1993. The perimeter of the landfill was capped in 1992. The landfill itself was capped in late 1994.

OCC installed 5 additional extraction wells in 2001 because the monitoring system indicated that there was not 100% capture of the contaminated groundwater. OCC upgraded its onsite treatment facility to process 400 gallons per minute in 2002. Even though OCC was effectively dewatering the aquifer, they could not demonstrate complete capture. OCC proposed a new site conceptual model in which there are 11 flow zones at the site and not just 3 aquifers. OCC conducted an extensive geophysical sampling program at the site in 2001 in order to better characterize the ground-water flow zones.

OCC, using an extensive monitoring system which was installed at the site during 2001 and 2002, concluded in the Remedial Characterization Report: Hydrologic Characterization (June 2003) that the contaminated groundwater surrounding the site was being captured by the extraction well system and that the requirements of the RRT were being achieved. OCC conducted a study to determine the relative age of the water near the site and determined that the relative age of the groundwater between the extraction wells and the Niagara Gorge is younger than the groundwater underlying the site. This indicates that the extraction wells are effectively preventing migration of groundwater from the landfill to the Niagara River. The seeps along the gorge were determined not to be groundwater discharge, but surface runoff, indicating that the APL wells have been effective at controlling the groundwater near the gorge.

Site Facts: In 1981, the EPA, the Department of Justice, the State, and a potentially responsible party, Occidental Chemical Corporation, signed a Consent Decree specifying OCC's responsibilities for cleanup of contamination at the site and maintenance of these remedies. In 1985, the EPA selected the final method to clean up the site. There is intense public scrutiny of activities related to this site. Two citizens' groups have intervened in the lawsuit against the potentially responsible party. The Canadian government also reviewed all of the program activities.

Cleanup Progress

The cleanup actions at the Hooker-Hyde Park site were completed in September 2003. The removal of contaminated soils and sediments and the leachate control and treatment operations have substantially reduced potential health risks and further environmental degradation.

Remedial construction included the installation of a system of extractions wells, both in the bedrock and overburden, to contain and collect NAPL & APL. A Leachate Treatment Facility was built on-site. Contaminated sediments were removed from Bloody Run.

Approximately 5 million gallons of ground water have been treated on-site; approximately 350,000 gallons of NAPL have been extracted from the site and incinerated; 46,720 tons of contaminated sediments were removed from Bloody Run.

Future Activities:

• Operation and maintenance of the ground-water extraction and treatment systems. Approximately 250 million gallons of groundwater will need to be treated over the next 30 years;

• NAPL is currently incinerated offsite at a facility in Texas.

Site Repositories

US EPA Western NY Public Information Office 186 Exchange Street Buffalo, New York 14204 716.551.4410

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THE UPPER DEVONIAN RHINESTREET BLACK SHALE OF WESTERN NEW YORK STATE – EVOLUTION OF A HYDROCARBON SYSTEM

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INTRODUCTION

Organic-rich, low permeability, sometimes fractured, shale deposits, once ignored by drillers seeking more readily understood plays and faster returns on their investments, are boosting the fortunes of midsized producers across the United States and becoming an increasingly important resource base. This burgeoning interest is driven largely by increased natural gas prices, improved completion technologies and a lack of conventional reservoir targets. However, there is no universal model applicable to each and every unconventional or continuous-type reservoir. Indeed, most vary in terms of basic stratigraphic facies distribution, mineralogy (i.e., quartz content and clay type and content), fracture parameters (length, orthogonal spacing, connectivity, anisotropy), porosity and permeability, and rock mechanical properties. There clearly is a need to further our understanding of these economically important deposits. This fieldtrip focuses on the evolution of the heavily fractured Upper Devonian Rhinestreet black shale, which has served not only as a hydrocarbon source rock, but also as its own seal and reservoir, the essence of the unconventional reservoir. Specifically, we address the Rhinestreet shale in terms of the hydrocarbon system, which comprises three basic elements (Demaison and Huizinga, 1994). The first component, *charge*, is the hydrocarbon volume available for entrapment and is dependent upon source rock richness and volume. *Migration*, the second parameter, can be thought of in terms of the lateral movement of hydrocarbons via permeable carrier beds versus vertical migration via faults and, more to the point of the Rhinestreet shale, fractures. The final component of the petroleum system is *entrapment*, which involves consideration of those rock properties that would favor the arrest of migrating hydrocarbons.

We will consider the Rhinestreet shale from deposition through early diagenesis, including mechanical compaction, followed by the early and relatively shallow generation of abnormally high pore fluid pressures caused by a marked increase in sedimentation rate at the end of the Devonian. Soon after this, at the onset of the Alleghanian orogeny, the Appalachian Basin of western New York state and western Pennsylvania was uplifted resulting in the initial phase of fracturing in the Rhinestreet shale. This was followed by rapid burial of the organic-rich rocks to the oil window and the generation of natural hydraulic fractures (joints) in response to catagenesis. Indeed, the jointing history of these rocks provides a record of the complex overpressure history of the Rhinestreet shale and other black shale units of the western New York state region of the basin.

Results reported herein represent the record of six years of work on the Upper Devonian shale-dominated succession exposed along the Lake Erie shoreline and general vicinity, the Lake Erie District (LED). This work is now being extended into the subsurface of western New York state, northwest

Pennsylvania, and northeast Ohio by use of wireline logs. During his involvement in the study of the Rhinestreet shale, DRB was supported by two SUNY Fredonia Summer Undergraduate Research Fellowships. Most recently, his work, which constitutes the better part of his MS thesis at the University of Buffalo, was supported by an AAPG grant-in-aid. We acknowledge Peter Bush and his staff at the University of Buffalo, South Campus Instrumentation Center, School of Dental Medicine, for their unflagging help with the electron microscopy, Ross Giese and Anja Dosen, University of Buffalo, for the XRD analyses, and Taury Smith, Rich Nyahay, and James Sandor of the Reservoir Characterization Group at The New York State Museum, Albany, New York. Finally, we thank Terry Engelder for many fruitful discussions on compaction, jointing, the geology of the Catskill Delta Complex, and more in the field, over the phone, and via E-mail.

GENERAL STRATIGRAPHY OF THE RHINESTREET SHALE

The Upper Devonian (Frasnian) Rhinestreet shale, West Falls Group, is best exposed along Eighteenmile Creek in northern Erie County, western New York state, where it comprises ~54 m of heavily jointed, finely laminated organic-rich grayish-black to medium-dark-gray (herein referred to as black) shale (~ 85% of the unit in this area), thin (2-3 m-thick) intervals of greenish-gray, dark-greenish-gray, and medium gray shale, thin black shale beds, discontinuous concretionary carbonate layers, the so-called "scraggy layers" of Luther (1903), several horizons of ellipsoidal carbonate concretions, and rare thin siltstone beds (Fig. 1). The Rhinestreet is underlain by the Cashaqua shale, the contact being abrupt and easily recognized in the field. The upper 2.75 m of the Cashaqua shale is composed of brownish-black to olive-gray finely laminated shale. Below this, the Cashaqua comprises dark-greenish-gray shale and discontinuous diagenetic carbonate horizons (Fig. 1). The top of the Rhinestreet is transitional into the overlying Angola gray shale and marked by an interval of interbedded black and gray shale (Fig. 1). We follow Luther (1903) and Pepper et al. (1956) in placing the upper contact of the Rhinestreet shale at the base of the upper scraggy layer (Fig. 1), a horizon readily recognized in the field and even on wireline logs as discussed below.

The Rhinestreet shale was named for exposures of black shale near Rhinestreet Road at Naples, Ontario County, New York (Clarke, 1903). However, the Rhinestreet shale at its type locality is markedly different from that of Erie County where black shale comprises more than 80% of the unit. Indeed, black shale is found only in the lower part of the Rhinestreet shale in Ontario County (Pepper et al., 1956). Recently, Lash (2006a) demonstrated a rapid eastward reduction in the thickness of the black shale interval (based on gamma-ray and bulk density wireline logs) of the Rhinestreet shale in the subsurface in eastern Allegany County, New York (Fig. 2). It is tempting to attribute this abrupt facies change to penecontemporaneous movement on the Clarendon-Linden Fault System, the trace of which trends ~ north-south in eastern Cattaraugus County (CLF; Chadwick, 1920; Jacobi, 2002) (Fig. 2). If this fault were active during Rhinestreet sedimentation, and relative offset was down-to-the-east, then the rapid eastward reduction in organic rich-sediment at the expense of organic-lean shale and siltstone (e.g., Pepper et al., 1956) may reflect the existence of a hinge that ponded more clastic-rich deposits to the east (hinge position 2, Fig. 3). Such a scenario accords with the interpretation of Jacobi and Fountain (1993) who suggested that during accumulation of the Genesee Group, the axis or hinge of the foreland basin lay to the east of the CLF (hinge position 1, Fig. 3) whereas by Dunkirk shale time, the basin axis had migrated to the west of the CLF (hinge position 3, Fig. 3). This intriguing question awaits further study of the Rhinestreet shale in the subsurface.

The Rhinestreet shale is one of several black shale units that comprise part of the Upper Devonian succession of the western New York state region of the Appalachian Basin (Fig. 3). Each organic-rich unit occupies the basal interval of an unconformity-bound sequence and is conformably overlain by organic-lean gray shale and siltstone (Fig. 3). Accumulation of black shale records the episodic cratonward advance of the Acadian fold and thrust load accompanied by rapid subsidence of the basin and deposition of clastic-starved, carbonaceous sediment (Ettensohn, 1985, 1987, 1992). Overlying organic-lean shale and siltstone reflects tectonic relaxation, establishment of terrestrial drainage systems and delta progradation (Ettensohn, 1985, 1992). Nevertheless, the correlation of some black shales with global marine transgressions suggests that eustatic oscillations and/or fluctuations, too, may have played a role in the accumulation of Upper Devonian black shale in the Appalachian Basin (Johnson et al., 1985; Algeo et al., 1995; Brett and Baird, 1996; Werne et al., 2002). Indeed, the two gray shale intervals within the Rhinestreet shale (Fig. 1) that reflect sedimentation in somewhat more oxygenated, and perhaps shallower, water than the black shale, are probably elements of fourth-order progradational/retrogradational marine cycles of glacio-eustatic origin like those recognized in late Frasnian deposits throughout much of the Appalachian Basin (Filer, 2002).

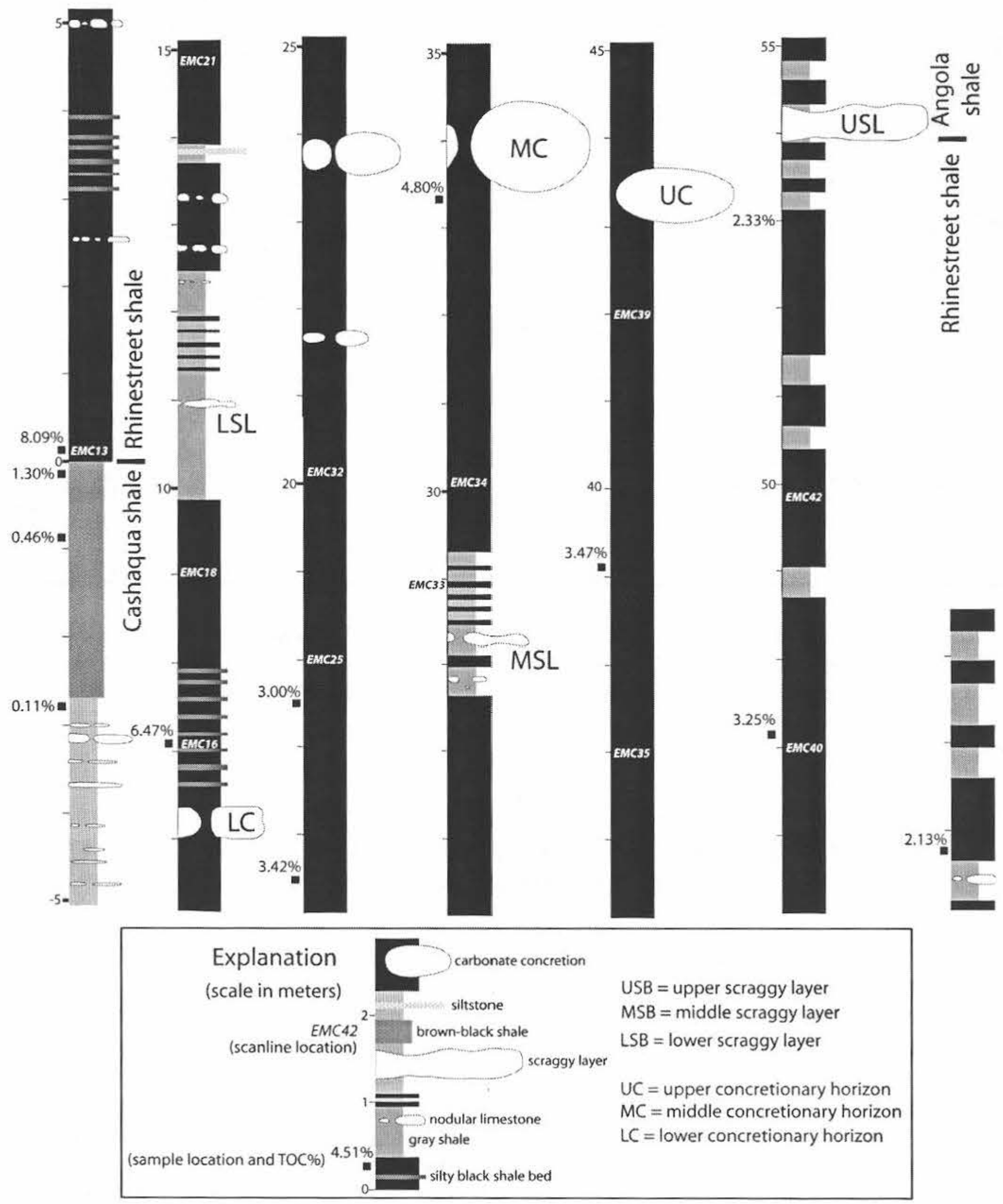


Fig. 1: Stratigraphic log of the Rhinestreet shale, Eighteenmile Creek.

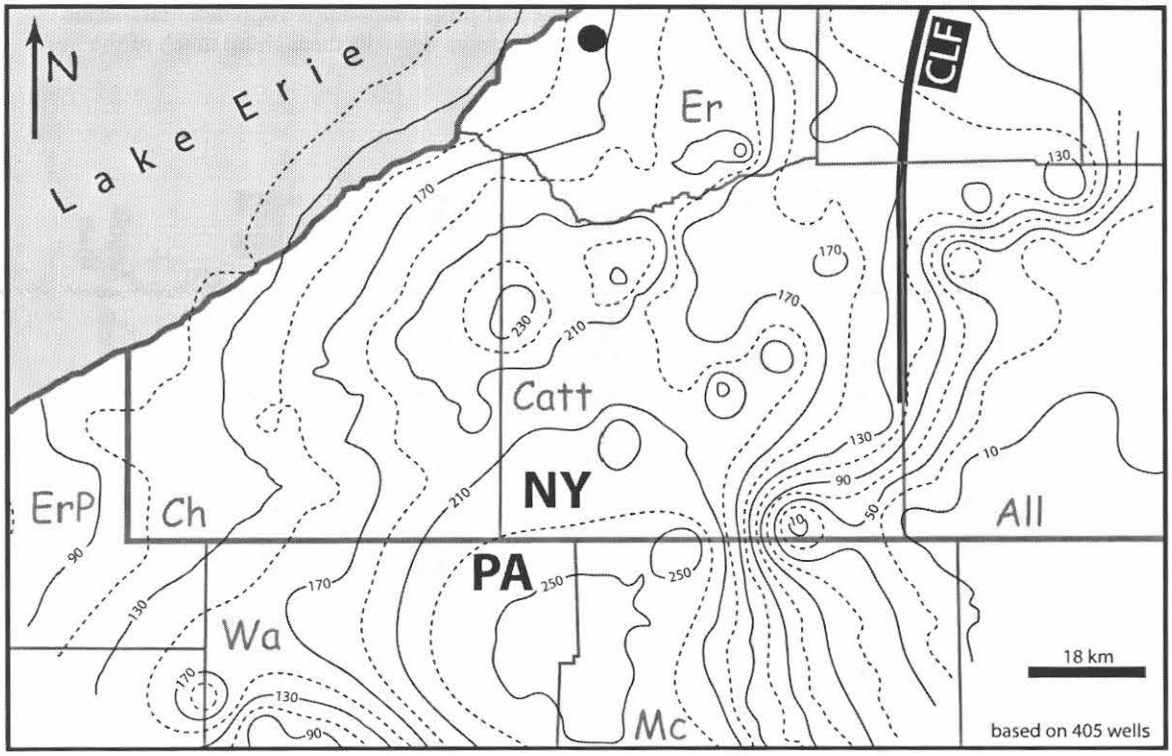


Fig. 2: Isopach map of the Rhinestreet shale in western New York state and northwest Pennsylvania showing the trace of the Clarendon-Linden Fault System (CLF). The thickness of the Rhinestreet shale is based on estimated TOC (Lash, 2006a). Contours in feet; filled circle marks the location of the Eighteenmile Creek measured section. New York state counties: Ch = Chautauqua; Catt = Cattaraugus; All = Allegany; Er = Erie. Pennsylvania counties: ErP = Erie; Wa = Warren; Mc = McKean.

RESERVOIR AND SOURCE ROCK CHARACTERISTICS AND BURIAL/THERMAL HISTORY

There is little doubt that the Rhinestreet shale of the Chautauqua-Cattaraugus-Erie counties region of western New York state and to the south into Pennsylvania has the makings of a quality source and reservoir rock. X-ray diffraction (XRD) analysis of three samples each of the Rhinestreet and Casahqua shales reveals noteworthy mineralogic differences. The Rhinestreet samples contain more quartz (Fig. 4), including non-detrital (i.e., lack of sharp edges) grains recognized in thin section and scanning electron microscopic (SEM) analysis. Clay minerals comprise a greater proportion of the Cashaqua samples than they do the Rhinestreet samples (Fig. 4). Illite is by far the dominant clay mineral in samples of both units, followed by chlorite and kaolinite. However, based on other thermal indicators, including vitrinite reflectance and Rock-Eval parameters, we suggest that the bulk of the illite is more a reflection of a low-grade metamorphic source terrane than a high level of thermal stress sustained by the Rhinestreet and Cashaqua shale samples (e.g., Lash, 2006b). Chlorite is consistently more abundant than kaolinite in Cashaqua shale samples; the reverse is the case for the Rhinestreet samples. Feldspar (microcline and plagioclase) averages 5.8% (±0.6%) in the Rhinestreet samples and 2.9% (±0.3%) in analyzed Cashaqua shale samples. Calcite, dolomite, and siderite, in decreasing order of abundance, are present in variable amounts in the Rhinestreet (0.9 – 2.2%) and Cashaqua (1.9 – 5.2%) shales. Indeed, gray shale of the LED, including the Cashaqua and Angola shales, typically contain more calcite in the form of cement

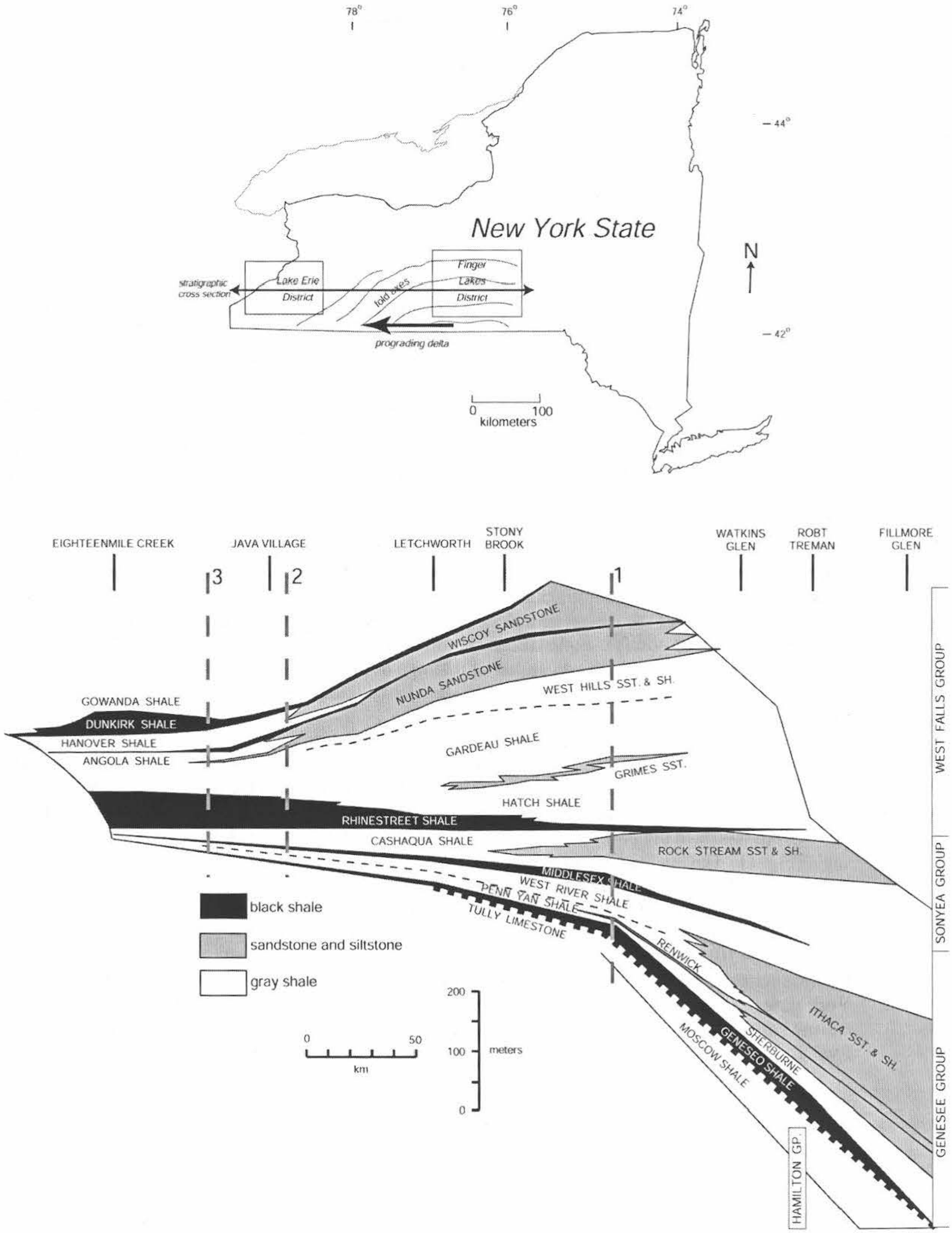


Fig. 3: Location map of the Lake Erie and Finger Lakes districts and a generalized east-west stratigraphic cross-section through the Catskill Delta Complex (modified after Woodrow et al., 1988) showing inferred positions of basin hinges.

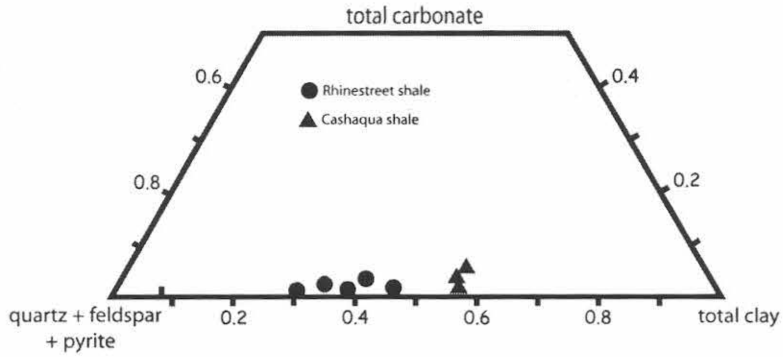


Fig. 4: Compositional plots for Rhinestreet and Cashaqua shale samples examined in this study.

(occasionally > 25%) than organic-rich shale, perhaps a reflection of the higher pre-lithification porosity of the former (Lash, 2006b).

Reservoir characteristics of the Rhinestreet and Cashaqua shales were investigated by mercury injection capillary pressure (MICP) measurements made on those samples analyzed by XRD. Samples were prepared so that pressure measurements were made perpendicular to bedding. Capillary entry measurements (10% saturation levels) for both sets of samples are quite high; however, the Rhinestreet entry pressures (~10,007 psia) are more than five times those of the Cashaqua shale samples (~1,800 psia) (Fig. 5). The calculated porosity of the Rhinestreet shale is 3.9% (±0.9%), less than one-half that of the Cashaqua shale (8.5% ±1.2%). MICP measurements indicate that the Rhinestreet shale is two orders of magnitude less permeable than the Cashaqua shale (10^{-20} m^2 vs. 10^{-18} m^2, respectively), which seems at odds with the generally coarser grained nature of the former. This likely reflects the very small average pore throat diameter of the Rhinestreet shale (8 nm) as opposed to that of the Cashaqua shale samples (19.2 nm). There are three possible explanations for this: (1) the strongly oriented planar microfabric of the black shale (see below), (2) the generation of bitumen during catagenesis of the Rhinestreet shale that clogged pore throats; and (3) the squeezing of abundant ductile organic matter in the Rhinestreet shale into void spaces (e.g., Lash, 2006b).

Total organic carbon (TOC) content of the Rhinestreet shale at its base along the Eighteenmile Creek section is 8.09%, diminishing irregularly upward to 2.3 % at its contact with the Angola shale (Fig. 1). TOC in the Cashaqua shale typically is less than 0.5%; however, TOC in the 2.75 m-thick brownish-black laminated shale interval at the top of the Cashaqua shale immediately beneath the Rhinestreet shale varies from 0.46% at the bottom of this interval to 1.3% at its top, 13 cm below the base of the Rhinestreet shale (Fig. 1). TOC in the Cashaqua gray-green shale 17 cm below its contact with the brownish-black interval is 0.11%, a value more typical of the Cashaqua shale (Fig. 1). Organic matter in the Rhinestreet shale consists principally of unstructured lipids (proprietary data); however, algal remains have been recognized in thin section and SEM analysis. The organic matter of the Rhinestreet is a combination of Type II/Type III oil and gas prone material (Fig. 6). The measured vitrinite reflectance, $\%R_o$, of Rhinestreet shale samples collected from both field exposures and gas well cuttings from western New York state range from 0.52% to 0.77% (Fig. 7). The bulk of the samples fall within the oil window for marine kerogen ($\%R_o > 0.6\%$; Tissot and Welte, 1984; Espitalie, 1986), as suggested by the plot of Rock-Eval T_{max} vs. hydrogen index (HI) as well as calculated production index values (Fig. 8).

The EASY$\%R_o$ kinetic model of vitrinite reflectance (Sweeney and Burnham, 1990) was used to model the burial/thermal history of the Rhinestreet black shale. This algorithm requires knowledge of (1) the age(s) of the unit(s) of interest (the base of the Rhinestreet shale), (2) at least a partial thickness of the local stratigraphic sequence, and (3) the measured vitrinite reflectance of the unit(s) of interest (average vitrinite reflectance of the base of the Rhinestreet shale = 0.74%). We estimate that the Cashaqua shale in the western New York state area was overlain by as much as 1,155 m of Devonian strata (Lindberg, 1985) and that the age of the base of the Rhinestreet shale can be dated by the Belpre ash bed at ~ 381 Ma (Kaufmann, 2006). Finally, our model assumes a geothermal gradient of 30° C km^{-1} and a 20° C seabed temperature (e.g., Gerlach and Cercone, 1993). We will make reference to the burial/thermal model of the Rhinestreet shale (Fig. 9) throughout the text. However, in light of the fact that vitrinite reflectance suppression has been recognized in Devonian black shales (e.g., Rimmer and Cantrell, 1988; Rimmer et al., 1993; Nuccio and Hatch, 1996; Rowan et al., 2004), including those from northwest Pennsylvania and western New York state (see Stop 2 discussion), we caution that the estimated 3.1 km maximum burial depth of the Rhinestreet shale (Fig. 9) should be viewed as a minimum maximum burial depth.

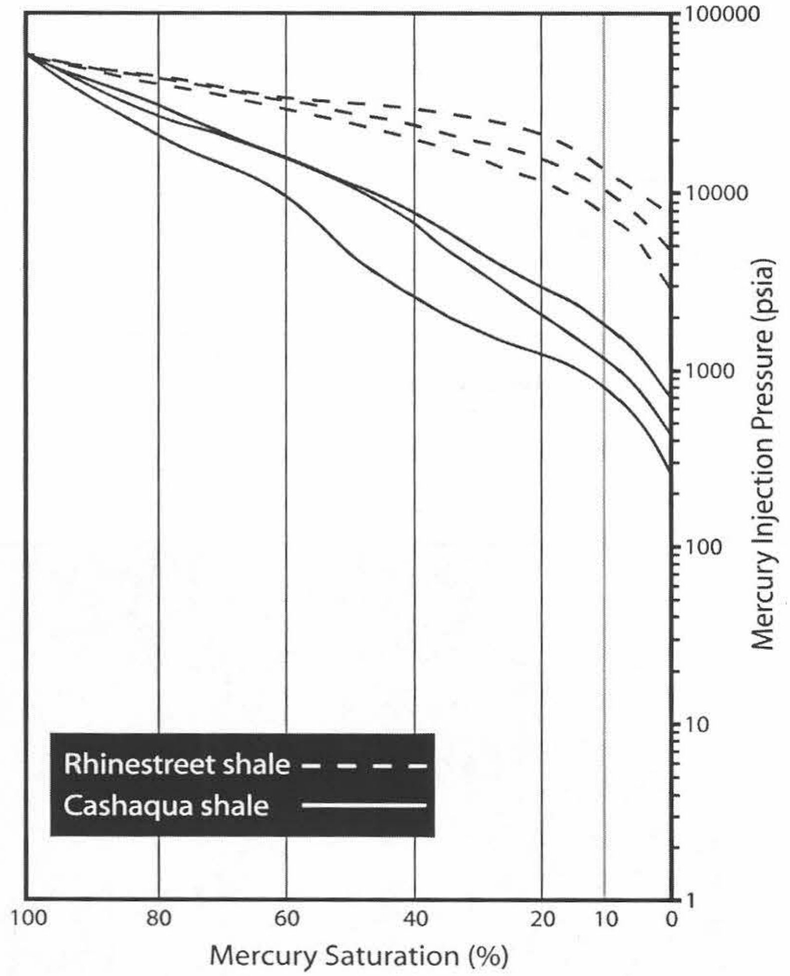


Fig. 5: Mercury injection capillary pressure curves for Rhinestreet and Cashaqua shale samples. Note the line denoting the 10% mercury saturation pressure (a proxy for saturation; see Lash, 2006b).

EARLY COMPACTION AND DIAGENESIS OF THE RHINESTREET SHALE

Mechanical Compaction

The important part that gravitational pressure plays in the progressive consolidation of mud and clay has been recognized for well over fifty years (Hedberg 1926, 1936) and perhaps longer (Sorby 1908). More recent studies (e.g., Heling 1970; Bennett et al. 1977, 1981; Faas and Crocket 1983; O'Brien and Slatt 1990; O'Brien 1995) have presented compelling evidence for the early diagenetic development of a strongly oriented particle orientation in response to burial-related compaction of high-porosity, water-rich flocculated clay. However, other studies have demonstrated that the strongly planar microfabric observed in some shales can form during progressive illitization of smectite by dissolution and re-precipitation under an anistropic stress field (Ho et al., 1999; Charpentier et al., 2003; Aplin et al., 2003).

Carbonate concretions hosted by the Rhinestreet shale provide the opportunity to study pre- and post-compaction clay fabrics of the encapsulating carbonaceous shale. Recently, we (Lash and Blood, 2004a) presented results of an SEM analysis of microfabric variations observed in Rhinestreet black shale samples collected from around early diagenetic carbonate concretions. Crucial to this was establishing that the internally laminated compaction-resistant concretions formed rapidly and, most importantly, close to the sediment-water interface. Field observations, including the wrapping of shale around concretions (Fig.

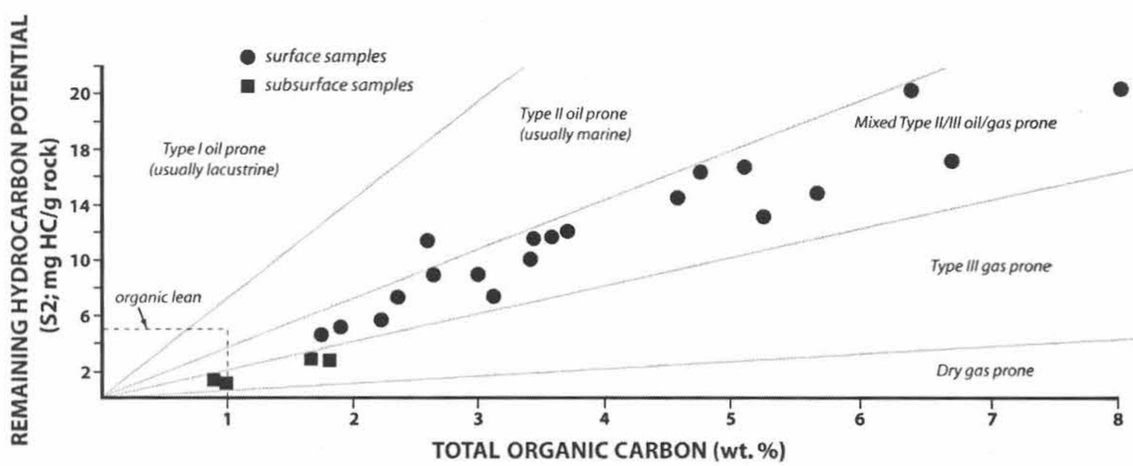


Fig. 6: Plot of TOC vs. Rock-Eval hydrogen index (HI) for Rhinestreet shale samples.

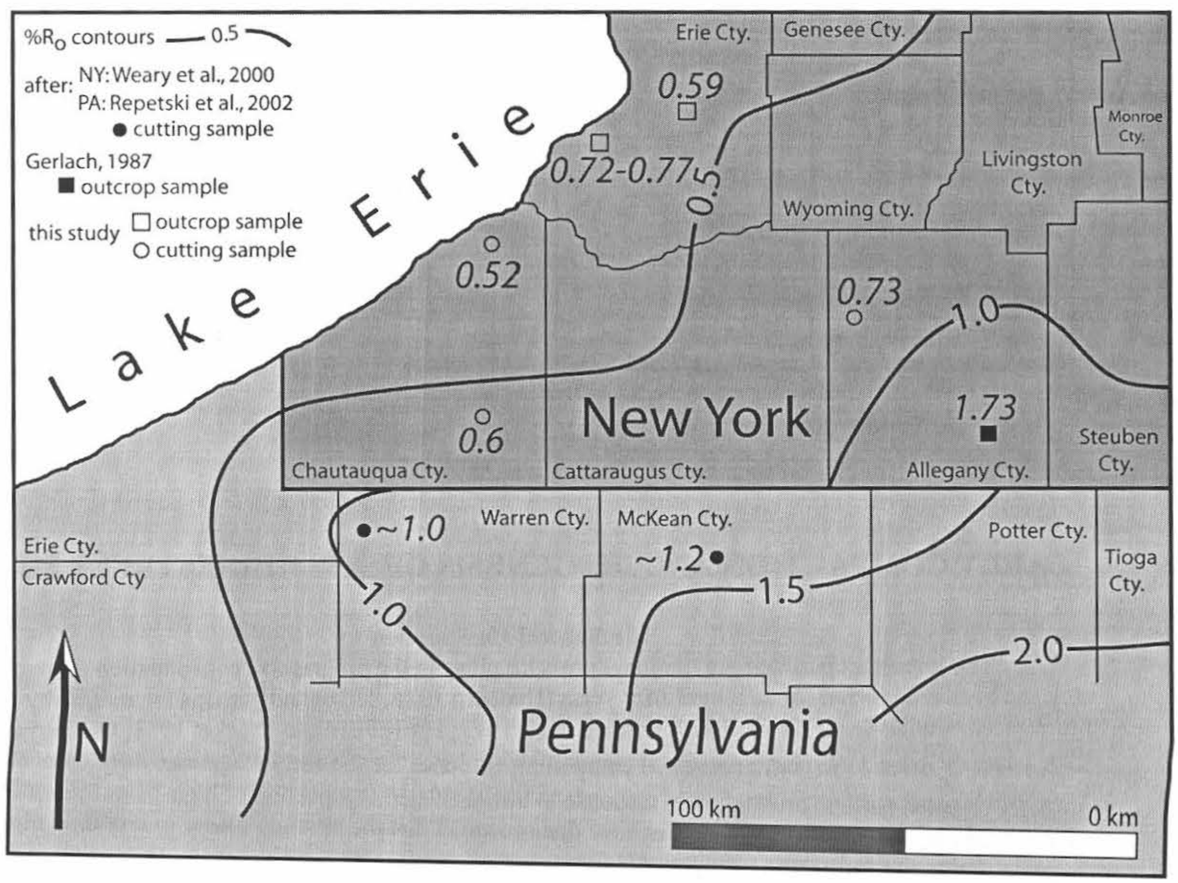


Fig. 7: Map of western New York state and northwest Pennsylvania showing thermal maturity data collected in this study and previous studies of Upper Devonian black shale.

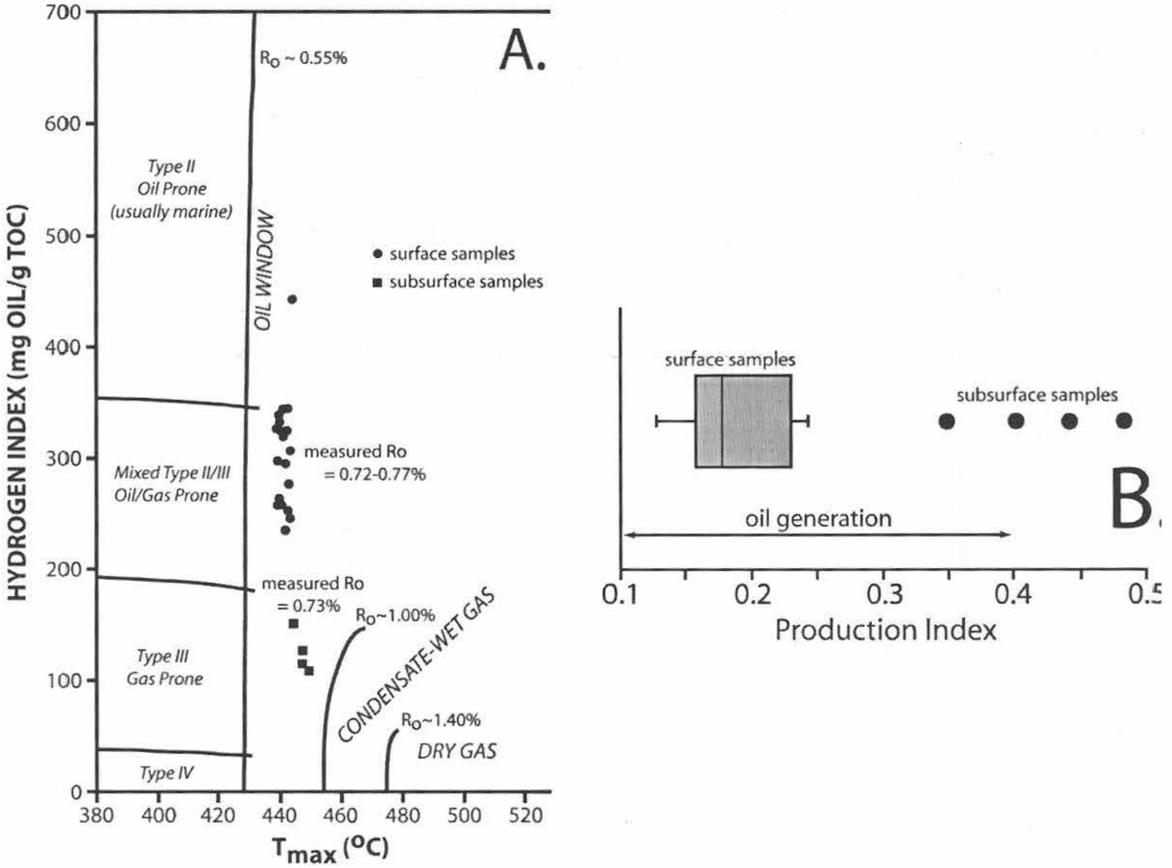


Fig. 8: (A) Van Krevelen plot of HI versus Rock-Eval T_{max} for the Rhinestreet shale samples; (B) box and whisker plot showing the range of production index (Rock-Eval S1/S1+S2) values for the Rhinestreet shale samples.

10A), the lack of center-to-edge deviation in laminae thickness (Fig. 10B), and the tilting of some concretions (Fig. 10C), demonstrate that the concretions formed rapidly at shallow burial depths, perhaps a meter or so below the seafloor (Lash and Blood, 2004a,b). A shallow depth of concretion growth is further suggested by estimates of host sediment porosity based on the assumption, addressed in more detail below, that carbonate cement precipitated passively within the water-filled pore space of the organic-rich clay (e.g., Lippmann 1955; Raiswell 1971; Gautier 1982). Accordingly, the volume percent of carbonate cement in the concretion matrix is a proxy for the porosity of the host sediment at the time of concretion growth (Raiswell 1971). Volume percent of 21 Rhinestreet concretion matrix samples collected from four concretions varies from 74 to 93% (mean = 83%), a range that encompasses the high end of porosity of modern marine clay deposits near the sediment-water interface (e.g., Müller 1967; Velde 1996). In summary, carbonate concretions of the Rhinestreet shale formed at very shallow depth, perhaps within a meter of the sea floor, by rapid, near-uniform, precipitation of diagenetic carbonate in void spaces of the host carbonaceous sediment (see Lash and Blood, 2004a,b for details). The shallow burial depth and rapid rate of growth of Rhinestreet concretions will allow us to use the thickness of layers within concretions (Fig. 10B) as a proxy for the original seafloor thickness of the carbonaceous mud layers immediately prior to, and coeval with, carbonate precipitation (e.g., Raiswell, 1971; Oertel and Curtis, 1972).

Our approach to defining the precompaction microfabric of the Rhinestreet black shale reflects the low compressibility of concretions relative to surrounding unconsolidated clay. The ellipsoidal concretions affected the local compaction-related elastic stress field such that sediment immediately adjacent to lateral edges of concretions - the concretion strain shadow - was shielded from overburden stress. The shielding effect of carbonate concretions in the Rhinestreet shale is obvious at the macroscopic scale. Argillaceous rock within strain shadows is not fissile (Fig. 11, see "ss") and is properly classified as mudstone (Pettijohn 1975). However, these *same* mudstones become fissile less than 30 cm distant the strain shadows, necessitating their reclassification as shale (Fig. 11, see "sh"; Pettijohn 1975). These observations alone

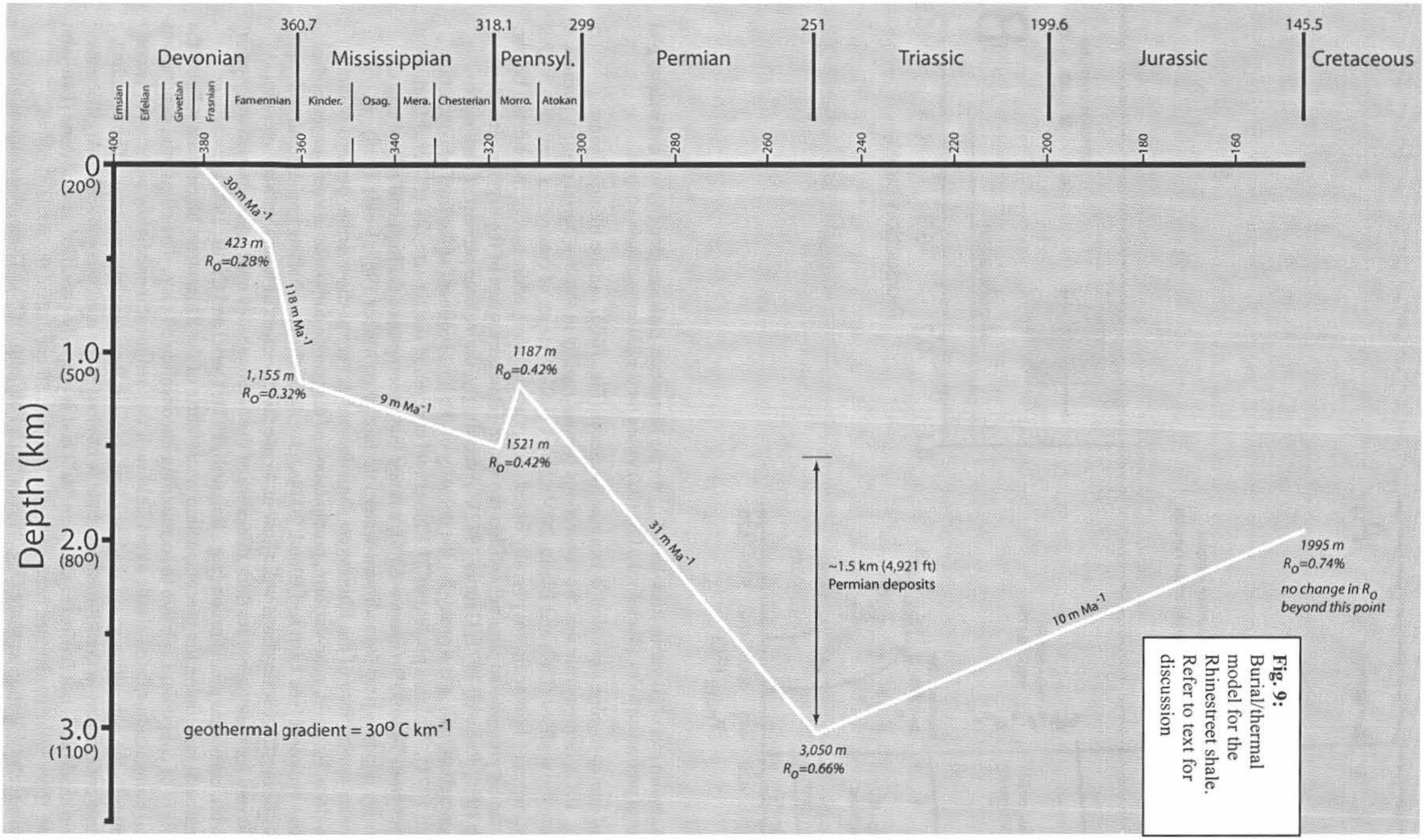


Fig. 9: Burial/thermal model for the Rhinestreet shale. Refer to text for discussion

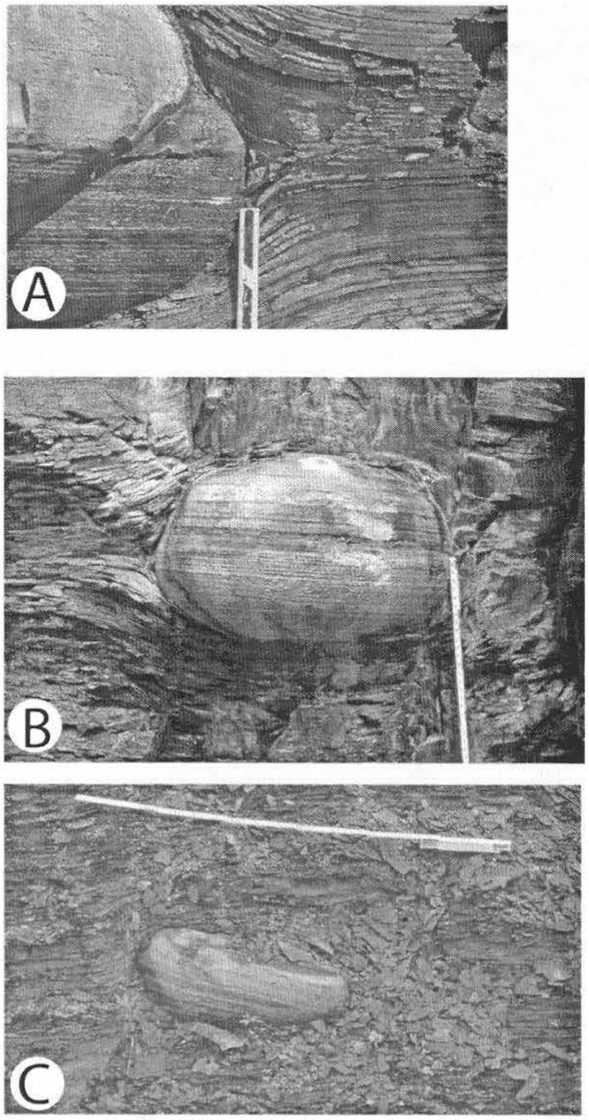


Fig. 10: Carbonate concretions of the Rhinestreet shale:: (A) close-up of differential compaction (note internal laminae); (B) internally laminated concretion; (C) tilted concretion (rule = 1 m).

suggest that the fissility was induced by mechanical compaction and that mudstone within the strain shadows preserves a physical record of the depositional (i.e., pre-mechanical compaction) fabric.

SEM analysis of mudstone samples collected from concretion strain shadows reveals an open fabric of randomly oriented platy particles, which higher magnification proves to be face-to-face clay flake stacks or domains (Fig. 12A). Domains typically are arranged in a low-density network or cardhouse fabric of edge-to-edge and edge-to-face contacts marked by large voids relative to the thickness of clay flakes and domains (Fig. 12B). Secondary electron images of fissile shale samples collected 20 to 30 cm from strain shadows, however, reveal a generally low-porosity microfabric marked by a moderately to strongly preferred orientation of clay flake domains (Fig. 12C). The almost negligible degree of compaction sustained by strain shadow mudstone confirms that gravitational compaction of the Rhinestreet shale was minimal until *after* carbonate concretions had become incompressible thereby affirming the shallow diagenetic origin of the concretions (Lash and Blood, 2004a). Moreover, SEM observations of concretion samples evince an open cardhouse arrangement of detrital clay grains typical of newly deposited flocculated clayey sediment preserved by precipitation of carbonate cement (Fig. 12D; Lash and Blood, 2004b).

The foregoing discussion suggests that the Rhinestreet black shale accumulated in dysoxic to anoxic bottom water as flocculated carbonaceous sediment that underwent rapid gravitational mechanical compaction resulting in the observed strongly anisotropic planar microfabric. Nevertheless, the possible role of chemical dissolution and re-precipitation in producing the planar microfabric documented from the Upper Devonian black shale deposits of the LED cannot be dismissed out of hand. Clues to this question can be gained by microfabric analysis of samples collected from gray shale intervals within the Rhinestreet shale (Fig. 1). Thin section and SEM analysis of these deposits confirms what appear to be the effects of pervasive bioturbation judging from field observations, including the lack of well-defined bedding planes. Notably, the microfabric of the gray shale typically is characterized by an irregular distribution of silt grains throughout a mottled clay matrix (Fig. 13A), the result of bioturbation that homogenized the sediment by mixing silt grains with the clay (e.g., O'Brien and Slatt 1990; Bennett et al. 1991). SEM observations reveal an open microfabric of variably oriented platy grains, many of which appear to be individual clay grains (Fig. 13B). Angular silt grains dispersed throughout the "swirled" clay matrix by bioturbation propped open larger voids during compaction precluding complete reorientation of the disrupted fabric (Fig. 13C). It is obvious that these deposits lack the strongly anisotropic microfabric of the immediately under- and over-lying organic-rich deposits.

We believe that the Rhinestreet gray shale deposits would have acquired a strongly planar microfabric had they had been buried deep enough for illitization kinetics to initiate. Charpentier et al. (2003) point out that this transition in the deepwater Gulf of Mexico sedimentary column occurs at ~5,600

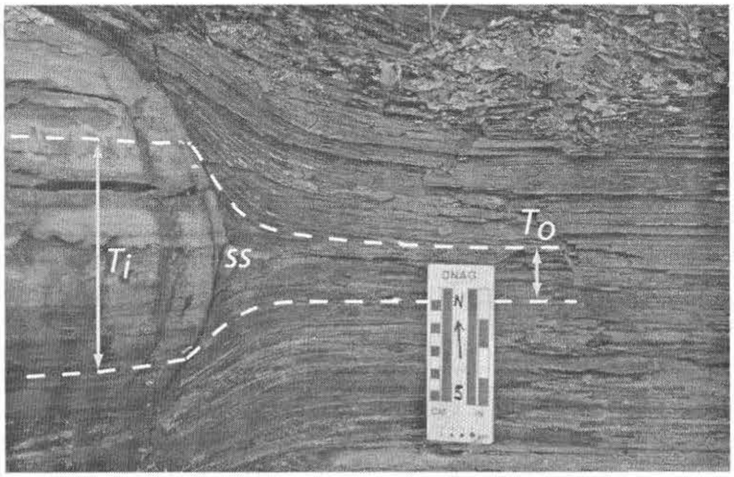


Fig. 11: Rhinestreet shale carbonate concretion and host shale showing elements that were measured for compaction strain analysis. T_i is the thickness of the measured bedding interval within the concretion (the depositional thickness); T_o is the thickness of the measured interval traced into the encapsulating black shale (the compacted interval); ***ss*** is the strain shadow of the concretion (it is here that the incompressible concretion perturbed the local compaction-related elastic stress field so that the sediment was shielded from overburden stress; Lash and Blood, 2004a). The calculated compaction strain of the Rhinestreet shale at this location is -0.56 (56%). The relative thicknesses of T_i and T_o are distorted because the concretion projects ~ 20 cm from the exposure toward the viewer.

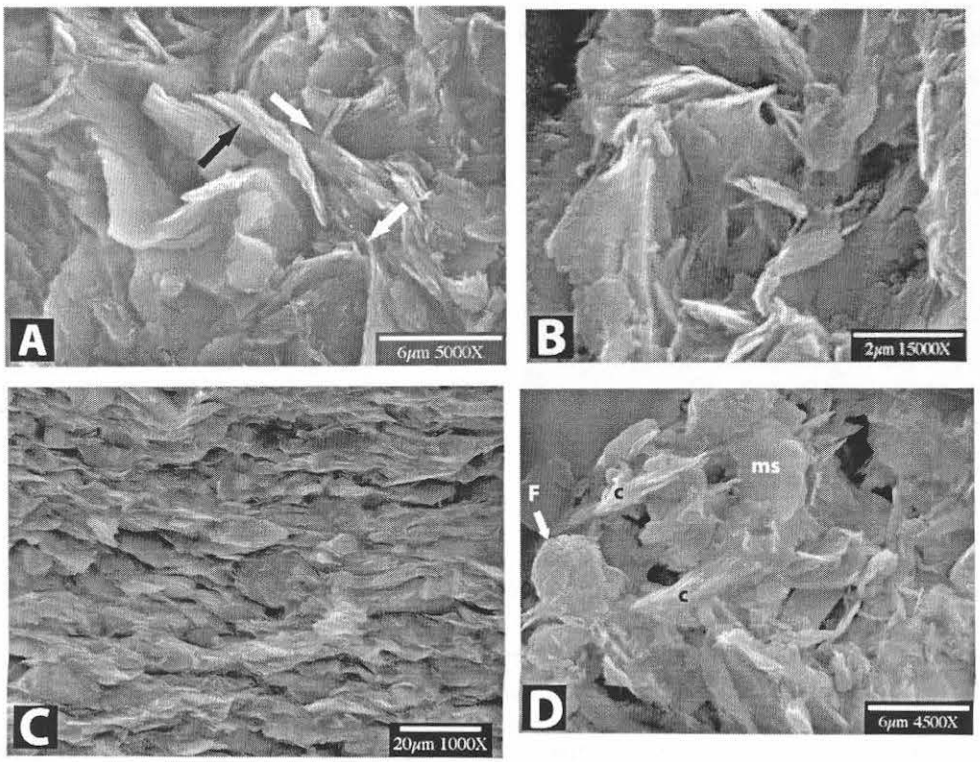


Fig. 12: Secondary electron images of Rhinestreet shale and concretion samples: (A) detail of clay domains and mudstone microfabric. Note face-to-face clay domains (black arrows) and edge-to-edge and edge-to-face domain contacts (white arrows); (B) detail of edge-to-edge domain contacts that define the cardhouse clay fabric. The porous nature of this fabric compares favorably with that of modern flocculated clay; (C) strongly planar (shale) fabric; (D) view of open framework of Rhinestreet carbonate concretions showing a framboid (**F**), randomly oriented clay grain domains (**c**), and microsparite (**ms**).

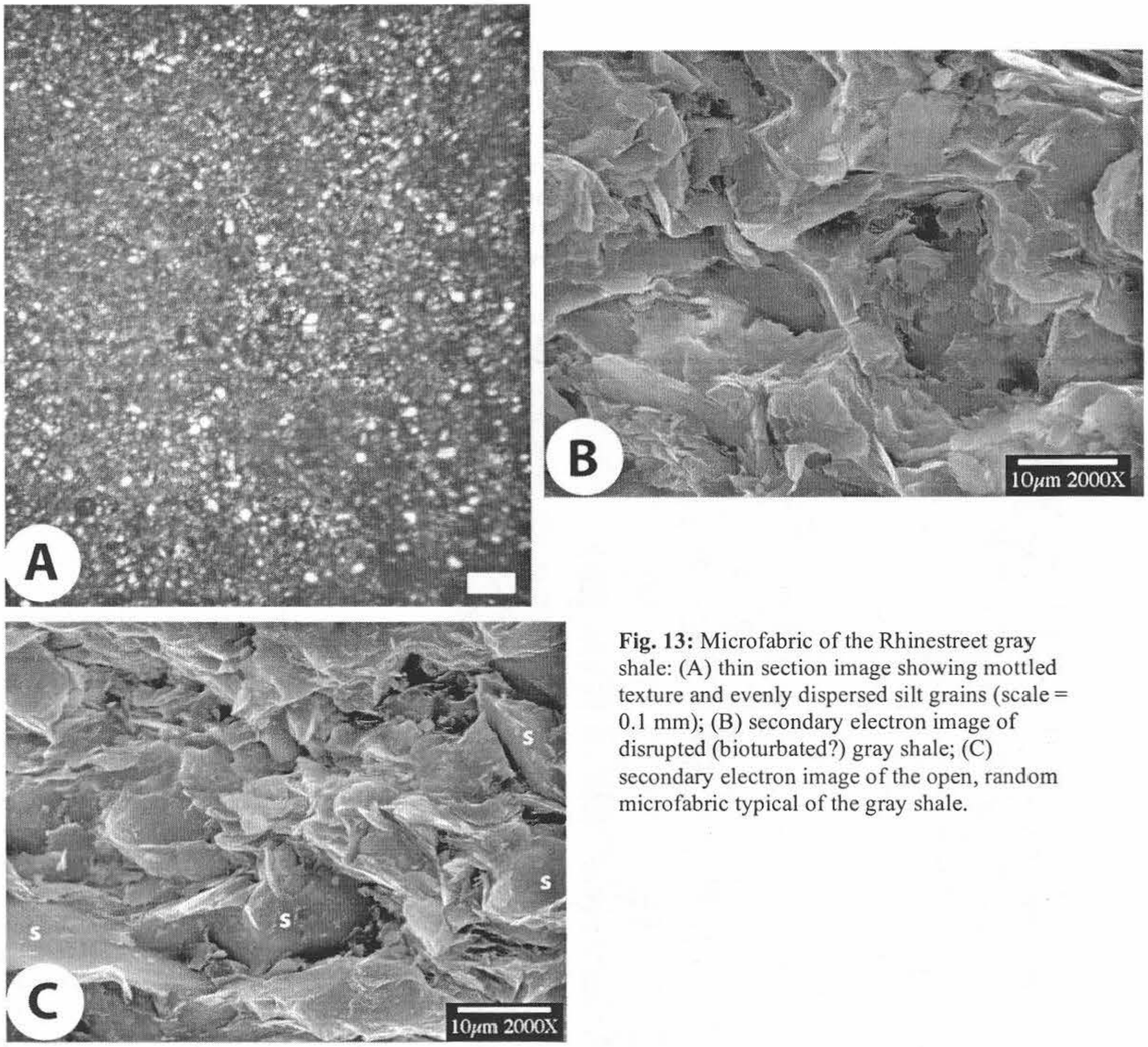


Fig. 13: Microfabric of the Rhinestreet gray shale: (A) thin section image showing mottled texture and evenly dispersed silt grains (scale = 0.1 mm); (B) secondary electron image of disrupted (bioturbated?) gray shale; (C) secondary electron image of the open, random microfabric typical of the gray shale.

m, well in excess of the modeled maximum burial of the Rhinestreet shale. However, Pearson and Small (1988), in their analysis of clay mineral diagenesis in North Sea deposits, demonstrated that illitization ccurs at depths of 2.4 to 3.5 km, within a vitrinite reflectance range of 0.54 and 0.72% and a temperature range of 87 to 100° C, placing the Rhinestreet shale within this zone. As described above, illite is by far the most common clay mineral in analyzed Rhinestreet shale samples, and most Devonian shale samples of the Appalachian Basin for that matter (Hosterman, 1993). However, rather than reflecting a high degree of thermal maturation, the plentiful illite probably tells of a source terrane rich in illite. Strickler and Ferrell (1989), for example, argued that the great abundance of illite in Louisiana Gulf Coast clays reflects derivation from the Appalachian orogen. Moreover, the location of the Rhinestreet shale in the LED, a region of very low illite crystallinity (Hosterman, 1993), is consistent with some of the low $\%R_o$ values measured in samples of Upper Devonian black shale collected from along the Lake Erie shoreline (see Fig. 7). It is difficult at best to envision the development of the strongly aligned microfabric of Rhinestreet black shale samples as a consequence of the smectite-illite transition while not seeing any evidence for this phenomenon in Rhinestreet gray shale samples. We suggest, instead, that the gray shale retained its relatively open fabric because the Rhinestreet was not buried deep enough for smectite-illite reaction kinetics to initiate. Thus, the planar microfabric of the Rhinestreet black shale was acquired rapidly and very early in the compaction history of the organic-rich clay, which, based on comparison with modern carbonaceous fine-grained sediment, was probably water-rich (e.g., Meade, 1966; Keller, 1982) and very compressible.

Carbonate Concretion Growth

Mechanical compaction is only one type of diagenetic modification of the Rhinestreet sediment; the other is the precipitation of carbonate cement in the form of concretions and the scraggy layers (Fig. 1). The majority of carbonate concretions of the Rhinestreet shale are found in three laterally persistent horizons (Fig. 1; Lash and Blood, 2004b). Most concretions are oblate ellipsoids with maximum diameters and thicknesses ranging up to 2.7 m and 1.1 m, respectively (Fig. 14A). Septarian fractures, where they can be observed, extend outward from concretion centers, narrowing to termination near the edges. Brown Fe-poor calcite lines fracture walls and is post-dated by yellow siderite. Pyrite, marked by low but uniform abundance throughout much of the interior of concretions, is concentrated along concretion carapaces as well as halos along septarian fractures. Broken and weathered concretions reveal depositional laminae inherited from the host Rhinestreet black shale that show no obvious systematic changes in thickness across concretions (Fig. 14B).

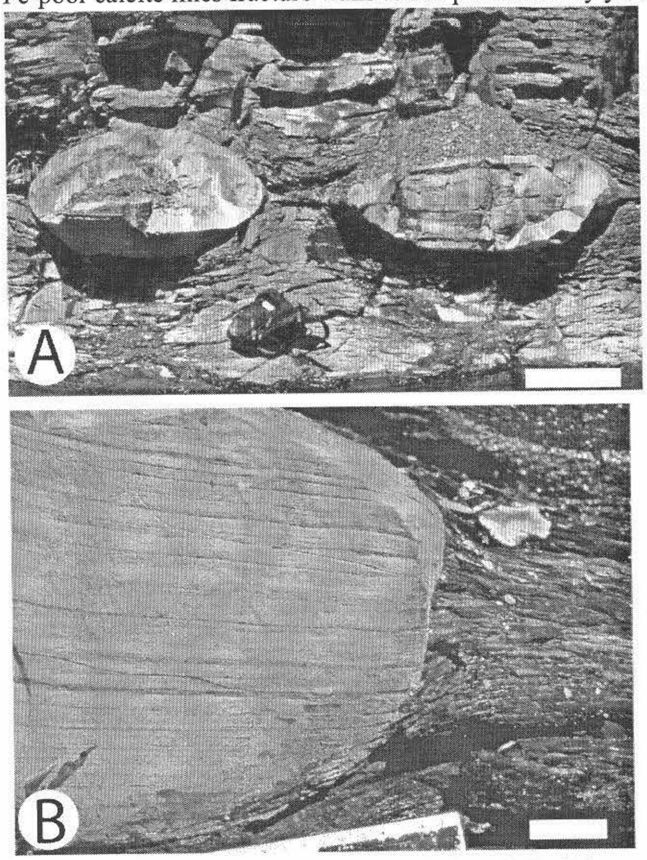


Fig. 14: Carbonate concretions of the Rhinestreet shale: (A) two of the larger ellipsoidal concretions of the middle concretion horizon (scale bar = 0.5 m); (B) carbonate concretion displaying laminae (bar = 3 cm).

The conventional explanation for the formation of carbonate concretions assumes that growth occurred concentrically by addition of successive layers of cement during burial (Oertel and Curtis, 1972; Mozley, 1996). Such a sequential growth history has been recognized by decreasing diagenetic carbonate percentage from the centers to edges of concretions reflecting progressively decreasing pore-space volume of the host sediment during concretion growth (e.g., Raiswell, 1971; Oertel and Curtis, 1972). Five studied Rhinestreet concretions show diminishing carbonate in center-to-edge traverses, the maximum reduction being 8% to a value of 74% at the edge of concretion EMC29 (Fig. 15), suggesting only a minor reduction in porosity of the host sediment during concretionary growth. One concretion (concretionUC5) is characterized by more than 90% carbonate in samples collected from the interior of the concretion whereas center and edge samples contain ~ 85% carbonate (Fig. 15). The minimal radial carbonate gradients exhibited by Rhinestreet concretions indicate that the concretions grew largely by rapid near-uniform precipitation of carbonate cement throughout the concretion body thereby confirming a similar interpretation based on the lack of center-to-edge deviation in laminae thickness (Lash and Blood, 2004b). Moreover, microscopic observations described earlier in this section indicate that concretions grew as compaction resistant frameworks of calcite micrite and microspar in porous organic-rich clay perhaps a meter or so below the seafloor (Fig. 12D; Lash and Blood, 2004b)). The open framework of the newly formed concretions provided strength yet reduced the mean density of most concretions precluding their sinking into the weak, low-density clay (e.g., Wetzel, 1992; Raiswell and Fisher, 2000).

Stable element isotopic data adds to our understanding of the diagenetic history of the Rhinestreet shale and its concretions. The range of $\delta^{13}C$ values is rather wide (-13.9 to +1.7 ‰ PDB) while that of $\delta^{18}O$ is limited (-6.8 to –3.8 ‰ PDB) (Fig. 16). Isotopic values of concretions sampled in the middle and lower

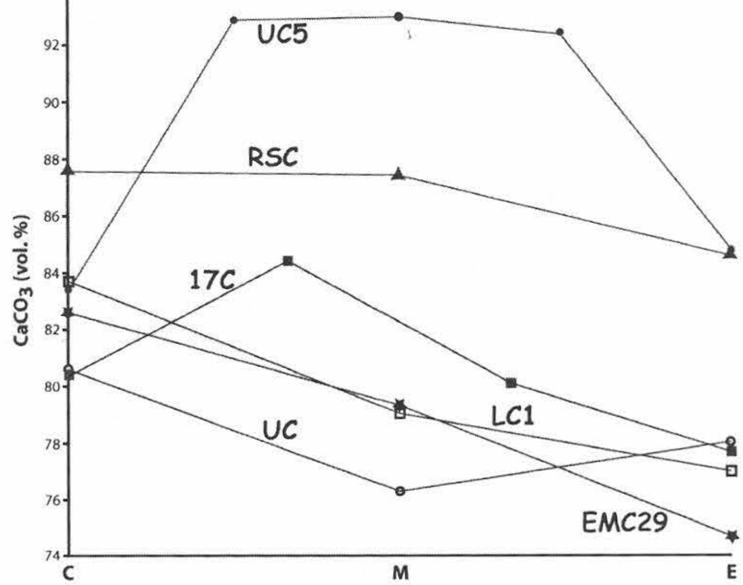


Fig. 15: Plot of diagenetic volume percent carbonate versus sample location within the concretion (C = center; M = middle; E = edge).

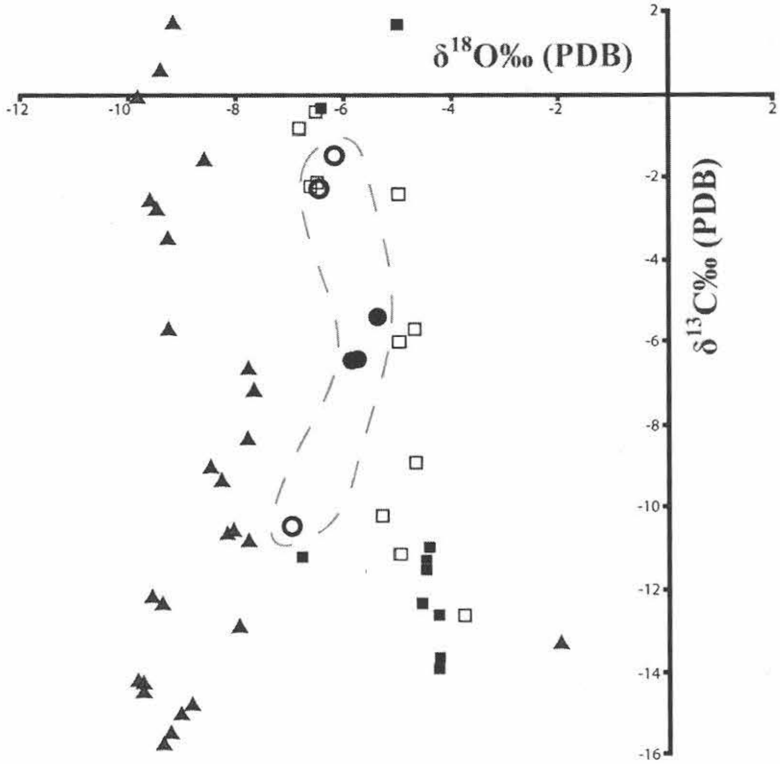


Fig. 16: $\delta^{18}O$ and $\delta^{13}C$ scatter diagram for Rhinestreet carbonate concretions (filled squares = lower concretion horizon; open squares = middle concretion horizon) and scraggy layers (filled circles = middle scraggy layer; open circles = upper scraggy layer). The plot includes the data of Dix and Mullins (1987) represented by triangles for comparison.

concretionary horizons are similar though less depleted ^{13}C values appear to be somewhat more common to concretions of the middle horizon (Fig. 16). Four of the six studied concretions show radial enrichment in ^{13}C (Fig. 17A). Nevertheless, concretion 17C displays minimal radial variation in $\delta^{13}C$, and center and edge $\delta^{13}C$ values of concretion UC5 are more enriched in $\delta^{13}C$ than samples collected from the mid-interior region of the concretion (Fig. 17A). Oxygen isotope compositions of the studied Rhinestreet concretions show minor radial variations (Fig. 17B). Four concretions display slight center-to-edge depletion in ^{18}O, and one concretion (concretion 17C) shows no change in $\delta^{18}O$. Center and edge samples collected from concretion UC5 are characterized by equally depleted $\delta^{18}O$ compositions whereas samples recovered from the interior of this concretion are less depleted (Fig. 17B). Concretions UC and EMC29 display center-to-edge enrichment in ^{13}C that is complemented by depletion in ^{18}O. However, concretions LC1 and RSC show almost identical radial increases in heavy carbon but dissimilar center-to-edge variations in $\delta^{18}O$. The significance of these seemingly random radial variations in stable element isotopes will be addressed below.

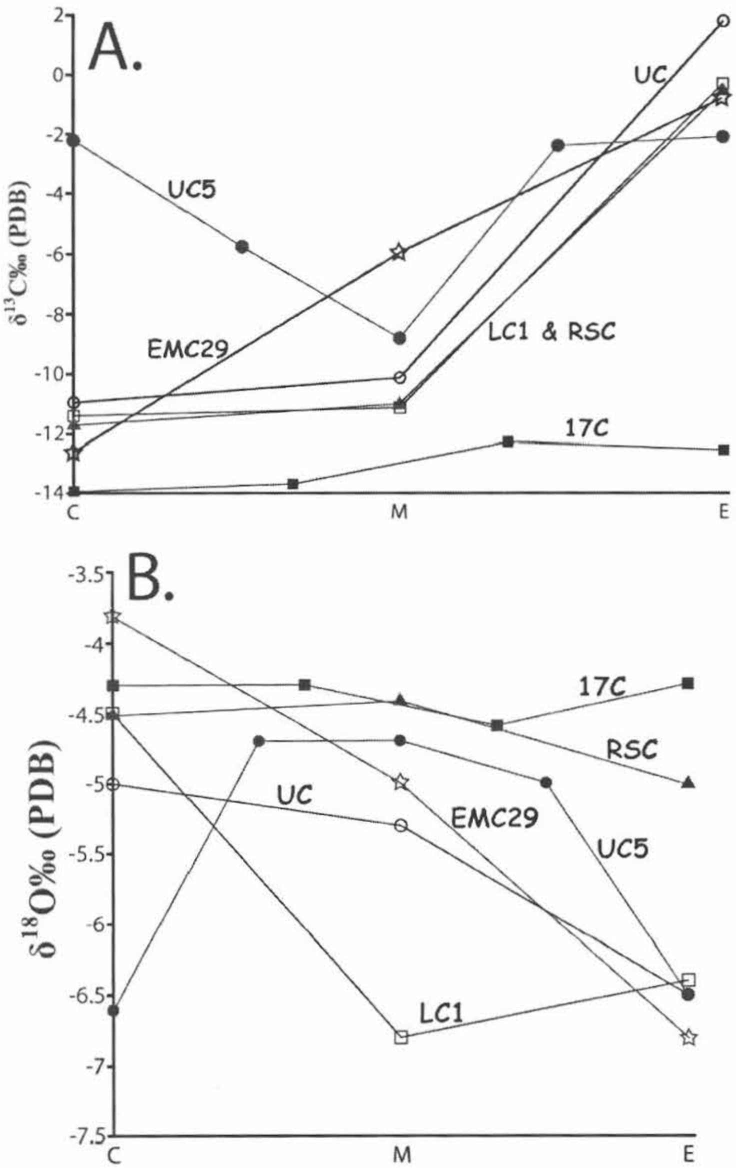


Fig. 17: Plots of $\delta^{13}C$ (A) and $\delta^{18}O$ (B) versus sample location within the concretion (C = center; M = middle; E = edge).

The formation of Rhinestreet concretions along specific horizons, an occurrence common to other black shale units (e.g., Raiswell, 1971; Hudson, 1978; Raiswell and White, 1978; El Albani et al., 2001), requires explanation. The siting of carbonate concretions has been related to the microbial reduction of locally concentrated organic matter, including skeletal remains and tissue (e.g., Weeks, 1957; Zangerl et al., 1969; Berner, 1980). However, we observed no textural evidence to suggest such a siting mechanism for the Rhinestreet concretions, nor do the concretions show a predilection for the organic-rich base of the unit (see Fig. 1). Moreover, organic matter within concretions is strikingly less than that in encapsulating shales (Lash and Blood, 2004b), though compaction-related volume loss of the shale no doubt elevated host rock TOC. Nevertheless, depleted carbon isotopic compositions (Figs. 16 and 17A) indicate that concretionary carbonate originated in part from the anaerobic bacterial reduction of organic matter. The large size of the Rhinestreet concretions suggests that the limited amount of *in situ* organic carbon was augmented by carbon derived from other sources during diagenetic precipitation of carbonate cement. A likely source of the additional carbon was the oxidation of biogenic methane at the base of the sulfate reduction zone by anaerobic methane oxidation (AMO) (Fig. 18; Raiswell, 1987, 1988). According to this mechanism, biogenic methane produced by microbial decomposition of organic matter in the zone of methanogenesis diffuses upward to the base of the sulfate reduction zone where carbonate precipitation occurs, usually within less than a meter of the seabed (Fig. 18). The isotopically light methane leaves behind a pool of ^{13}C-enriched carbon dioxide (Irwin et al., 1977) some of which migrates upward through the sediment. The siting of concretions within a concretionary horizon reflects the locations of permeable vertical pathways capable of conducting methane to the base of the sulfate reduction zone (Raiswell, 1987). Moreover, differences among $\delta^{13}C$ gradients of the studied Rhinestreet concretions (Fig. 17A) probably reflect local variations in the flux of methane and dissolved carbonate diffusing upward from the zone of methanogenesis (e.g., Raiswell, 1988).

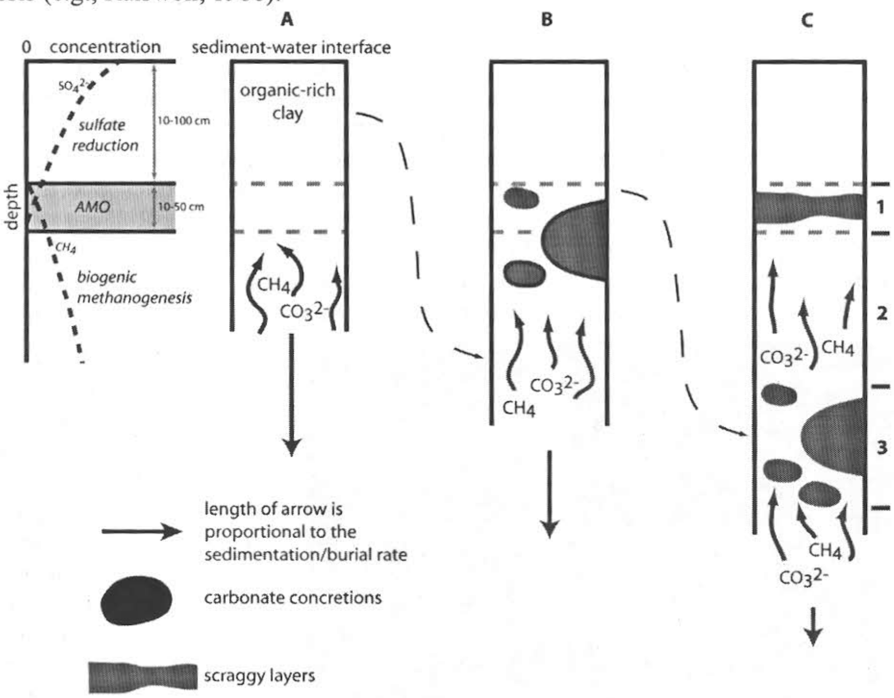


Fig. 18: Schematic representation of the inferred role of subsidence/burial rate in the growth of carbonate concretions by anaerobic methane oxidation (modified from Raiswell, 1987, 1988). The column on the left shows the diagenetic zones in anaerobic sediment; AMO = zone of anaerobic methane oxidation. Depth and thickness of the zone of AMO is based on Raiswell (1987) and references therein. Column A: subsidence/burial rates are high enough to carry sediment through the zone of AMO so fast that little more than the diffuse precipitation of diagenetic calcium carbonate occurs. Column B: a reduced burial rate enables concretions to grow in a relatively wide zone as sediment passes slowly through the zone of AMO (e.g., the lower concretionary horizon in the Rhinestreet shale and level 3 in column C). Column C: a near or complete cessation of sedimentation (very slow passage of sediment through the zone of AMO) results in the formation of a scraggy layer. Level 2 in column C reflects an increase in subsidence/burial rate following formation of the concretionary horizon represented by level 3.

Crucial to the precipitation of diagenetic carbonate as a consequence of AMO is a marked reduction in, or complete stoppage of, sedimentation/burial, which would stabilize the AMO zone thereby enabling protracted carbonate precipitation (Fig. 18; Raiswell, 1987, 1988). Two observations provide indirect evidence of a reduction in sedimentation/burial rate during concretion growth: 1) minimal change in estimated porosity (based on carbonate volume percentage) outward from the centers of studied concretions and 2) negligible variation in laminae thickness within concretions (see Figs. 10B and 14B). Both points suggest that the concretions grew in the near-absence of sedimentation/burial and associated compaction while the concretions formed (e.g., Raiswell, 1971).

Oxygen isotopic compositions of calcite are known to be a function of the temperature of precipitation and the $\delta^{18}O$ of the water from which the calcite precipitated (Hudson, 1977). If one variable is reasonably well known, the other can be estimated (Hudson, 1978). We used the equation of Anderson and Arthur (1983) to calculate paleotemperatures of concretionary calcite precipitation:

$$T^{o} = 16.0 - 4.14\,(\delta C - \delta W) + 0.13\,(\delta C - \delta W)^2 \qquad (1)$$

in which $\delta C = \delta^{18}O$ PDB of diagenetic calcite and $\delta W = \delta^{18}O$ of seawater on the SMOW scale. Estimates of seawater $\delta^{18}O$ for the Late Devonian include –1 ‰ SMOW of van Geldern et al. (2001) and –2 ‰ SMOW of Hudson and Anderson (1989). A Devonian seawater isotopic composition of –2 ‰ SMOW yields a temperature range of carbonate precipitation of 24° – 39° C, slightly to moderately in excess of the inferred 20° C bottom water temperature of the subtropical Catskill Delta basin (e.g., Gerlach and Cercone, 1993) and, therefore, incompatible with a shallow depth (~ 1 m below the seafloor) of concretion growth argued for by textural and field observations. Indeed, the range of oxygen isotopic values of Rhinestreet concretionary carbonate suggests that diagenetic precipitation occurred 130 to 630 m below the seafloor, assuming a geothermal gradient of 30° C/km.

An influx of ^{18}O-depleted meteoric water through the forming Rhinestreet concretions could account for their isotopic compositions, yet the location of the Rhinestreet depocenter, far distant from the Devonian shore would seem to preclude such a scenario. Direct precipitation of the Rhinestreet concretions from seawater at the temperatures suggested by the depleted oxygen isotopic values would require abnormally warm bottom waters during the Late Devonian, a hypothesis not supported by any geological data. We suggest, instead, that the oxygen isotope compositions of the Rhinestreet concretions resulted from the incomplete recyrstallization of primary calcite micrite to microspar (e.g., Folk, 1965) and re-equilibration of isotopic values at depth, perhaps by warm fluids migrating through the section. Although four of the Rhinestreet concretions studied by us show minor center-to-edge depletion in ^{18}O (Fig. 17B), estimated temperatures of precipitation of the centers of these concretions are too high to suggest precipitation from normal seawater. Further, concretion 17C shows no radial oxygen isotope gradient, and the center of concretion UC5 is marked by an especially depleted (-6.6 ‰ PDB) value (Fig. 17B). It appears, then, that warm fluids had some degree of access to concretion interiors. Dix and Mullins (1987) described a similar occurrence of recrystallized carbonate concretions in the more deeply buried Middle Devonian sequence of central New York state that they attributed to the migration of warm connate fluids, as much as 30°C warmer than those fluids inferred to have affected the Upper Devonian Rhinestreet concretions of western New York state (see Fig. 16).

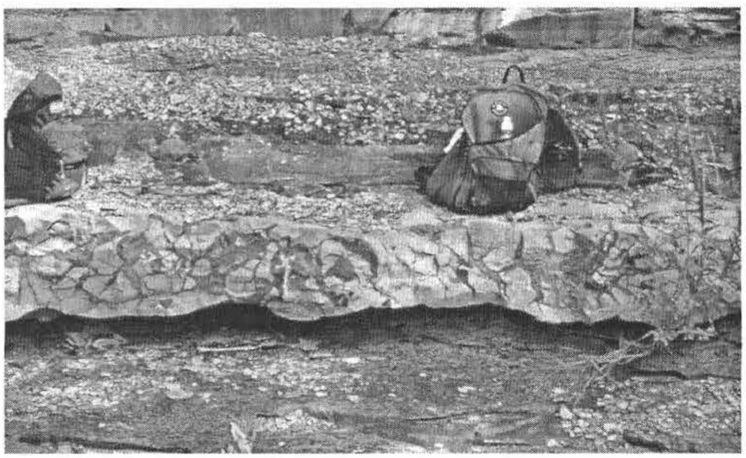


Fig. 19: Upper scraggy layer, Eighteenmile Creek.

Finally, we consider the mode of formation of the scraggy layers (Fig. 19). The occasional presence of ellipsoidal carbonate concretions embedded within scraggy layers indicates that they, too, are diagenetic and likely formed in a manner similar to the concretions; i.e., AMO. Indeed, stable isotope values of samples recovered form the middle and upper scraggy layers (see Fig. 1) fall within the range of values of the carbonate concretions (Fig. 16). The geometry of the scraggy layers, noticeably different from the ellipsoidal concretions, is interpreted to reflect a near stoppage of sedimentation/burial. Building upon our discussion of AMO, the thickness of a diagenetic carbonate horizon is interpreted to be proportional to the change in sedimentation/burial rate; i.e., a complete hiatus yields the minimum vertical range of carbonate precipitation because burial is halted and the zone of AMO is stabilized (Raiswell, 1987, 1988). The thickness and geometry of the scraggy layers likely reflects a greater duration of the break in sedimentation/burial during their growth than during the development of the concretionary horizons and, therefore, a protracted period of time that isotopically light methane and ^{13}C-enriched dissolved carbonate were delivered to the stationary AMO zone (compare columns B and C, Fig. 18; Raiswell, 1987, 1988). That is, the near cessation of sedimentation and, most importantly, burial, resulted in the confinement of carbonate precipitation to a narrow, fixed zone defined by the upper limit of methane and CO_3^{2-} diffusion and the maximum depth to which seawater sulfate can diffuse (Raiswell, 1988). The scraggy layers, by virtue of their thickness and their semi-continuous geometry, then, probably formed as a consequence of the stoppage of sedimentation and burial so that the zone of AMO was held in place long enough for diagenetic carbonate to accumulate laterally rather than vertically.

We have recognized the upper scraggy layer in wireline logs from hydrocarbon wells as distant as Chautauqua County (Fig. 20). Assuming that basinwide dynamics were responsible for the prolonged cessation of sedimentation and burial, the scraggy layers might be used both in outcrop and subsurface work to create a time-stratigraphic framework in the Rhinestreet shale. This is a fertile direction for future work.

EARLY (ACADIAN) OVERPRESSURE EPISODE – DISEQUILIBRIUM COMPACTION

The Rhinestreet shale likely followed a normal compaction history that can be described in terms of the exponential decay function for shale first proposed by Athy (1930),

$$\phi = \phi_i e^{-cz} \qquad (2)$$

where z is depth in meters, ϕ_i is the initial porosity at $z = 0$, and e is the compaction coefficient, for some time, perhaps to its maximum burial depth. Burial-induced normal mechanical compaction of argillaceous sediment is accomplished by a loss of porosity as sediment particles respond to increasing effective stress by reorienting into more mechanically stable arrangements (i.e., perpendicular to overburden stress) and pore fluid is expulsed (Hedberg, 1936; Terzaghi and Peck, 1948; Hamilton, 1976; Magara, 1978; Jones and Addis, 1985; O'Brien and Slatt, 1990; Goulty, 2004). The majority of porosity-depth algorithms created from empirical data (e.g., Sclater and Christie, 1980; Huang and Gradstein, 1990; Hansen, 1996; among others) define by a rapid reduction in porosity at shallow depth, followed by a reduced rate of porosity occlusion in progressively older, more deeply buried sediment.

Under certain conditions, however, notably when fluid expulsion during burial is retarded due to low permeability and/or rapid sedimentation, mechanical compaction fails to keep pace with increasing vertical effective stress such that pore pressure is greater than hydrostatic (Swarbrick et al., 2002). This phenomenon, termed *disequilibrium compaction*, has been documented from modern basins including the Carnarvon Basin, western Australia (van Ruth et al., 2004), the Gulf Coast Basin, offshore Louisiana (Dickinson, 1953; Harrison and Summa, 1991; Hart et al., 1995; Audet, 1996), the Caspian Sea (Bredehoeft et al., 1988), offshore eastern Trinidad (Heppard et al., 1998), the North Sea (Mann and Mackenzie, 1990; Audet and McConnell, 1992; Swarbrick and Osborne, 1998), and southeast Asia (Harrold et al., 1999). Overpressure caused by disequilibrium compaction begins at that depth where sediment permeability becomes so low that mechanical compaction and, therefore, porosity reduction is retarded (e.g., Swarbrick et al., 2002). The onset of overpressure occurs at the *fluid retention depth*, which is recognized on porosity-depth profiles as that depth at which porosity shows no reduction with increasing burial depth and on pressure-depth profiles as that depth where pore pressure exceeds hydrostatic (Fig. 21; Audet, 1996; Harrold et al., 1999; Swarbrick et al., 2002).

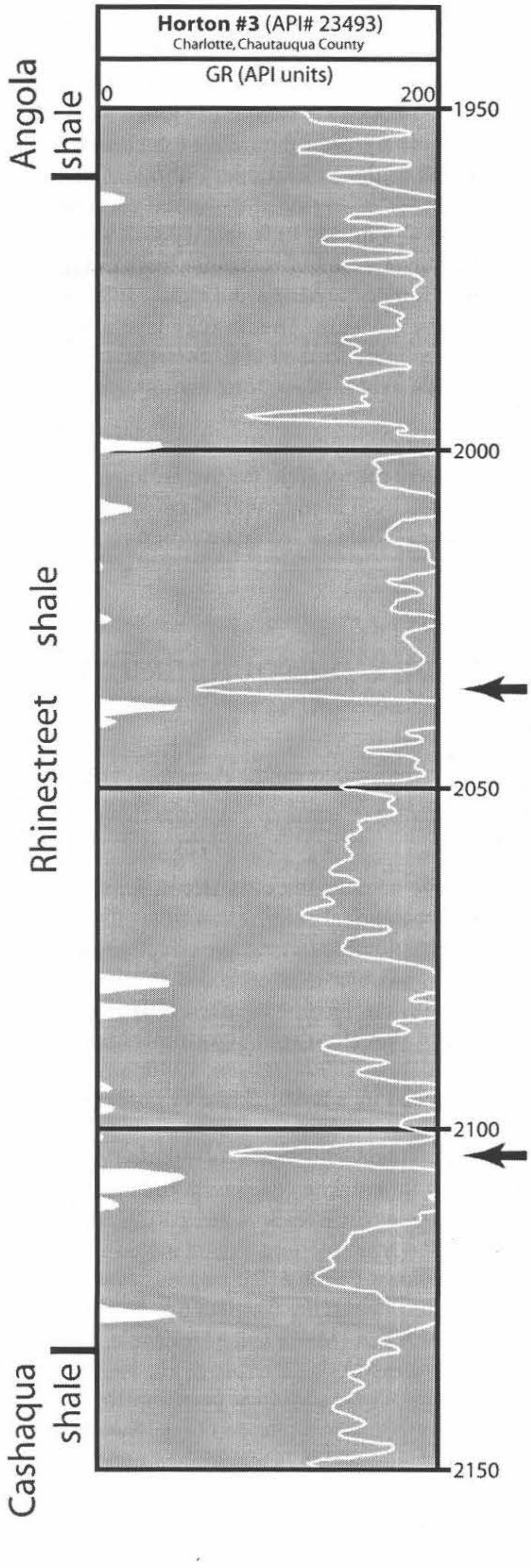


Fig. 20: Gamma-ray log of the Horton #3 well, Chautauqua County, New York. Black arrows point to inferred scraggy layers of the Rhinestreet shale.

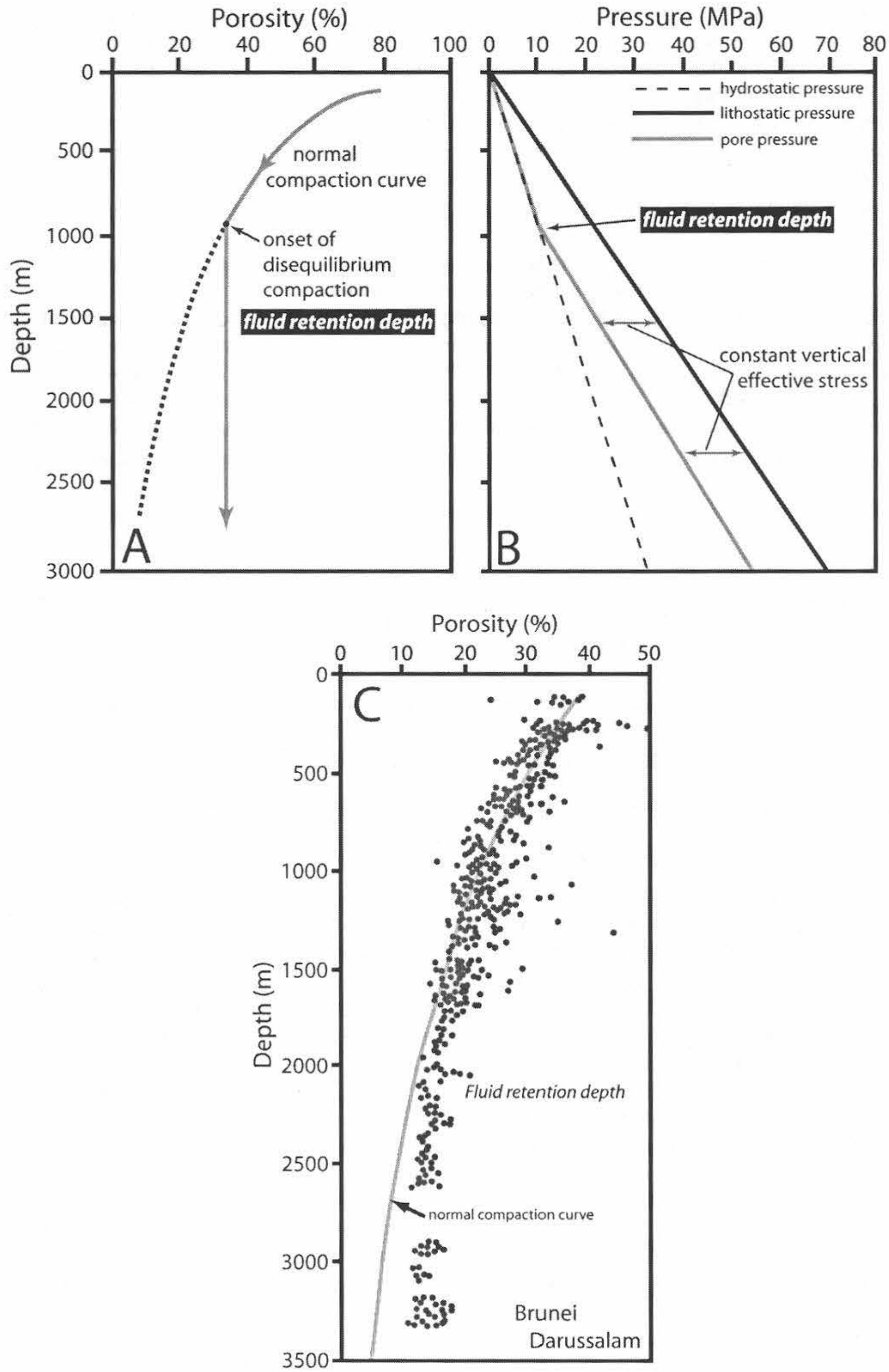


Fig. 21: Relationship between (A) porosity and depth and (B) pressure and depth for a shale that becomes overpressured by disequilibrium compaction at the fluid retention depth (modified after Harrold et al., 2000). (C) Porosity-depth profile for Brunei Darussalam (modified after Tingay et al., 2000).

Abundant natural hydraulic fractures (NHFs) or joints in the Upper Devonian shale-dominated succession of the western New York state region of the Appalachian Basin indicate that these deposits were episodically overpressured during their burial history (Engelder and Oertel, 1985; Lacazette and Engelder, 1992; Lash et al., 2004; Lash and Engelder, 2005). As described below and elsewhere, overpressuring of the shale and resultant natural hydraulic fracturing resulted from the transformation of kerogen to hydrocarbons in organic-rich source rocks and occurred close to or at maximum burial depth during the Alleghanian orogeny (Lash et al., 2004; Engelder and Whitaker, 2006). Because disequilibrium compaction alone is insufficient to generate pore pressures, P_p, capable of driving NHFs (Hart et al., 1995; Kooi, 1997), such an overpressuring event in the burial history of a basinal succession may go unrecognized. However, analysis of compaction strain data described below suggests that the Rhinestreet shale became overpressured well before attaining its maximum burial depth.

The most obvious measure of gravitational compaction strain sustained by a volume of sediment following accumulation on the seafloor is the change in layer thickness from the concretion into correlative layers of the encapsulating shale (see Fig. 11). We measured the thickness of bedding or a bedding interval inside concretions (T_i) and the thickness of that same interval in the shale (T_o), a presumed proxy for the original seafloor thickness of the host sediment as argued above (Fig. 11). Gravitational compaction strain of the shale outside the strain shadow of the concretion, ε_3, is calculated by the following expression,

$$\varepsilon_3 = \frac{T_i - T_o}{T_i}. \qquad (3)$$

The mean ε_3 of the Rhinestreet black shale based on the analysis of 118 concretions and encapsulating shale throughout the unit, expressed as a negative value, is –0.518 or 51.8% (± 4.9%). This value is noteworthy because normally compacted marine shales typically compact more than 65% upon burial to depths comparable to the maximum burial depth of the Rhinestreet shale (Burland, 1990).

The compaction strain of the Rhinestreet shale can be used as a measure of the porosity achieved at the termination of gravitational mechanical compaction if we assume that all volume loss was caused by vertical shortening, a reasonable assumption based on the lack of layer-parallel shortening caused by Alleghanian tectonics in these rocks (Hudak, 1992). Compaction strain is converted to paleoporosity, ϕ_p, by the following equation derived by Jacob (1949),

$$\phi_p = \frac{\phi_o + 100\varepsilon_3}{\varepsilon_3 + 1} \qquad (4)$$

in which ϕ_p is expressed as volume percent. However, in order to calculate the ϕ_p of the Rhinestreet shale, we first must obtain a reasonable value for the porosity of the sediment at the onset of normal compaction, ϕ_o. Textural evidence described above and in more detail by Lash and Blood (2004b) relates the formation of Rhinestreet concretions to the passive precipitation of diagenetic carbonate in void spaces of the clayey organic-rich host sediment. As noted earlier, the porosity of the Rhinestreet sediment at the time of concretion growth ranged from 74 to 93%. However, some authors have suggested that the porosity of newly deposited clayey sediment decreases from as much as 90% to perhaps 60-65% within ten m or so of the seafloor (Weller, 1959; Von Engelhardt, 1977; Magara, 1978; Luo et al., 1993). Moreover, Kawamura and Ogawa (2004) demonstrated an especially rapid reduction in void ratio of pelagic clay equal to a 5% drop in porosity down to a depth of 10 cm below the seafloor. Luo et al. (1993) maintain that such marked losses of porosity within several meters to a few tens of meters of the seafloor should be considered a continuation of the depositional process rather than the initial phase of normal load-induced mechanical compaction. Thus, we interpret the range in $CaCO_3$ volume in analyzed Rhinestreet concretions to reflect the rapid occlusion of porosity as the carbonaceous clay entered and passed through the sulfate reduction zone where concretionary growth occurred (Lash and Blood, 2004b). However, minor geochemical (i.e., $CaCO_3$) zonation observed in individual Rhinestreet concretions (see Fig. 15) suggests that concretionary growth occurred rapidly relative to the rate of porosity reduction.

In light of the preceding discussion, we arbitrarily set $\phi_o = 70\%$ for our calculation of the Rhinestreet shale ϕ_p. This value is a bit higher than the 60-65% porosity level that Luo et al. (1993) consider to mark the onset of normal compaction principally because of the typically high water content of organic-rich clays (e.g., Meade, 1966; Keller, 1982). Our calculated Rhinestreet shale ϕ_p using equation (4) is 37.8% (±7.1%), a value markedly higher than that expected for shale normally compacted to the 3.1 km maximum burial depth of the Rhinestreet shale (see Fig. 9). Because porosity reduction by normal compaction is essentially nonelastic (Harrold et al., 1999), uplift of the Rhinestreet shale cannot account for the high ϕ_p. Indeed, the present porosity of the Rhinestreet shale based on MICP measurements is 3.9% (±0.9%), roughly one-tenth the calculated ϕ_p based on compaction strain analysis. We suggest that the Rhinestreet shale ϕ_p reflects the arrest of gravitational compaction short of its maximum burial depth. The reduction of porosity from the ϕ_p to the present porosity likely reflects a combination of processes, including pressure solution (Towarak, 2006), precipitation of cement, neoformation of clay minerals (chlorite, kaolinite), the production of bitumen when the Rhinestreet shale entered the oil window (Lash, 2006c), and the crushing and/or squeezing of soft grains, including kerogen, into pore throats and voids (e.g., Lash, 2006b).

The relatively low ϕ_p of the Rhinestreet shale suggests that normal compaction of the low-permeability organic-rich clay was arrested at some depth shy of its maximum burial depth by P_p in excess of hydrostatic. This depth, the paleo-fluid retention depth (PFRD), can be estimated in two ways; (1) comparison of the Rhinestreet ϕ_p with published porosity-depth profiles for shale and (2) calculation of the depth at which a porosity of 37.8% is reached during normal compaction of shale by use of empirical porosity-depth algorithms. Published porosity-depth relationships vary among basins as a consequence of such factors as the age of the sediment, effective mean stress, mineral composition, sedimentation rate, lateral stress, and pore fluid chemistry (e.g., Rieke and Chilingarian, 1974; Magara, 1980; Hermanrud, 1993; and many others). Indeed, there is no universal porosity-depth profile (Liu and Roaldset, 1994). Recognizing this, we seek only to bracket the PFRD of the Rhinestreet shale. Plotting the calculated mean Rhinestreet ϕ_p and its upper and lower standard deviation limits on Rieke and Chilingarian's (1974) summary plot of normal compaction curves for shale suggests that the PFRD could have ranged from a depth of ~690 m to ~1,380 m, less than half-way to the modeled maximum burial depth (Fig. 22).

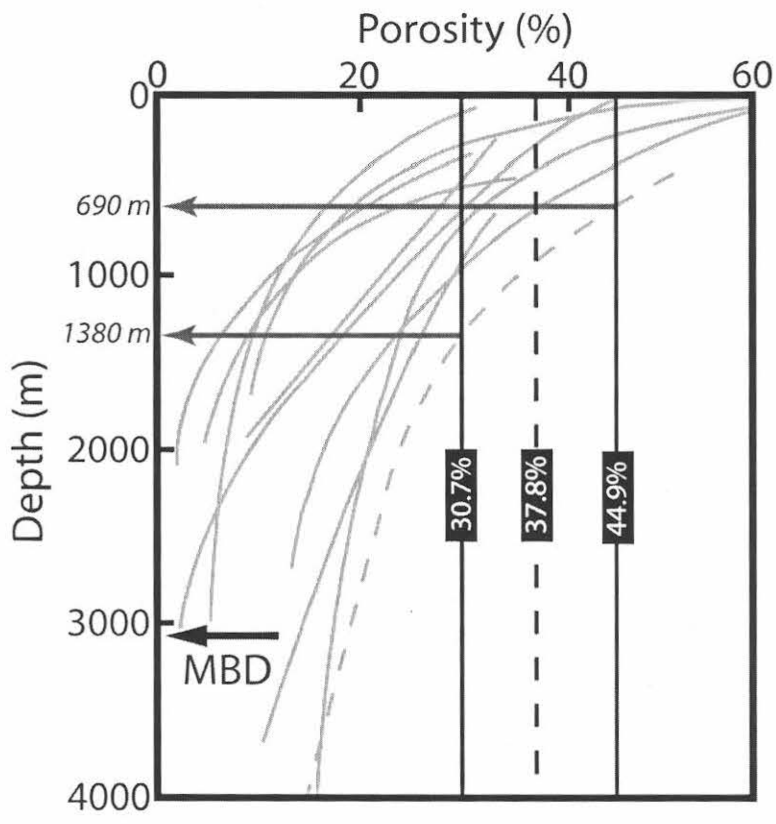


Fig. 22: Porosity versus depth of burial relationships for shale and clayey sediments (modified after Rieke and Chilingarian, 1974) showing the estimated maximum Rhinestreet paleo-fluid retention depth (PFRD) range based on the mean and the upper and lower standard deviation ranges for the calculated paleoporosity of the Rhinestreet shale. The dashed normal compaction trend, which was used to define the maximum depth of the PFRD, is from Ham (1966). MBD = modeled maximum burial depth of the Rhinestreet shale.

Our second approach to estimating the PFRD of the Rhinestreet shale involves the use of some of the more widely cited empirically derived porosity-depth algorithms for normally compacted shale (Gallagher, 1989). The majority of these functions describe a reduction in porosity that varies exponentially with depth, following, to various degrees, the relationship proposed by Athy (1930). The Rhinestreet shale PFRD was estimated using the shale exponential functions of Hansen (1996; Norwegian Shelf), Huang and Gradstein (1990; deep sea claystone), Tingay et al. (2000; Baram Basin, Brunei Darussalam), Hermanrud et al. (1998; offshore mid-Norway), Sclater and Christie (1980; North Sea), Gallagher and Lambeck (1989; Eromanga Basin, Australia), and Hegarty et al. (1988; southern margin of Australia). Estimated depths at which a porosity of 37.8% is reached during equilibrium compaction range from a low of 456 m based on the porosity-depth function of Hegarty et al. (1988) to 1,543 m, half the maximum burial depth of the Rhinestreet shale, using the algorithm of Tingay et al. (2000) (Fig. 23). Most calculated depths derived from exponential functions fall within the depth range defined by comparison of the Rhinestreet ϕ_p with normal compaction curves (Fig. 23). Three linear shale porosity-depth relations were used to calculate the Rhinestreet PFRD; Hansen (1996; Norwegian Shelf), Velde (1996; clay-rich sediments and Recent and Tertiary sediments from Japan and Italy; burial depths > 500 m), and Falvey and Deighton (1982). These functions yield PFRDs of 1,344 m, 622 m, and 500 m, respectively (Fig. 23). The minimum PFRD, 295 m, was obtained from the power-law equation of Baldwin and Butler (1985) for normally compacted shale. Finally, the parabolic porosity-depth function of Liu and Roaldset (1994) placed the PFRD of the Rhinestreet shale at 2,100 m, one km above its modeled maximum burial depth (Fig. 23).

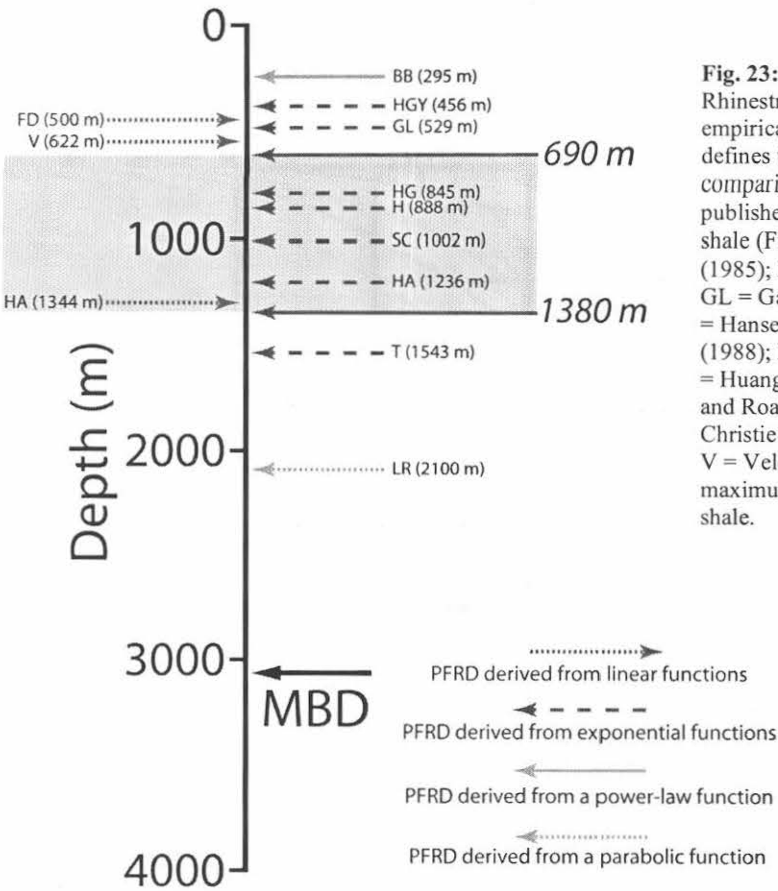


Fig. 23: Calculated PFRD values for the Rhinestreet shale (ϕ_p = 37.8%) using empirical algorithms. Shaded interval defines the PFRD range based on comparison of the Rhinestreet ϕ_p with published normal compaction curves for shale (Fig. 22). BB = Baldwin and Butler (1985); FD = Falvey and Deighton (1982); GL = Gallagher and Lambeck (1989); HA = Hansen (1996); HGY = Hegarty et al. (1988); H = Hermanrud et al. (1998); HG = Huang and Gradstein (1990); LR = Liu and Roaldset (1994); SC = Sclater and Christie (1980); T = Tingay et al. (2000); V = Velde (1996); MBD = modeled maximum burial depth of the Rhinestreet shale.

Normal gravitational mechanical compaction of sediment with a ϕ_o of 70% to > 2 km burial depth typically reduces porosity to ≤ 20% (Rieke and Chilingarian, 1974; Magara, 1978; Giles et al., 1998). Estimation of the Rhinestreet PFRD by use of published porosity-depth curves for normally compacted shale and empirically derived porosity-depth functions suggests that this organic-rich unit departed its normal compaction trend at a depth of 850 to 1,380 m, short of its maximum burial depth (Fig. 23). The relatively shallow depth range of the Rhinestreet PFRD indicates that the overpressure mechanism was not linked to the transformation of kerogen to hydrocarbons, which did occur in the Rhinestreet shale of the LED, but at much greater depth (Lash, 2006c). The most likely candidate for the early generation of overpressure in the Rhinestreet shale is disequilibrium compaction. Disequilibrium compaction, while capable of pushing P_p above hydrostatic at a given depth, the fluid retention depth, cannot drive it high enough to create an effective tensile stress of zero as the increase in P_p is linked inextricably to the rate of loading or the lithostatic gradient (see Fig. 21; Hart et al., 1995; Kooi, 1997; Swarbrick et al., 2002). Indeed, this phase of overpressuring in the shale-dominated succession of the LED did not result in the propagation of NHFs.

There is widespread agreement that the principal controls of disequilibrium compaction include sedimentation (loading) rate and sediment permeability, and, secondarily, sediment compressibility (Luo and Vasseur, 1992; Kooi, 1997; Swarbrick et al., 2002; among others). Swarbrick et al. (2002) argue convincingly that the fluid retention depth in low permeability/highly compressive rocks, like the organic-rich Rhinestreet shale, would be shallower than that expected for more permeable, less compressible sand- and/or silt-rich deposits. Further, given the same sediment type, the fluid retention depth will be relatively shallow in those rocks that have been affected by high sedimentation rates. For example, Grauls (1998) points out that a slow sedimentation rate (~50 m Ma^{-1}) in offshore Angola resulted in fluid retention at a depth of 1,200 m; on the Nile Delta, on the other hand, where sedimentation rates > 800 m Ma^{-1}, disequilibrium compaction begins at a depth of only 800 m (Mann and Mackenzie, 1990).

In order to better evaluate the timing of disequilibrium compaction in the Upper Devonian shale succession of the LED, we need to fully understand the Upper Devonian-Mississippian stratigraphy of this region of the basin (Fig. 24). Exhumation caused by passage of a peripheral bulge at the onset of the Alleghanian orogeny across western New York state (Faill, 1997, and discussed below) resulted in the loss of much of the Mississippian section in the LED (Fig. 24; Lindberg, 1985). However, the presence of the lower Kinderhookian Knapp conglomerate conformably overlying Devonian strata along the New York state-Pennsylvania border immediately south of the LED suggests that the Cashaqua shale is overlain by 1,155 m of Upper Devonian deposits (Fig. 24). Approximately 148 m of Frasnian strata, including the Rhinestreet shale, were deposited on top of the Cashaqua shale at a rate of 30 m Ma^{-1}. The remaining 1,007 m of Upper Devonian sediment accumulated during the Famennian stage at an average rate of 65 m Ma^{-1}, a more than two-fold increase in sedimentation rate from the Frasnian stage. Moreover, the sedimentology of the lower part of the Famennian succession in the LED, the Dunkirk black shale and overlying Gowanda gray shale (see Fig. 3), suggests that the relatively low Frasnian sedimentation rate continued into the Famennian (Baird and Lash, 1990). However, overlying progressively siltier deposits that record a regressive shift from delta front to shallow-marine shelf environments (Baird and Lash, 1990; Dodge, 1992) evince an increase in sedimentation rate in post-Gowanda shale time. Accumulation of the 275-m-thick Dunkirk shale-Gowanda shale succession (Lindberg, 1985) at the above-cited 30 m Ma^{-1} Frasnian sedimentation rate would have encompassed 9.2 Ma of the 15.4 Ma Famennian stage (Kaufmann, 2006). The remaining 732 m of Famennian deposits, then, would have accumulated at the markedly higher rate of 118 m Ma^{-1}.

The transition from the Devonian to the Mississippian in the Appalachian Basin of northwest Pennsylvania is defined by a marked drop in sedimentation rate to 25 m Ma^{-1} based on 172 m of Kinderhookian deposits preserved in this area (Fig. 24; Lindberg, 1985). However, if the Knapp conglomerate accumulated at the very end of the Devonian, as suggested by Dodge (1992), the postulated drop in sedimentation rate would have occurred shortly before the Mississippian. Regardless, the sedimentation rate appears to have diminished throughout the Mississippian (e.g., Beaumont et al., 1987) as evidenced by a calculated sedimentation rate of 9 m Ma^{-1} for the 366 m (Lindberg, 1985) of Kinderhookian to Chesterian deposits preserved in north-central Pennsylvania, immediately south of the Finger Lakes District of western New York state (Fig. 24). The marked reduction in sedimentation rate at the Devonian-Mississippian boundary in Finger Lakes and Lake Erie districts appears to confirm Bond and Kominz's (1991) contention that the Early Carboniferous of the central Appalachians lacked significant tectonic activity.

We believe that the early and relatively shallow onset of disequilibrium compaction in the low-permeability shale-dominated Upper Devonian succession of the LED was induced by the marked increase in sedimentation rate in the latter half the Famennian related to an acceleration, perhaps glacio-eustatic in origin (Veevers and Powell, 1987), in the rate of progradation of the Catskill Delta Complex. Moreover,

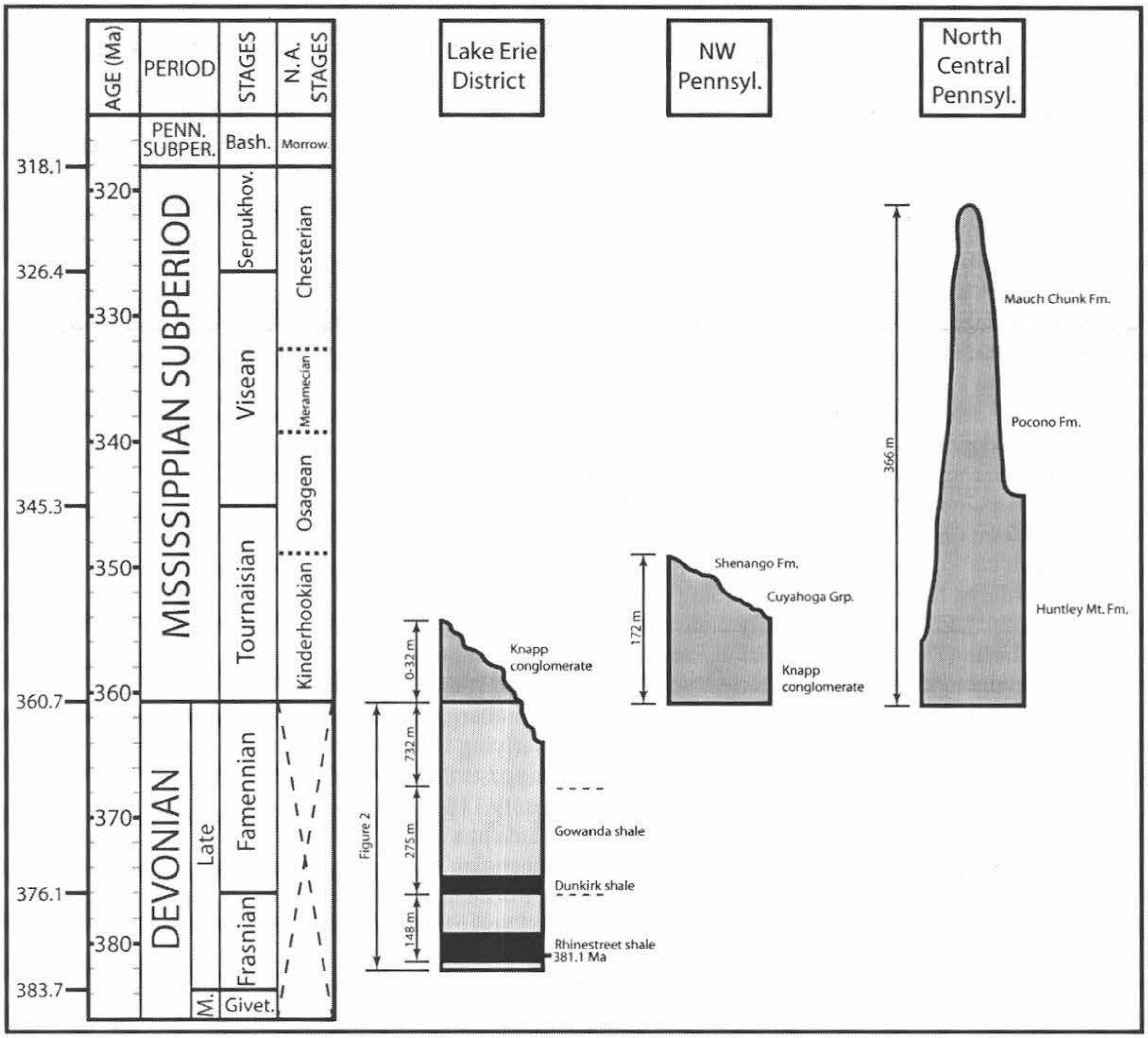


Fig. 24: Upper Devonian-Mississippian stratigraphic relations of western New York state (Lake Erie District) and northwest and north-central Pennsylvania. Stratigraphic columns, including thicknesses, are modified from Lindberg (1985). Devonian time scale (including the Devonian-Mississippian boundary) is from Kaufmann (2006); Mississippian time scale is from Gradstein et al. (2004).

the presence of a 1,100 to 1,200 m of Devonian strata on the Cashaqua shale-Rhinestreet shale contact suggests that the onset of disequilibrium compaction in the Upper Devonian succession occurred at a depth of ~1.1 km, well within the depth range defined by comparison of the Rhinestreet shale ϕ_p with normal compaction curves and porosity-depth algorithms for shale (see Fig. 23). The earliest overpressure episode recorded by rocks of the LED, then, is an Acadian event that finds recent analogues in the Louisiana Gulf Coast (Dickinson 1953) and the Mahakam Delta, Indonesia (Burrus, 1998). Indeed, our estimated PFRD of the Rhinestreet shale and the maximum Famennian sedimentation rate of 118 m Ma^{-1} are consistent with data from modern basins overpressured by disequilibrium compaction (Fig. 25; Swarbrick et al., 2002).

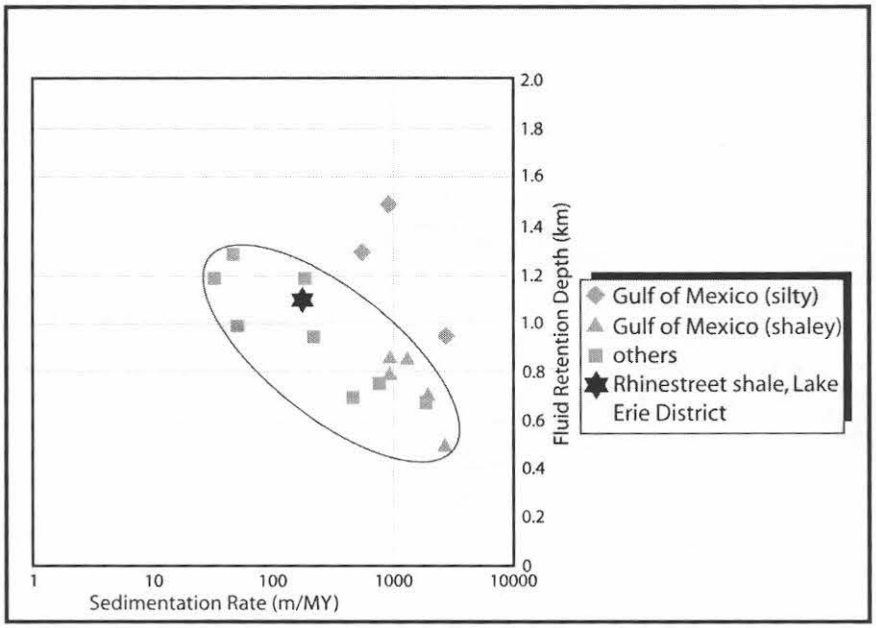


Fig. 25: Plot of sedimentation rate versus fluid retention depth showing data from the Gulf of Mexico and other basins and the Rhinestreet shale (modified after Swarbrick et al., 2002). Refer to text for discussion of the sedimentation rate that induced disequilibrium compaction of the Rhinestreet shale and estimation of the Rhinestreet PFRD.

ONSET OF THE ALLEGHANIAN OROGENY - FRACTURE EVIDENCE FOR PASSAGE OF A PERIPHERAL BULGE

The Rhinestreet shale was overpressured as a consequence of the Acadian orogeny, yet the nature of the overpressuring mechanism – disequilibrium compaction – could not have driven P_p high enough to induce NHFs. Thus, jointing of the Upper Devonian shale-dominated sequence did not occur until after the Acadian orogeny. The following discussion emphasizes joint-driving mechanisms; thus, this is an appropriate time to introduce the dominant joint sets represented in the Rhinestreet black shale and other shale units of the LED. It is worth noting that even relatively recent publications relate the pervasive fractures carried in Upper Devonian black shales of western New York state and western Pennsylvania to glacial unloading and differential isostatic rebound (e.g., Schmoker and Oscarson, 1995). Although we have catalogued joints that formed under these conditions (Lash et al., 2004), the systematic joints readily observed in shale exposed along the lakeshore formed much earlier and under very different loading configurations. Indeed, the pioneering work of Terry Engelder and his students at Penn State, grounded in linear elastic fracture mechanics, has demonstrated the variety of natural loading configurations and related joint-driving mechanisms under which rocks can fracture (Engelder, 1985, 1987; Engelder and Lacazette, 1990; Engelder and Gross, 1993; Engelder and Fischer, 1996; Lacazette and Engelder, 1992). Moreover, each loading configuration may be represented during a single tectonic cycle (Engelder, 1985).

The Rhinestreet shale carries as many as five systematic joint sets (e.g., Lash et al., 2004). The most pervasive sets, regionally and stratigraphically, are NW (298°-313°)- and ENE (060°-075°)-oriented sets (Fig. 26). Systematic NS (352°-007°)-trending joints, while displaying a strong stratigraphic control as described below, are also recognized over much of the LED (Fig. 27). Abutting relations among joints of the three sets indicate that the NS joints are the oldest structures and systematic ENE-trending joints are the youngest (Lash et al., 2004). Both the ENE- and NW-trending joints show a strong affinity for black shale, especially the basal, more organic-rich intervals of these units (Lash et al., 2004). Joints of these two sets seldom extend downward from the bases of black shale units into underlying gray shale; instead, they appear to have propagated farther upward into the black shale and then into overlying gray shale (Fig. 28; Lash et al., 2004).

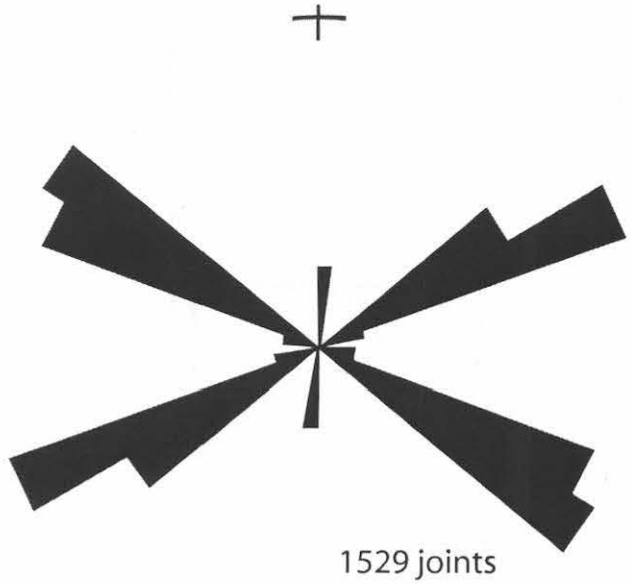


Fig. 26: Rose plot of joints in the Rhinestreet shale, northern Erie County, New York.

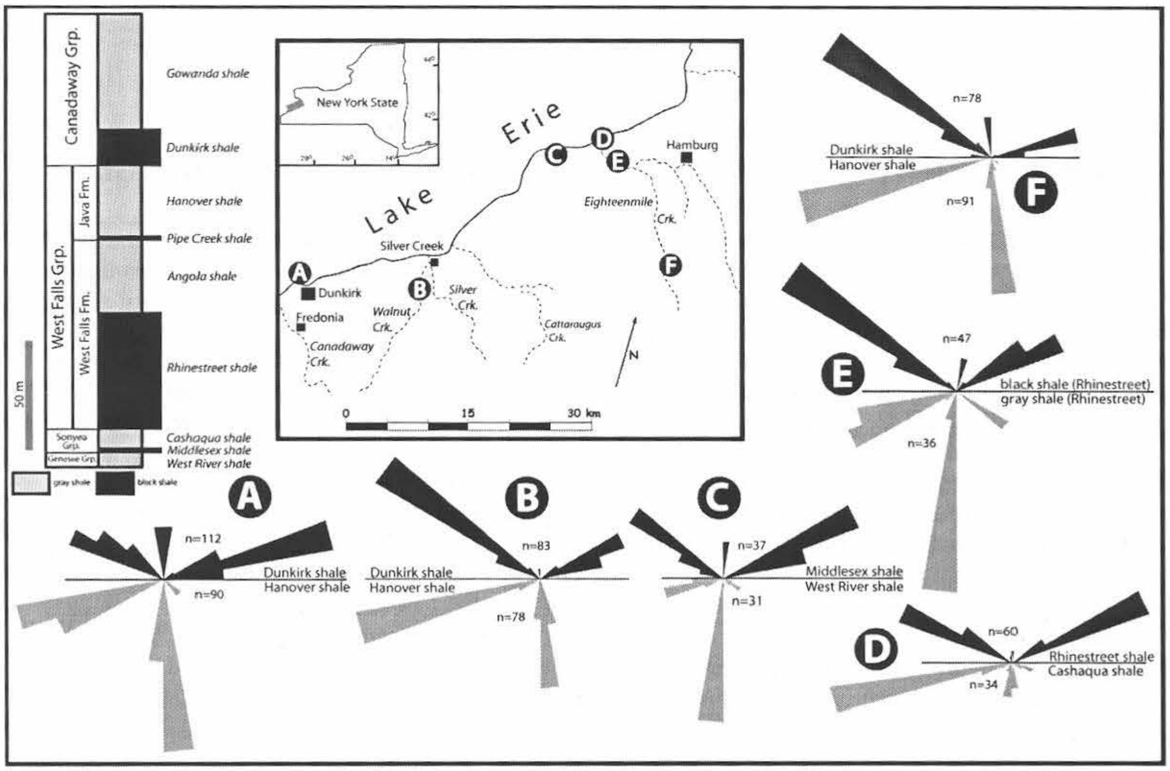
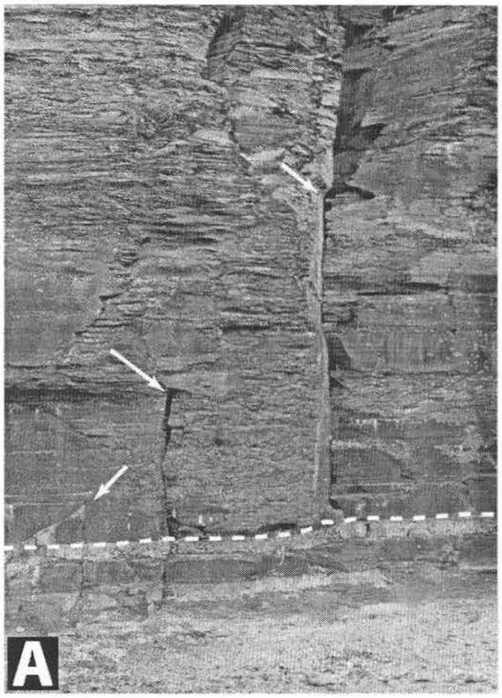


Fig. 27: Generalized stratigraphic column showing the Upper Devonian shale succession of the Lake Erie District and selected rose plots of systematic NS, NW, and ENE joints at gray shale-black shale contacts throughout the Lake Erie District. The lettered rose plots are keyed to the location map.

Fig. 28: Tall, close-spaced NW joints at the contact (dotted line) of the Dunkirk shale and underlying Hanover gray shale along Eighteenmile Creek.

Systematic NS joints formed preferentially at the tops of gray shale units and basal intervals of overlying black shale deposits (Fig. 27; Lash, 2006b). These joints have been observed at the bottom contacts of all Upper Devonian black shale units of the LED and at almost all exposures of these contacts (Fig. 27). However, they are most densely developed (i.e., lowest median orthogonal spacing) at the Hanover shale-Dunkirk shale contact, the Cashaqua shale-Rhinestreet shale contact, and at contacts of gray shale intervals and overlying black shale within the Rhinestreet shale (Fig. 27). Few NS joints observed in gray shale (~20%) extend more than 20 cm into overlying black shale (Fig. 29A); roughly 50-60% of all NS joints examined by us either were arrested below or at the base of the overlying black shale unit (Fig. 29B). Finally, NS joints are rarely found exclusively in black shale; indeed, >90% of NS joints observed in black shale units can be readily traced into underlying gray shale from which they appear to have propagated.

The joint-driving mechanism(s) for the three dominant systematic joint sets carried in Upper Devonian shale of the LED is not immediately obvious. Few joint surfaces of these fine-grained rocks contain any sort of ornamentation that might be informative regarding the origin of the joints. However, rare plumose structures consistent with incremental propagation decorate ENE and NW joint surfaces. Such morphology is typical of fluid loading during natural hydraulic fracturing (Lacazette and Engelder, 1992). Moreover, the close orthogonal spacing of joints of each set relative to their height offers further evidence for their propagation under conditions of fluid loading (Ladeira and Price, 1981; Fischer et al., 1995; Engelder and Fischer, 1996). Indeed, it is tempting to attribute the seemingly similar nature of the NS joints to the same loading configuration as the NW and ENE joints. However, further insight into differences in joint-driving mechanisms among the three dominant joint sets carried in the LED may be found in the nature of joint-carbonate concretion interactions (e.g., McConaughy and Engelder, 1999). Studied diagenetic carbonate concretions in the Upper Devonian shale succession of the LED exist in two general forms that appear to be controlled by host shale type: (1) large ellipsoidal (aspect ratio ~ 2) internally laminated concretions most common to black shale units as described above (Lash and Blood, 2004a,b) and (2) smaller lenticular (aspect ratio > 4) internally massive concretionary bodies most frequently found in bioturbated gray shale.



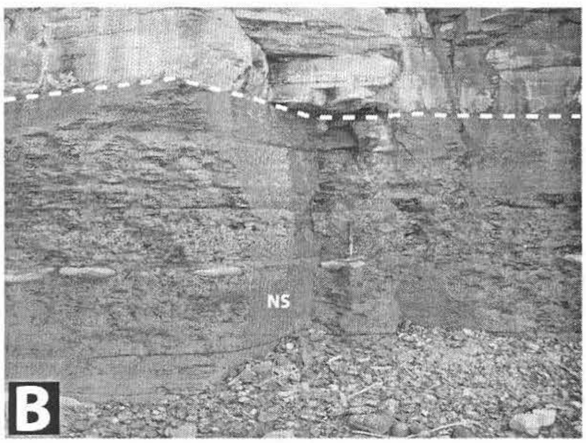


Fig. 29: NS joints terminating at gray shale-black shale contacts: (A) close-spaced NS joints at the Hanover shale-Dunkirk shale contact (dashed line). Arrows point to tips of NS joints that have propagated from the Hanover shale into the Dunkirk shale; (B) NS joint (**NS**) arrested at the contact of a gray shale (below white dashed line) interval and overlying black shale within the Rhinestreet shale. Note concretionary carbonate horizon in the gray shale at the hammer.

Systematic NW and ENE joints show no deflection in propagation path as they approach concretions; instead, the joints propagated in-plane above and below the concretions (Fig. 30A). However, the occasional joints confined to a mechanical layer narrower than the width of the concretion define propagation paths that deflected into an orientation approximately normal to the interface. The majority of systematic NW and ENE joints (>90%) failed to penetrate concretions (Fig. 30B); one in ten ENE joints were observed to have propagated no more than a centimeter into concretions before arresting.

Systematic NS joints, while grossly similar to NW and ENE joints, differ from their younger counterparts in the nature of their interactions with embedded concretions. Anywhere from 30% to 70% of NS-trending joints, depending upon one's field location, completely cleave lenticular concretions (Fig. 31A). Few NS joints interface with larger ellipsoidal concretions common to black shale. However, the infrequent NS joints that propagated upward from gray shale into concretion-bearing interval within black shale typically penetrate deeply (> 4 cm) into the concretions (Fig. 31B). Occasionally, NS joints in gray shale can be observed cutting concretions in one stratigraphic interval but are arrested at interfaces with concretions in immediately over-and/or under-lying horizons (Fig. 31C). The few NS joints whose width is less than that of the concretions with which they interact display propagation paths that curve toward interfaces (Fig. 31D).

NW- and ENE-trending joints of the LED are NHFs that propagated as a consequence of the transformation of kerogen in the organic-rich black shale to hydrocarbons at burial depths >2.3 km (Lash et al., 2004; Lash and Engelder, 2005), an interpretation confirmed by their interactions with embedded concretions. Specifically, P_p in black shale was driven well above hydrostatic by catagenesis and sustained by a tight, strongly anisotropic microfabric (e.g., Lash and Engelder, 2005). Increasing P_p ultimately reduced the compressive minimum horizontal stress to an effective tensile stress initiating the NW- and later ENE-trending fluid loaded joints. The relatively stiff (i.e., high modulus) concretions embedded in the overpressured black shale, however, remained under effective compressive stress thereby suppressing penetration by NHFs beyond the carapace of the concretion (McConaughy and Engelder, 1999).

The critical difference between systematic NW and ENE joints and the older NS joints is that the latter penetrate deeply and/or completely cut many, but by no means all, concretions. We suggest that these joints originated within the higher modulus concretionary carbonate by a thermoelastic contraction driving mechanism. This scenario requires a uniform level of extensional elastic strain distributed over the entire Upper Devonian shale succession of the LED. Presuming elastic extension, the higher modulus diagenetic carbonate would fail in tension before the lower modulus shale, which remained under effective compression at a given depth of burial (e.g., McConaughy and Engelder, 1999). This interpretation is borne out by the local presence of NS joints confined to lenticular concretionary carbonate at the tops of the gray shale units (Fig. 32). However, the fact that not all concretions originated NS joints may reflect variable $CaCO_3$ abundance among concretions and its effect on elastic properties. The lack of NS joints originating in the ellipsoidal (lower aspect ratio) concretions in black shale suggests that tensile stress builds most effectively in the higher aspect ratio concretionary carbonate.

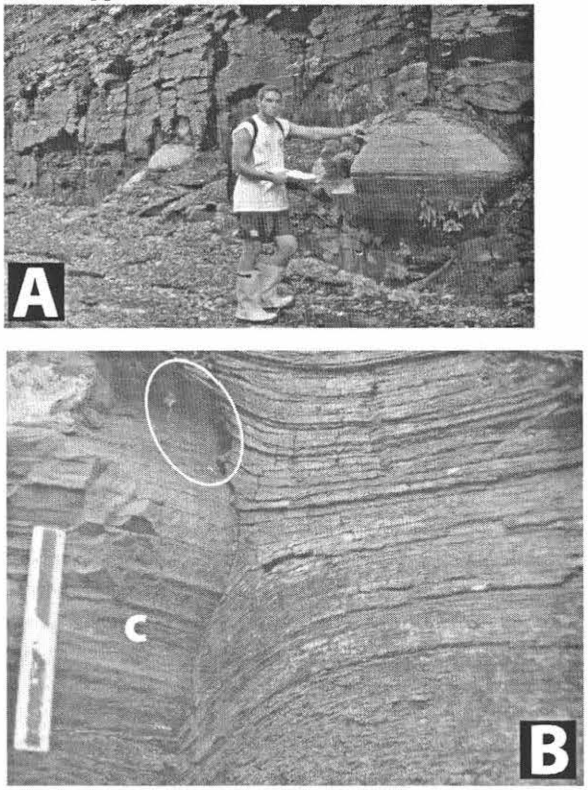


Fig. 30: (A) Tall ENE joint that propagated in-plane around two large concretions in the Rhinestreet shale. Person = 1.9 m; (B) Close-up of the interface of a concretion (**c**) and a NW joint that propagated in-plane around the concretion. Note the lack of penetration in the circled area where the black shale has been eroded. Scale = 14 cm.

Propagation of NS joints out of concretions and into the lower modulus host gray shale was likely aided by modestly overpressured pore fluid in the latter that migrated into the open joints. Elevated P_p within newly formed joints in the diagenetic carbonate would have enabled them to propagate through the shale under a fluid decompression mechanism. Indeed, the preferential growth of NS joints at the tops of the gray shale units probably reflects some degree of overpressure at these stratigraphic intervals. Deeper in the gray shale, however, lower (near hydrostatic) P_p would not have been high enough to create an effective tensile stress necessary for joint propagation. Thus, whereas NS joints were initiated in higher modulus diagenetic carbonate by thermoelastic contraction, they were driven through the lower modulus shale by fluid decompression. Evidence for this interpretation includes NS joints that cut concretions at one stratigraphic level (point of initiation) but failed to cleave or even penetrate concretions in immediately over- and under-lying stratigraphic intervals. Instead, the joints propagated in-plane around the rigid inclusions (Fig. 31C). Further, the curving of some NS joint trajectories toward normal to concretion interfaces (Fig. 31D) is more typical of fluid loaded joints than joints that originated by thermoelastic contraction (McConaughy and Engelder, 1999).

The switch in joint-driving mechanism recorded by the Rhinestreet shale from thermoelastic contraction supplemented by fluid decompression early in the deformation history of these deposits (*NS joints*) to a solely fluid decompression mechanism later in the tectonic cycle (*NW* and *ENE joints*) must be considered in terms of our present understanding of Alleghanian tectonics. The inception of the Alleghanian orogeny in the central Appalachian Basin is nominally dated by the initial accumulation of the Chesterian Mauch Chunk delta deposits shed from an eastern source area over the Greenbrier and Loyalhanna limestones (Hatcher et al., 1989; Faill, 1997). The end of the Mississippian, however, witnessed a change in sedimentary facies from the low-energy Mauch Chunk facies to westward spreading high-energy fluvial conglomeratic deposits of the Pottsville Formation (Meckel, 1967; Colton, 1970). This transition has been attributed to renewed uplift caused by crustal thickening during oblique convergence between the African portion of Gondwana and Laurentia, perhaps with microcontinents such as the Goochland terrane sheared between them (Faill, 1997). Oblique convergence early in the Alleghanian orogeny (Atokan and younger) is manifested by sequential dextral slip deformation along much of the central and southern Appalachian internides and development of ENE-trending (present direction) face cleat in coal in the central and southern Appalachians and ENE-trending joints in Frasnian black shales of the Appalachian Plateau of New York state (Fig. 33; Gates and Glover, 1989; Hatcher, 2002; Engelder and Whitaker, 2006).

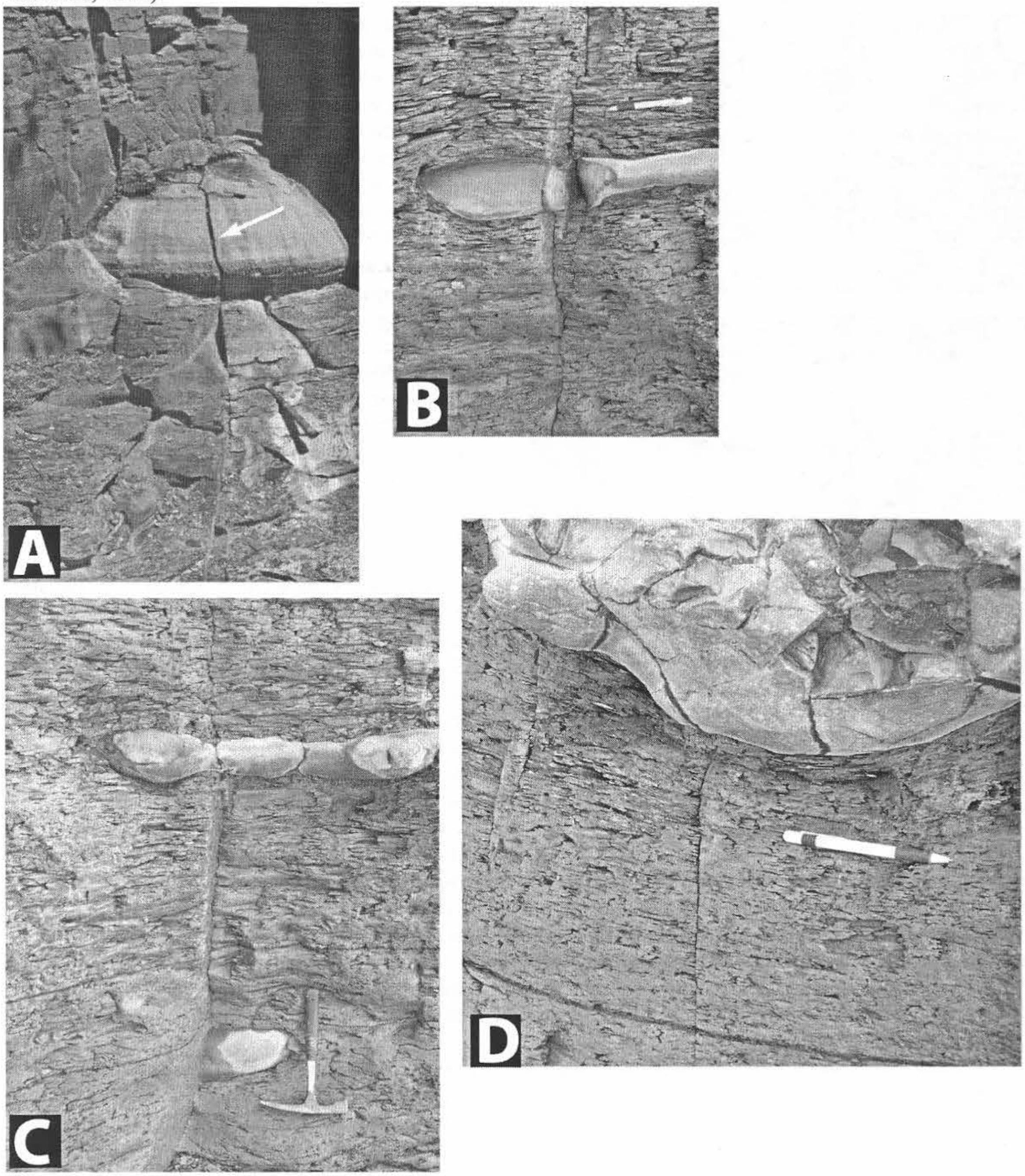


Fig. 31: (A) NS joint that cut a concretion in a gray shale interval of the Rhinestreet shale. Pen = 11 cm; (B) NS joint (indicated by arrow) that penetrated a large carbonate concretion in the Rhinestreet shale. An ENE joint (**ENE**) propagated in-plane (parallel to the picture) around the concretion. Note hammer for scale; (C) NS joint that cut the upper concretionary carbonate body but propagated in-plane around the lower concretion; (D) the propagation path of a narrow (relative to the width of the concretion) NS joint deflecting into a path perpendicular to the concretion interface. Pen = 11 cm.

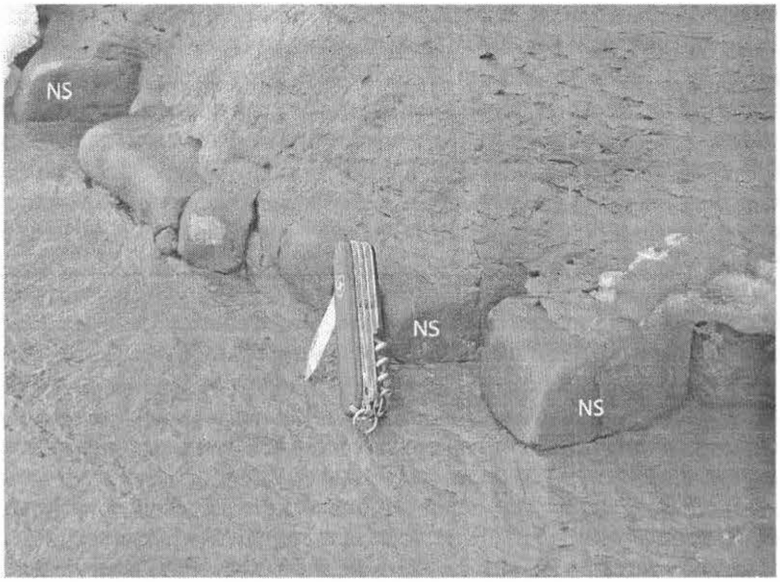


Fig. 32: NS joints (**NS**) that failed to propagate out of a thin carbonate layer at the top of the Hanover shale. Knife = 8 cm long.

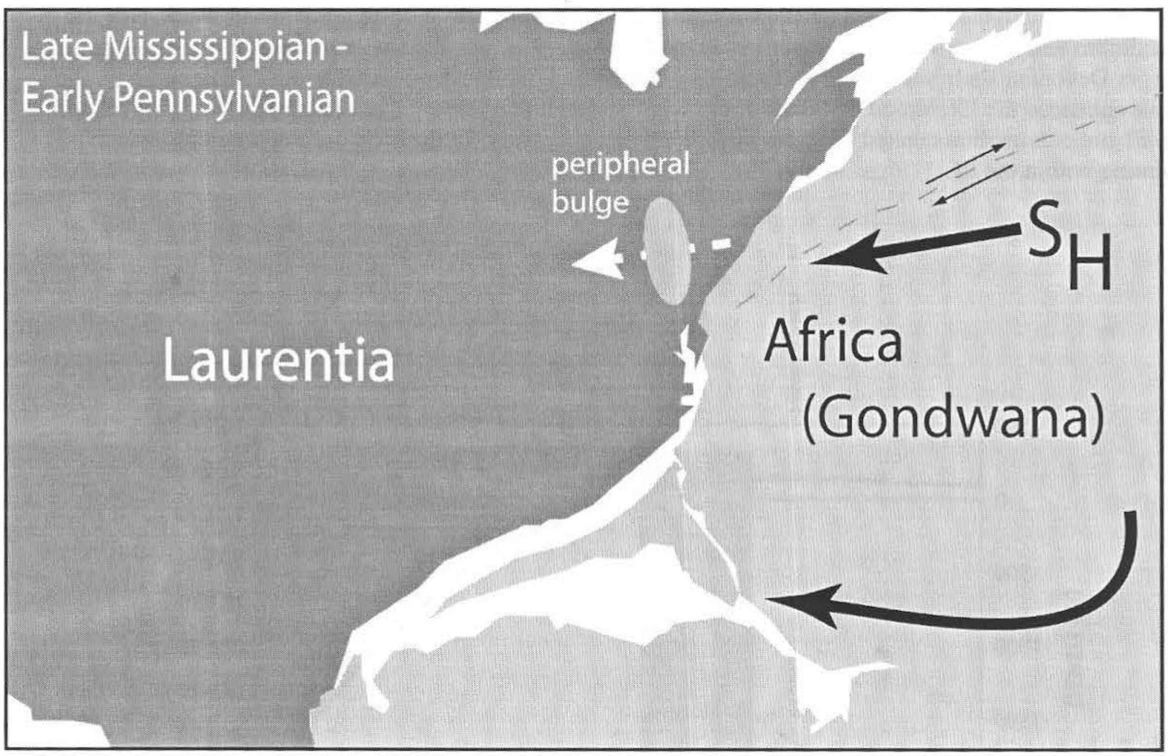


Fig. 33: Closure of the Theic ocean early in the Alleghanian orogeny by oblique convergence of Gondwana against Laurentia (modified after Hatcher, 2002). Refer to text for discussion.

The onset of Alleghanian tectonics is also manifested by a widespread Morrowan unconformity present in parts of the Appalachian system, including the southern tier of western New York state (e.g., Berg et al., 1983; Lindberg, 1985) (see Fig. 24). This unconformity is attributed to the cratonward migration of a peripheral bulge caused by early Alleghanian thrust loading in the southern Appalachians (Ettensohn and Chesnut, 1989). Similarly, loading during oblique convergence of Gondwana against Laurentia in New England and the Canadian Maritimes Appalachians drove a contemporary peripheral bulge across New York state and Pennsylvania (Fig. 33; Faill, 1997). The switch in joint-driving mechanism demonstrated by the jointing history of the Rhinestreet shale is consistent with passage of a peripheral bulge across the basin followed by burial of the shale succession to the oil window. The peripheral bulge was driven by crustal loading during the earliest stages of closure of Africa against the Laurentian plate in New England in Morrowan time (Fig. 33; Faill, 1997). The propagation of NS joints in gray shale of the LED prior to the formation of ENE-trending coal cleat elsewhere in the Applachian Basin (Engelder and Whitaker, 2006) and NW joints in the Rhinestreet shale indicates that the Upper Devonian shale succession of the LED had not yet been buried to the oil window by the time of the Morrowan unconformity, consistent with our burial/thermal model of the Rhinestreet shale (see Fig. 9). Indeed, the earliest coal cleat in the Appalachian Basin, which records oblique convergence in the New England Appalachians, post-dates the Morrowan (Engelder and Whitaker, 2006). Thus, the NS-trending joints in the LED are among the earliest structures that can be attributed to the Alleghanian orogeny.

Modeling of the peripheral bulge as having formed in response to the imposition of a line load at the edge of Laurentia and an overthrust fill of 2 km (Stewart and Watts, 1997) predicts that flexural curvature was highest in the approximate location of the Hudson Valley (Fig. 34). However, the great depth of burial and consequent high compressive stress in this more proximal area of the basin suppressed curvature-related tensile stress keeping the rocks under effective compression. Flexure-related uplift resulted in deep erosion in the LED and to the south in western Pennsylvania (Berg et al., 1983; Lindberg, 1985). Although the calculated tensile stress in this region of the Appalachian Basin was ~ one-half the maximum tensile stress magnitude generated by the bulge (Fig. 34), the shallow depth of burial of the Upper Devonian shale succession and resulting diminished overburden-related compressive stress would have enhanced the likelihood that these rocks were placed under effective tensile stress. Flexural-related uplift and exhumation created the near-surface tension necessary for thermoelastic contraction-driven jointing within the LED (Figs. 34 and 35).

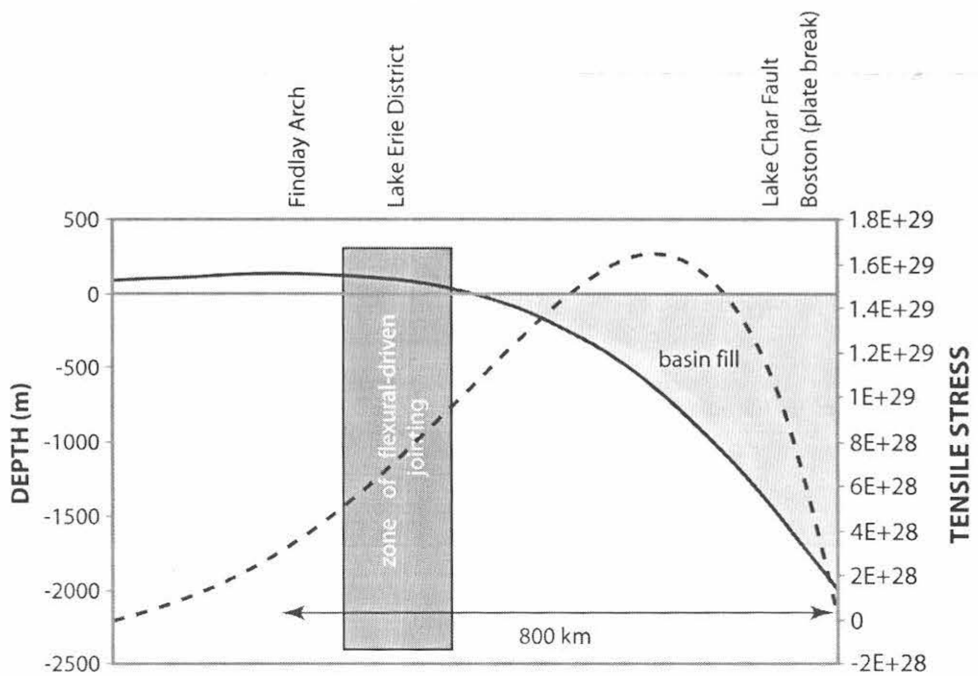


Fig. 34: Modeled peripheral bulge showing variations in the calculated tensile stress across the Appalachian Basin.

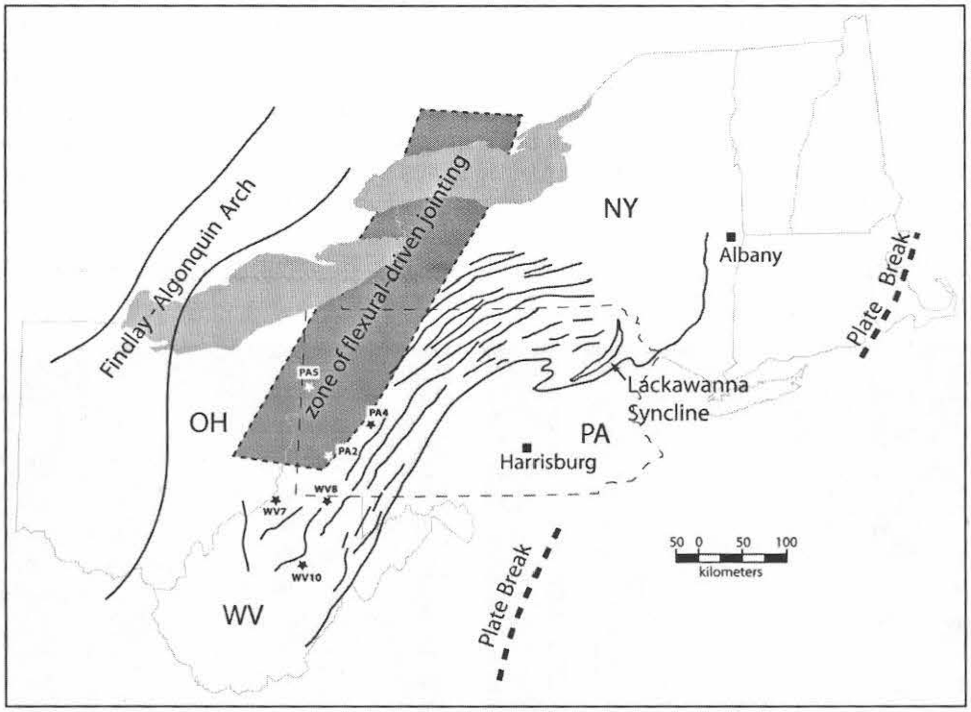


Fig. 35: Map showing the modeled area of NS jointing related to migration of the peripheral bulge. Black stars show locations of EGSP (Eastern Gas Shales Project) cores in which NS joints were observed in Middle Devonian rocks by Evans (1994).

NS joints initiated in higher modulus diagenetic carbonate by thermoelastic loading propagated through the host gray shale aided by modestly elevated P_p, which maintained an effective tensile stress in the lower modulus shale. Preferential propagation of NS joints by fluid decompression immediately beneath the black shale units suggests that the gray shale at these stratigraphic intervals was overpressured. Elevated P_p within the gray shale probably was caused by the increase in sedimentation rate at the end of the Devonian that attended rapid progradation of the Catskill Delta plain into the western New York region of the basin as described above. Overpressure generated by disequilibrium compaction may have been maintained by the formation of gas capillary seals induced by biogenic methanogenesis in the organic-rich Rhinestreet deposits (Blood and Lash, 2006; Lash, 2006b). Termination of the NS joints at the base of, or a short distance within, the black shale may reflect a somewhat lower modulus in the organic-rich deposits that suppressed the tensile stress carried by these rocks.

The proposed influence of the early Alleghanian peripheral bulge on the jointing history of the LED appears to have extended well to the south. The modeled region of the Appalachian Basin where uplift related to lithospheric flexure induced the systematic NS joints can be traced south into western and southwestern Pennsylvania where Evans (1994) described early NS-trending joints in Middle Devonian shale cores (Fig. 35). The locus of NS jointing in the Appalachian Basin was limited to the west by diminishing tensile stress; its eastern extent likely was controlled by the increasing depth of burial and consequent rise in compressive stress (Fig. 34).

MAIN STAGE JOINTING IN THE ALLEGHANIAN – EVOLUTION OF THE RHINESTREET FRACTURED RESERVOIR

The peripheral bulge had subsided by Atokan time when the Pottsville-equivalent Olean Conglomerate accumulated unconformably over eroded Upper Devonian and Kinderhookian strata (Edmunds et al., 1979; Berg et al., 1983; Lindberg, 1985; Dodge, 1992). Continued subsidence through the balance of the Pennsylvanian into Permian time carried the Upper Devonian shale succession of the LED into the oil window by ~ 275 Ma (Fig. 9; Lash, 2006c). The initial response to the Rhinestreet's entry into the oil window appears to have been the initiation of horizontal microcracks identical to those identified in the Upper Devonian Dunkirk shale (Lash and Engelder, 2005; Fig. 36). The intriguing aspect of these fractures is that they formed as open-mode cracks under a basinal stress field in which the greatest principal stress was vertical. Lash and Engelder (2005) explained this as a consequence of two factors: (1) a marked compaction-induced layer-perpendicular strength anisotropy and abundant flattened kerogen particles and (2) poroelastic deformation of these low permeability deposits pressurized by conversion of kerogen to bitumen and consequent establishment of a local *in situ* stress field favorable to the propagation of the microcracks in the horizontal plane.

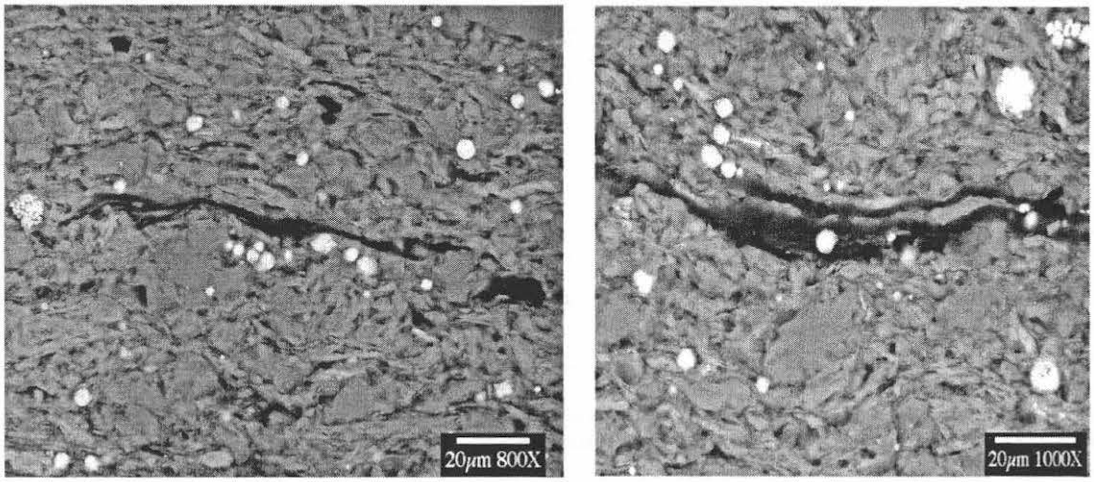


Fig. 36: Secondary electron images of bitumen-filled microcracks in Rhinestreet shale samples.

Continued burial of the Rhinestreet shale and further production of hydrocarbons resulted in the initiation of upward-propagating systematic fluid-loaded NW-trending joints or NHFs. The timing of the entry of these rocks into the oil window ties this phase of jointing to the emplacement of the Blue Ridge-Piedmont Megathrust sheet by rotational transpression of Gondwana along the southern Appalachians (Fig. 37; Hatcher, 2002). The transition from horizontal open-mode fractures to the propagation of vertical NHFs defies simple explanation. There is no evidence that the least principal stress direction flipped from vertical to horizontal. Indeed, in order to avoid violation of Occam's Razor, we assume that the regional stress field remained constant during the time that the horizontal microcracks and NW joints propagated. We propose that the switch from horizontal microcracks produced by catagenesis to vertical NHFs involved some degree of horizontal tensile stress that diminished the effective minimum horizontal stress induced by burial to zero enabling the vertical joints to form (e.g., Zhao and Jacobi, 1997). The origin of the hypothesized tensile stress may be found in the irregular nature of the Laurentian margin (i.e., the Pennsylvania salient; Hatcher, 2002) and its effect on plate edge stresses during rotational transpression of Gondwana along the central and southern Appalachians.

Systematic ENE joints formed toward the end of the Alleghanian tectonic cycle after the Rhinestreet shale had experienced greater hydrocarbon production (see Fig. 9). Further oblique convergence in the New England Appalachians, may have provided a level of lateral extension that, when combined with elevated P_p in the Rhinestreet shale, resulted in the propagation of the ENE joints. The stress field responsible for the ENE joints carried by the Rhinestreet shale had earlier produced ENE-trending systematic joints in Frasnian black shales in the Finger Lakes District as well as an early face cleat in coal deposits throughout the Appalachian system (Engelder and Whitaker, 2006). This stress system, unlike that which produced the NW joints, appears to have remained in place for the duration of the Alleghanian orogeny.

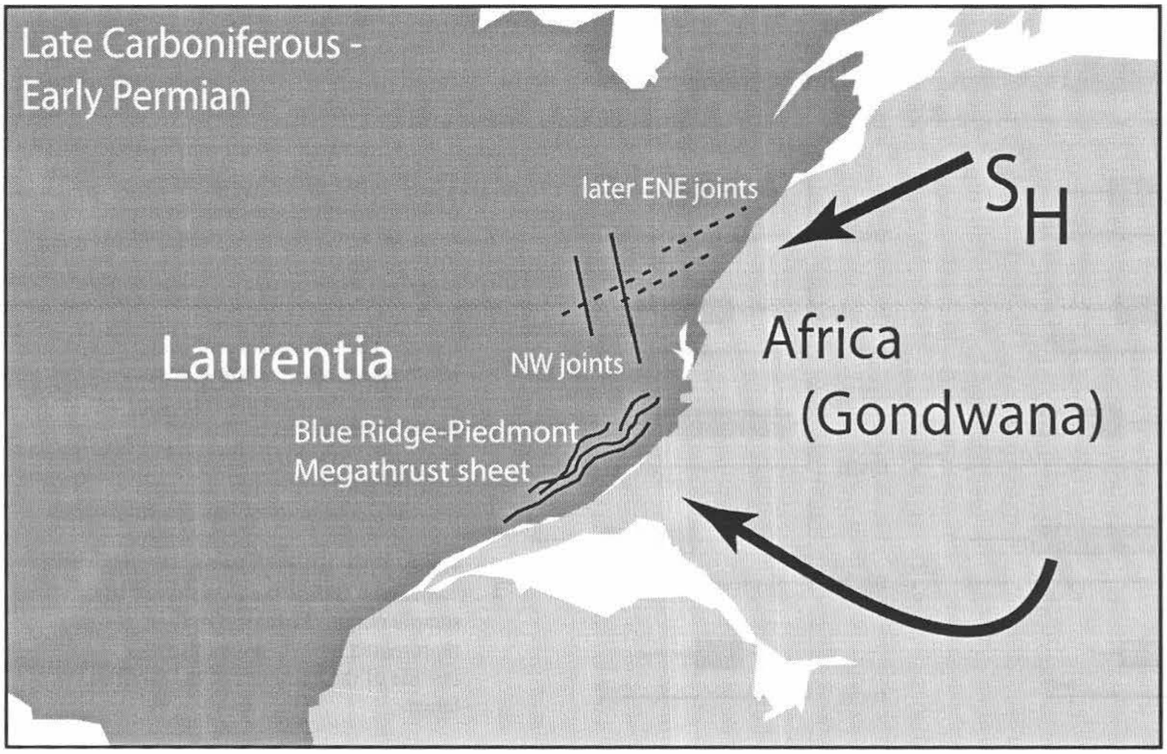


Fig. 37: Late Carboniferous-Early Permian head-on collision resulting in emplacement of the Blue Ridge-Piedmont Megathrust sheet (modified after Hatcher, 2002). Refer to text for discussion.

The Rhinestreet shale continued to undergo thermal maturation even as it was uplifted in post-Alleghanian time (see Fig. 9). ENE joints may have continued to propagate during this time, perhaps induced by post-Alleghanian flexural rebound of the Appalachian Basin (Blackmer et al., 1994). That is, at about the time the Rhinestreet shale had reached its peak thermal maturity (and hydrocarbon generation), unroofing of the Appalachian Plateau brought the overpressured black shale closer to the surface, resulting in thermoelastic contraction. Relaxation of the minimum horizontal compressive stress affecting the overpressured impermeable organic-rich deposits provided the mechanism for the preferential development of effective tensile stress within the black shale relative to gray shale units. Moreover, the ENE orientation of these joints may reflect the Early Cretaceous change in the remote stress system from one dominated by rift-related dynamics to one of compression caused by sea floor spreading of the North Atlantic Ocean (Miller and Duddy, 1989). Indeed, we are increasingly of the mind that there are, in fact, two sets of ENE joints – an older ENE set that was fluid-driven and a younger, less planar, set that may have been induced by thermoelastic contraction. As such, ENE joints in the LED may provide an example of similarly oriented joints having formed by different loading configurations. This intriguing story awaits much more work.

According to our jointing chronology for the Rhinestreet shale (as well as other Upper Devonian black shale units of the LED; Lash et al., 2004), fluid loaded ENE joints formed when the Rhinestreet had been in the oil window for a longer period of time than for formation of the NW joints. Evidence for this interpretation can be found in the relative density of joints of each set as manifested by their spacing characteristics. Joint spacing data for each joint set were obtained by simple scanline techniques (refer to Lash et al., 2004, for details of this method and statistical treatment of the data). Scanline analysis from the top of the Cashaqua shale through the Rhinestreet shale to its contact with the Angola shale reveals three trends regarding the degree of development of NW and ENE joints as measured by orthogonal spacing (Fig. 38). First, and perhaps most obvious, is the great range and high median spacing of ENE joints in the Cashaqua shale immediately below the Rhinestreet shale. NW joints are poorly represented at the top of the Cashaqua, an observation made of other gray shale units in the LED (see Fig. 27). Second, ENE joints

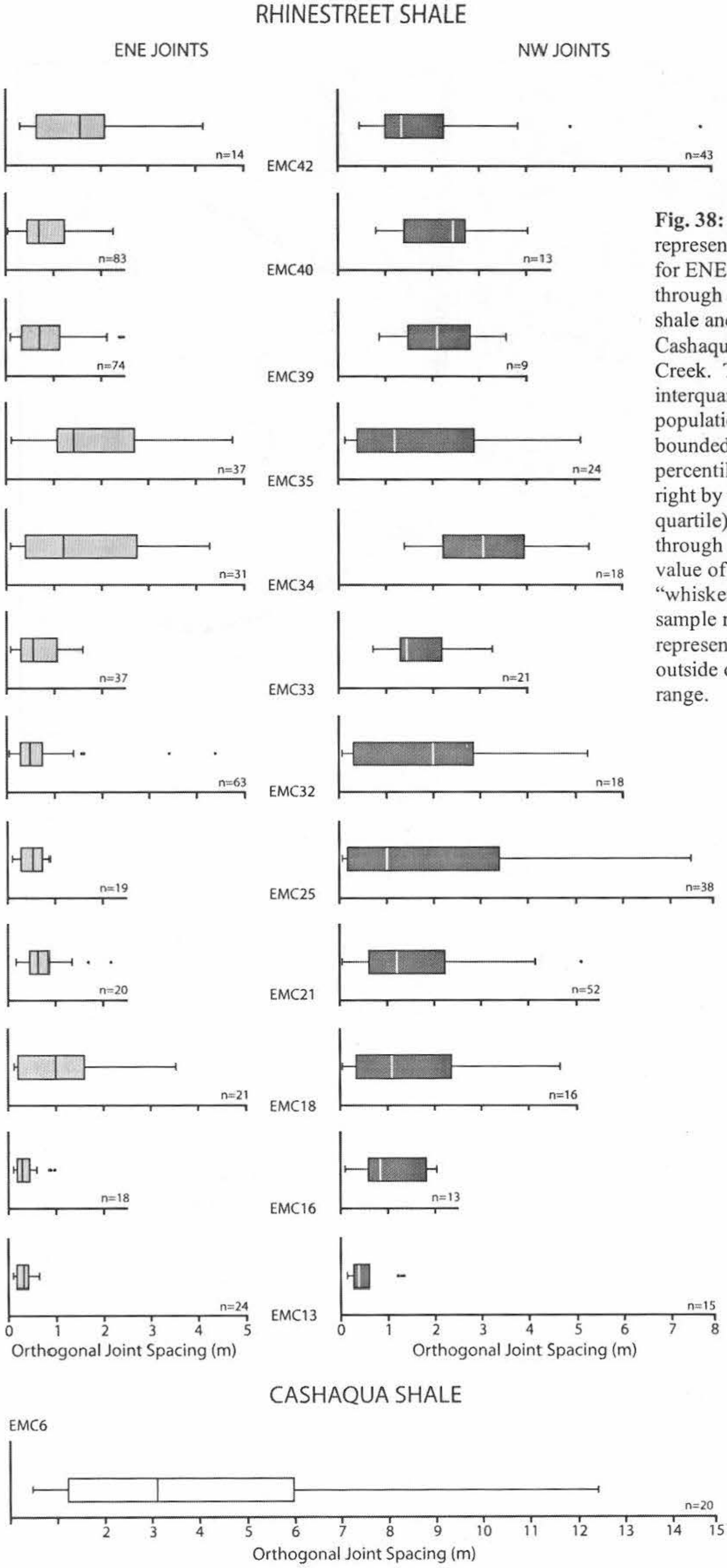


Fig. 38: Box-and-whisker diagrams representing joint spacing distribution for ENE and NW joints from the base through the top of the Rhinestreet shale and ENE joints at the top of the Cashaqua shale along Eighteenmile Creek. The box encloses the interquartile range of the data set population; the interquartile range is bounded on the left by the 25th percentile (lower quartile) and on the right by the 75th percentile (upper quartile). The vertical line drawn through the box defines the median value of the data population, and the "whiskers" define the extremes of the sample range. Statistical outliers are represented by data points that fall outside of the extremes of the sample range.

are more closely and uniformly spaced at a particular stratigraphic horizon than are NW joints at that same horizon. Finally, both NW and ENE joints demonstrate a subtle reduction in density upsection from the high TOC base of the Rhinestreet shale to its contact with the Angola shale (Fig. 38). The scanline data, then indicates that (1) both NW and ENE joints are most densely formed in the most organic-rich deposits, (2) within the Rhinestreet shale, ENE joints are more densely developed than are the older NW joints, and (3) the density of both NW and ENE joints correlate positively with TOC (Fig. 38). Trends similar to these have been described from elsewhere in the Middle and Upper Devonian succession of the Appalachian Plateau (Loewy, 1995; Lash et al., 2004).

Differences in spacing characteristics of the NW and younger ENE joints of the Rhinestreet shale can be thought of in terms of the progressive thermal maturation of a source rock. It is widely held that a set of natural fractures does not form instantaneously; rather, it evolves over a period of time becoming progressively denser (Fig. 39). Rives et al. (1992), for example, have demonstrated that the orthogonal spacing of a systematic joint set evolves from a negative exponential distribution early in its history to a log-normal distribution and finally to a normal distribution. Similarly, Narr (1991) and Becker and Gross (1996) maintain that as joint systems develop, orthogonal spacing becomes progressively smaller and more uniform. Ultimately, overlapping stress shadows between adjacent joints precludes the formation of new fractures in intervening unjointed volumes of rock (Fig. 39). The rock is said to be fracture saturated at this point. Time is a critical factor in this process - the longer the duration of fracturing under a given set of conditions, the closer a specific set approaches saturation. The Rhinestreet shale had just entered the oil window when it was affected by the remote stress field that gave rise to the NW joints. That is, production of hydrocarbons in the Rhinestreet shale had not proceeded long enough when the rocks reacted to the stress field responsible for emplacement of the Blue Ridge-Piedmont Megathrust sheet for the resultant joints to form with a density even close to saturation. Indeed, the wider range in joint spacing of the NW joints is more consistent with a negative exponential or log-normal distribution. Moreover, the general dearth of NW joints at the tops of gray shale units (see Fig. 27) may reflect a level of production in the underlying black shale incapable of driving these joints well up into the overlying gray shale. ENE-trending joints, having formed after the Rhinestreet had occupied the oil window for a prolonged period of time, are more densely and uniformly (close to saturation) developed than NW joints reflecting a greater level of production. Moreover, the great abundance of ENE joints at the tops of gray shale units (see Fig. 27) likely records the higher level of hydrocarbon production by the time the ENE joints were initiated in these rocks. An identical relationship of NW and younger ENE joints has been described from the Dunkirk shale (Lash et al., 2004).

CONCLUSIONS

The Rhinestreet shale is an unconventional or continuous-type hydrocarbon accumulation that satisfies most criteria of these systems, including a lack of obvious trap and seal. Indeed, the Rhinestreet shale, like other Devonian black shale units, is its own seal and reservoir; it is a self-sourced reservoir. However, in spite of general similarities among all continuous-type hydrocarbon accumulations, most differ in terms of more subtle parameters, including mineralogy (e.g., silt content, dominant clay mineral type), complexity of fractures and fracture history, reservoir thickness and internal stratigraphy, compaction and burial/thermal history, kerogen type, TOC and variations in TOC, both vertically and laterally. Thus, it is important to treat each continuous-type hydrocarbon accumulation as an individual situation and take from it what might be applied to the study of other similar accumulations.

The Rhinestreet shale has a rather complex burial/thermal history (see Fig. 9). Its early compaction history records the fitful burial of the sediment resulting in the formation of the laterally persistent concretionary horizons that could have influenced fluid migration. The marked increase in sedimentation rate in the latter half of the Famennian, likely induced by rapid progradation of the Catskill Delta complex, perhaps the result of glacio-eustacy, brought normal mechanical compaction of the Rhinestreet shale to an end, well shy of its maximum burial depth. However, this initial overpressure event caused by disequilibrium compaction, an Acadian event, failed to initiate fractures.

Morrowan time was marked by uplift of the Devonian-Mississippian succession in western New York state and western Pennsylvania, a result of the westward migration of a peripheral bulge produced by loading of the Laurentian plate edge at the onset of the Alleghanian orogeny. Burial/thermal modeling of

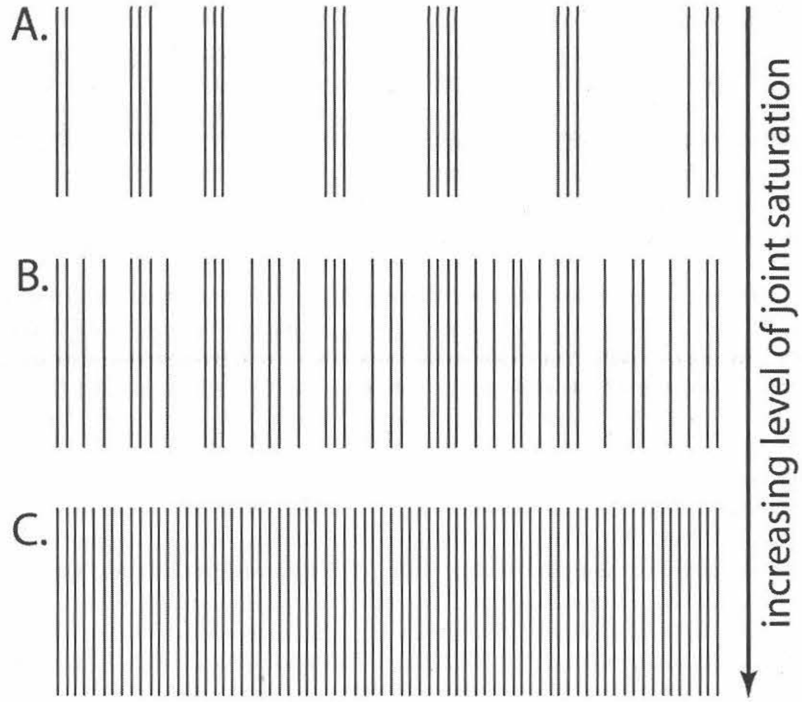


Fig. 39: Schematic diagram showing the evolution of fracture spacing (modified after Peacock and Mann, 2005). Refer to text for discussion.

measured vitrinite reflectance of the Rhinestreet shale suggests that the organic-rich rocks had not yet been buried to the oil window when they were uplifted (see Fig. 9). Thermoelastic contraction of the uplifted sequence in concert with reduced burial-related stress induced NS-trending joints in the higher modulus diagenetic carbonate. Moderately elevated P_p in the host gray shale as a consequence of the above-mentioned disequilibrium compaction enabled the newly formed joints to propagate through the shale as natural hydraulic fractures.

Renewed subsidence in Atokan time carried the Rhinestreet shale into the oil window by ~275 Ma. The initial result of catagenesis was the formation of horizontal bitumen-filled microcracks, which caused this tight unit to become even more impermeable. Soon after this, NW-trending vertical natural hydraulic fractures formed. The Rhinestreet appears to have entered the oil window roughly coeval with emplacement of the Blue Ridge-Piedmont Megathrust sheet. However, because the Rhinestreet may not have been in the oil window long enough to achieve little more than modest production, NW joints set did not come close to saturating the source rock. The Rhinestreet shale continued to produce hydrocarbons as it approached its modeled maximum burial depth of ~3.1 km. As a result, ENE joints that formed under the influence of the continued oblique plate convergence in the New England Appalachians came closer to saturating the Rhinestreet source rock. These joints, having more drive behind them, propagated higher in the sequence. ENE joints may have continued to form during post-Alleghanian uplift of the Appalachian Plateau as a consequence of thermal contraction.

Clearly, the geologic history of the Rhinestreet shale differs from other continuous-type hydrocarbon accumulations most notably in the timing of its entry into the oil window and resultant chronology, orientation and density of fracturing. These factors are crucial to exploration and production strategies of such a unit as the Rhinestreet shale. Our ongoing work demonstrates that a fuller understanding of the intricacies of a petroleum system like that of the Rhinestreet shale requires a multi-faceted approach that entails the analysis of the system from the submicroscopic scale to the lithospheric stress level.

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ROAD LOG AND STOP DESCRIPTIONS

Total Miles	Miles from Last Point	Route Description
0.0	0.0	Depart Adams Mark heading east on Church St. to its intersection with Route. 5.
0.2	0.2	Bear right onto Route 5 West.
15.3	15.1	Intersection of Route 5 with South Creek Road. Turn right onto South Creek Road.
15.6	0.3	Turn left onto Lake Shore Road.
17.5	1.9	Park in the small lot at the Erie County Water Authority building or at the intersection of Sweetland Road a short distance ahead. Walk down to Pike Creek.

STOP 1: **PIKE CREEK/LAKE ERIE SHORELINE**

Significance: I) the preferential generation of early NS-trending joints at the contacts of gray shale and black shale;
II) geographic variations in the magnitude of the pre-Upper Devonian or Taghanic unconformity and the significance of the so-called "Driller's Tully";
III) preferential propagation of NW and ENE joints in the Rhinestreet black shale.

I. Working down creek toward Lake Erie, the first exposure encountered reveals the contact of the West River gray shale and overlying Middlesex black shale (Fig. 40). A prominent NS-trending joint set as well as ENE-trending joints can be observed here (Fig. 41). The NS joints terminate at the contact or a short distance above it, a theme common to other Upper Devonian black shale units of western New York state as described earlier. We interpret the generation of the NS joints to be a response to the passage of a peripheral bulge produced by loading of the Laurentian plate edge at the onset of the Alleghanian orogeny across the western New York state region of the Appalachian Basin. Reduced compressive stress as a consequence of uplift-related thermoelastic loading initiated joints in higher modulus diagenetic carbonate. Propagation of these joints through the host gray shale was aided by moderately overpressured pore fluids at the top of the gray shale, perhaps sealed in by the overlying low permeability black shale.

II. The mouth of Pike Creek exposes the Middle-Upper Devonian interval from the Tichenor limestone (perhaps the top of the Ludlowville shale, depending upon the lake level) to roughly two-thirds the way up the Middlesex black shale (Fig. 40). This exposure provides the opportunity to (1) consider the magnitude of the Taghanic unconformity in the subsurface of western New York state and (2) offer some preliminary thoughts on the subsurface stratigraphy of the Moscow-Tully interval in this region of the basin. De Witt and Colton (1978) placed the unconformity at the contact of the Penn Yan shale of the Geneseo Formation and the North Evans limestone of the underlying Moscow shale, Hamilton Group (Fig. 40). Rocks absent this exposure include, in descending order, an unknown thickness of the Penn Yan shale and the underlying Geneseo black shale, the Tully limestone, and some thickness of the Moscow shale (Windom Member). The hiatal extent or magnitude of the Taghanic unconformity diminishes to the south and east as the Moscow shale increases in thickness and the Geneseo black shale and Tully limestone appear and thicken (Fig. 42). De Witt et al. (1993) maintained that the disconformity confined the Middle Devonian Geneseo black shale to the eastern region of the basin in New York state; however, Upper Devonian black shales, including the Dunkirk, Pipe Creek, and Rhinestreet shales were largely restricted to the western part of the basin by the unconformity. However, the distribution of the Middle Devonian Oatka Creek black shale of the Marcellus Formation (Hamilton Group) appears to have been unaffected by the Taghanic unconformity.

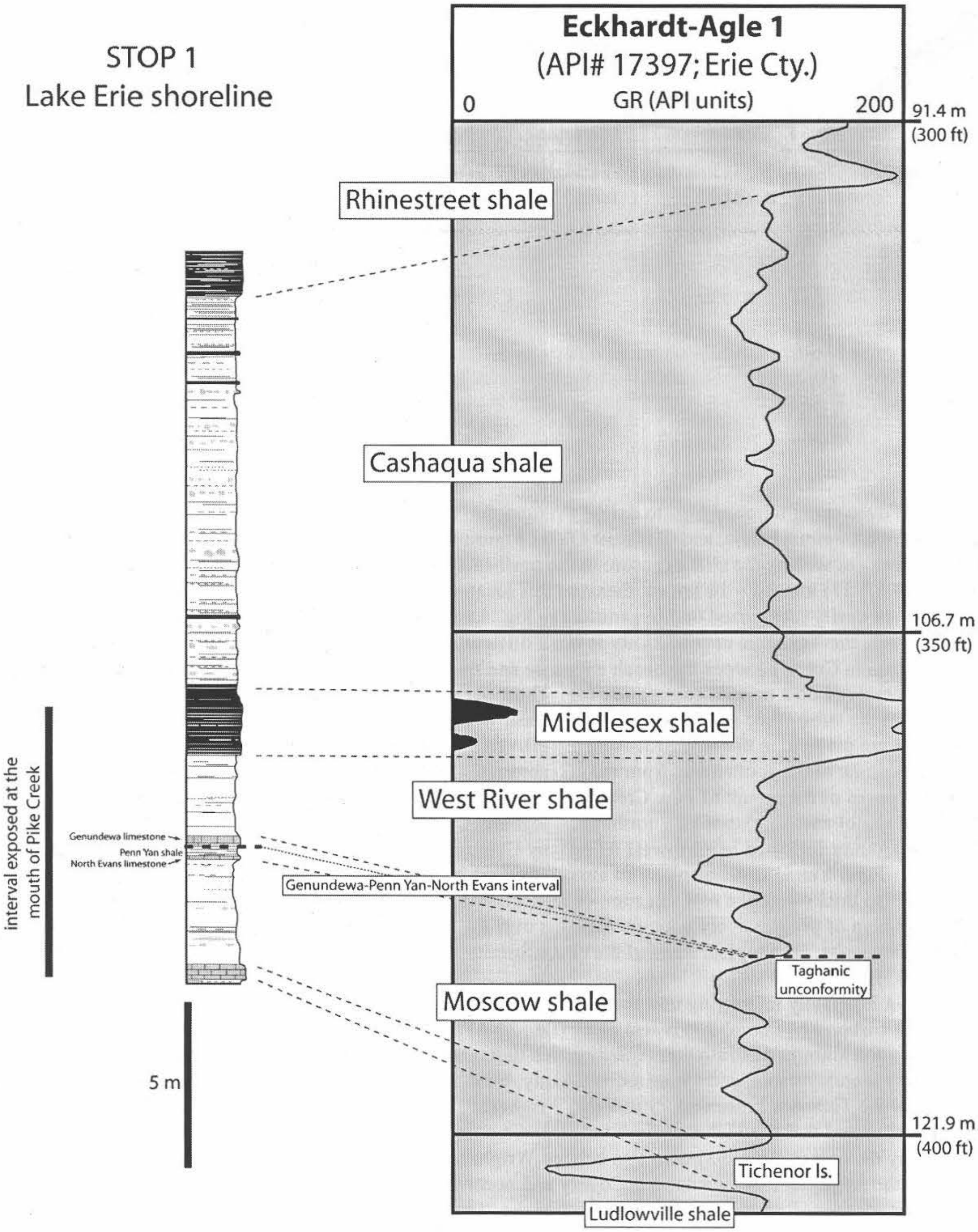


Fig. 40: Stratigraphic section of the beach cliff northeast of the mouth of Pike Creek (showing the interval exposed at the mouth of Pike Creek). The lithologic log is based on measurements made along the lakeshore and Eighteenmile Creek. This section is traced into the subsurface to the Eckhardt-Agle 1 gas well (see Fig. 42 for location of this well).

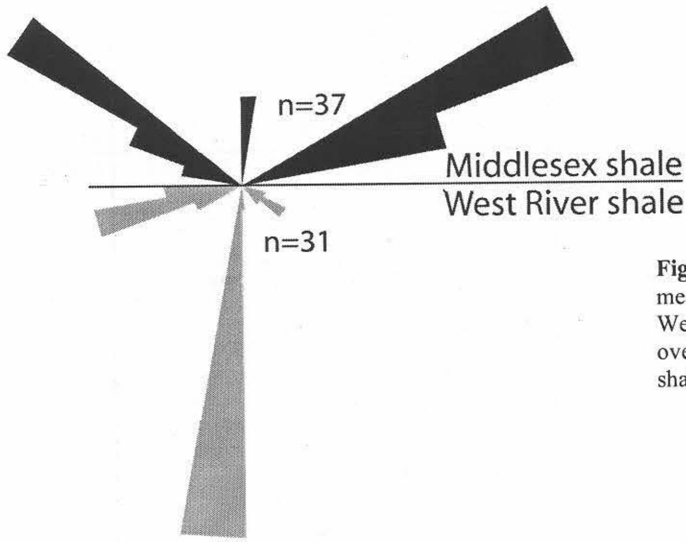
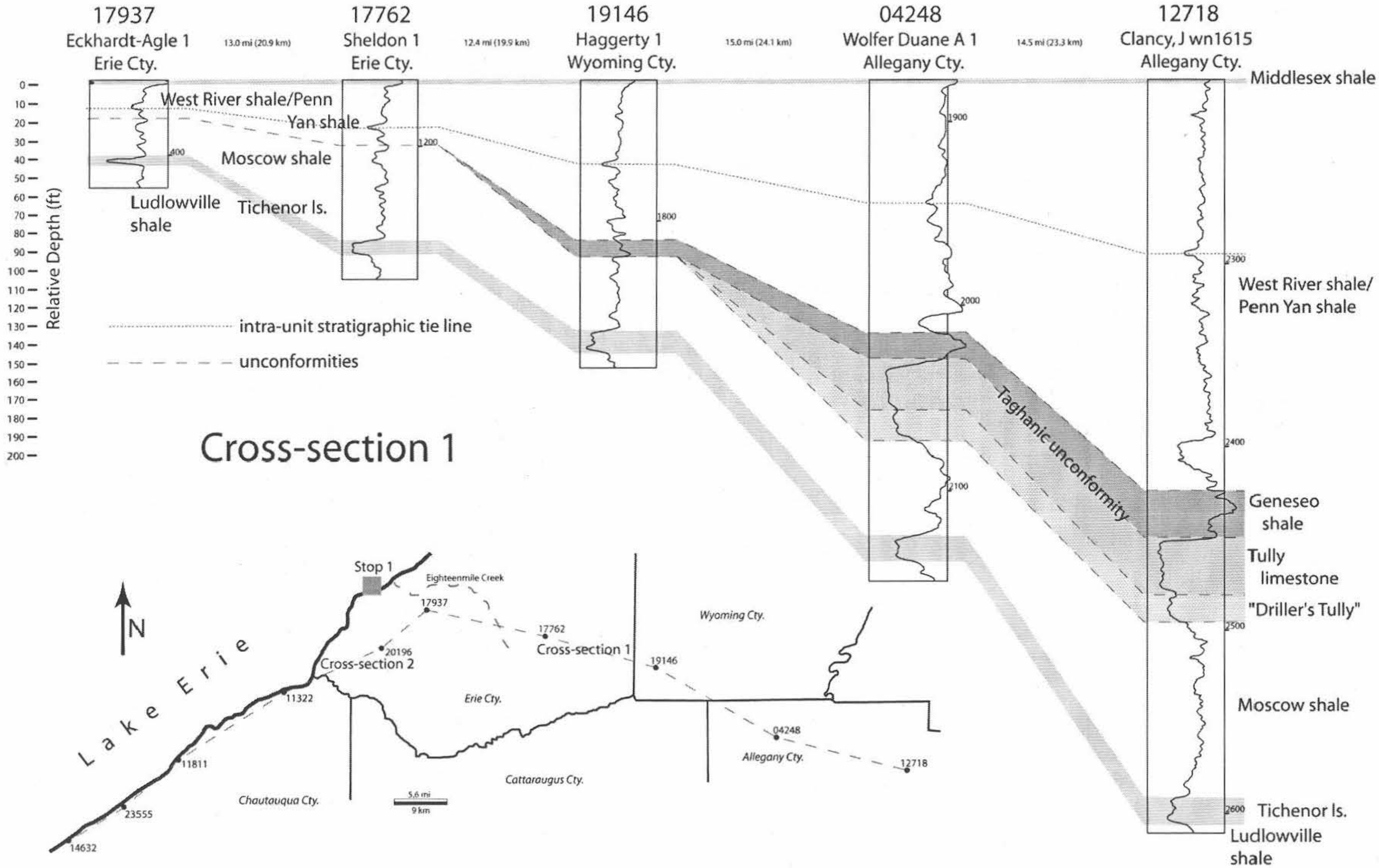


Fig. 41: Rose plot of joints measured at the contact of the West River gray shale and overlying Middlesex black shale along Pike Creek.

We are using wireline logs to define stratigraphic trends in the subsurface of western New York state, northwest Pennsylvania, and northeast Ohio. Figure 40 links the stratigraphy observed at the mouth of Pike Creek (as well as the cliff exposure to the northeast along the beach) to the Eckhardt-Agle 1 gas well (API# 17937) located ~ 10 km southeast of this location (Fig. 42). The Middlesex shale and Tichenor limestone are easily recognized on the gamma-ray log. Locating the North Evans-Genundewa interval, and, therefore, the Taghanic unconformity is more difficult. The Genundewa limestone, ~21 – 30 cm thick on Eighteenmile Creek between the beach exposure and the Eckhardt-Agle 1 well (Fig. 42), is difficult at best to be resolved on a normal gamma-ray log. Thus, our placement of the unconformity is based on extension of the subsurface stratigraphy from the east, where the Tully limestone and overlying Geneseo black shale are present, to the Eckhardt-Agle 1 well. There is a marked increase in the thickness of the West River shale to the south away from the lakeshore (Fig. 40). Cross-section 1 (Fig. 42) carries the interval exposed at the mouth of Pike Creek (i.e., the base of the Tichenor limestone to the base of the Middlesex shale) east-southeast into northeast Alleghany County. The Geneseo black shale first appears in Wyoming County followed in northwest Alleghany County by the Tully limestone. Notice also that the West River/Penn Yan interval displays an almost six-fold increase in thickness to the east. The Tichenor limestone also thickens to the east. In sum, cross-section 1 displays, over a relatively short distance, the westward onlap of progressively younger deposits over the Taghanic unconformity as well as the truncation of underlying units, both typical traits of the unconformity elsewhere in the Appalachian Basin (Hamilton-Smith, 1993).

An especially interesting unit that we are mapping in the subsurface, the so-called "Driller's Tully," is a carbonate interval that is less clean (i.e., 20 API units or more) than the true Tully limestone. There is a good amount of confusion regarding this unit; Oliver et al. (1969) placed the "Driller's Tully" of Erie County, Pennsylvania, and Chautauqua County, New York, well down the Hamilton Group, perhaps at the level of the Tichenor limestone. Similarly, Wright (1973) referred to the "Driller's Tully" as limestone "B" and suggested that the Tichenor limestone is a tongue of limestone "B." This interpretation was accepted by de Witt et al. (1993) who also echoed Wright's proposition that the Tully limestone is not present in the subsurface of Chautauqua County and northwest Pennsylvania. De Witt et al. (1993) pointed out that conodonts collected from the upper part of the "Driller's Tully" in a core from the Eastern Gas Shales Project well PA3 in Erie County, Pennsylvania, included elements of *Polygnathus linguiformis*,

Fig. 42 (next page): Stratigraphic cross-section (gamma-ray logs) of the interval observed at the mouth of Pike Creek east into northern Allegany County (datum = base of the Middlesex shale).


17937
Eckhardt-Agle 1
Erie Cty.
13.0 mi (20.9 km)
17762
Sheldon 1
Erie Cty.
12.4 mi (19.9 km)
19146
Haggerty 1
Wyoming Cty.
15.0 mi (24.1 km)
04248
Wolfer Duane A 1
Allegany Cty.
14.5 mi (23.3 km)
12718
Clancy, J wn1615
Allegany Cty.
Middlesex shale
Relative Depth (ft)
0
10
20
30
40
50
60
70
80
90
100
110
120
130
140
150
160
170
180
190
200
West River shale/Penn Yan shale
Moscow shale
Ludlowville shale
Tichenor ls.
1200
1400
1800
1900
2000
2100
2300
2400
2500
2600
intra-unit stratigraphic tie line
unconformities
Taghanic unconformity
West River shale/ Penn Yan shale
Geneseo shale
Tully limestone
"Driller's Tully"
Moscow shale
Tichenor ls.
Ludlowville shale
Cross-section 1
N
Lake Erie
Stop 1
Eighteenmile Creek
17937
17762
20196
Cross-section 2
Cross-section 1
19146
11322
11811
23555
14632
04248
12718
Wyoming Cty.
Erie Cty.
Allegany Cty.
Cattaraugus Cty.
Chautauqua Cty.
5.6 mi
9 km

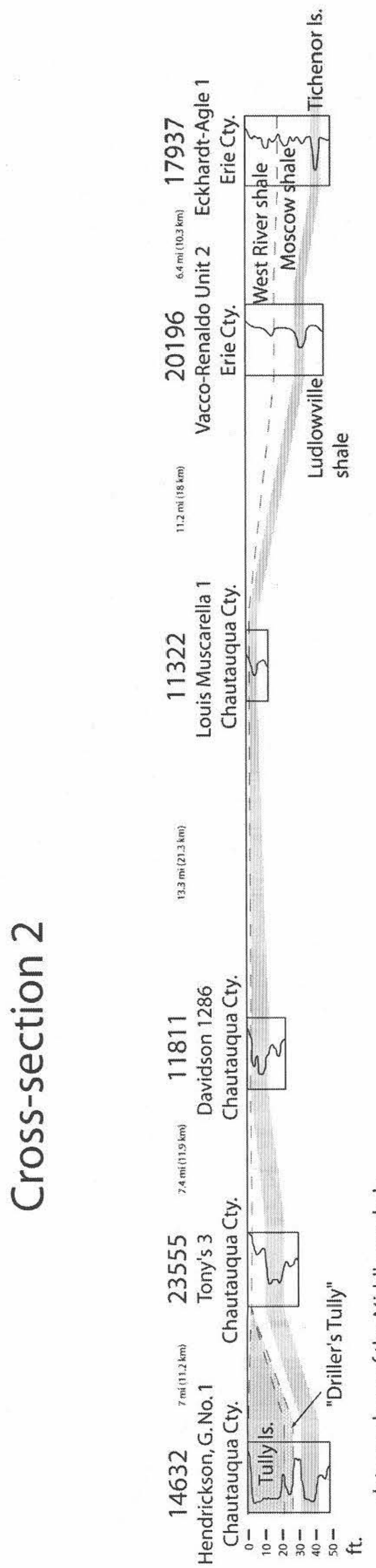


Fig. 43: Stratigraphic cross-section (gamma-ray logs) of the interval observed at the mouth of Pike Creek west into northern Chautauqua County (datum = base of the Middlesex shale).

We agree with Rickard (1989) that the Tully limestone is present in the subsurface of Chautauqua County and northwest Pennsylvania. Indeed, our analysis of close-spaced wells in far western Chautauqua County indicates that Wright's (1973) Tichenor limestone and limestone "B" is actually the Tully limestone and underlying "Driller's Tully," the latter resting unconformably on the Tichenor limestone. However, we interpret the "Driller's Tully" to be a discrete unit in erosional contact with the Moscow shale, and locally the Tichenor limestone, rather than a stratigraphic equivalent of the Moscow shale and Tichenor limestone as suggested by Rickard (1989). Truncated calcareous intervals of the Moscow shale and locally abrupt changes in thickness of the "Driller's Tully" confirm the erosional contact of this unit with underlying deposits. The "Driller's Tully" is recognized in the subsurface of northern Allegany County, that area of Chautauqua County where the Tully is present (Fig. 43) and parts of Warren and Erie counties, Pennsylvania. The contact of the "Driller's Tully" and overlying Tully limestone is a thin shaley interval (Figs. 42 and 43). However, the local absence of this interval and as well as the occasional absence of the "Driller's Tully" beneath the Tully suggests that the contact of these units is an unconformity.

Cross-section 2 extends southwest along the Lake Erie shoreline from the Eckhardt-Agle 1 well to Westfield, New York, northern Chautauqua County (Fig. 43). Noteworthy features in this section include: (1) the thickening of the Tichenor limestone to the west of Erie County; (2) the appearance of the Tully and underlying "Driller's Tully" in the Westfield area; and (3) the thinning of the interval between the base of the Tichenor limestone and the base of the Middlesex shale to the southwest (the Middlesex and Tichenor are separated by less than one meter of gray shale in the Muscarella 1 well; Fig. 43) followed by its rapid thickening as the Tully reappears near Westfield. This latter point is consistent with localized uplift. That the Middlesex shale and younger deposits show no evidence of this event provides clues to the timing of uplift, perhaps a reflection of localized tectonic activity of the post-Tully basin in advance of the Acadian thrust load and related forebulge (e.g., Tankard, 1986; Beaumont et al., 1988; Hamilton-Smith, 1993; Ver Straeten and Brett, 2000; Filer, 2003). Indeed, Filer (2003) defined a Famennian forebulge in West Virginia based on isopach and sedimentary facies maps that, when traced north, extends through the eastern Chautauqua County-Erie County region. We are just beginning to consider this event in a regional sense based on subsurface information from western New York state, northwest Pennsylvania, and northeast Ohio.

The Taghanic unconformity appears to be erosional throughout western New York state (Johnson, 1970; Wright, 1973; Baird and Brett, 2003). Heckel (1973) maintained that the contact is a slight diastem in central New York state, increasing in stratigraphic magnitude to the west where the disconformity represents an increasing period of nondeposition. However, our ongoing investigation of the subsurface stratigraphy of western New York state reveals a more complex situation made more so by the presence of the "Driller's Tully." Indeed, the initial event may have been that phase of erosion that occurred before deposition of the "Driller's Tully," which locally rests unconformably on the Tichenor limestone.

Heckel (1973) pointed out that the Geneseo black shale conformably overlies the Tully limestone east of Cayuga Lake. West of this meridian, however, the contact becomes one of disconformity as progressively younger deposits onlap westward above the unconformity (de Witt et al., 1993) until the Geneseo shale feathers out in easternmost Chautauqua County into eastern Wyoming County (Fig. 42). Rickard (1989) suggested that the post-Tully unconformity was entirely subaqueous. Similarly, Baird and Brett (1991) postulated that the erosional contact of the Geneseo black shale and underlying Tully limestone is a submarine erosional feature (maximum flooding surface) related to the abrupt increase in water depth, likely the result of the sudden sinking of a forebulge in response to tectonic loading to the east (e.g., Hamilton-Smith, 1993).

III. Moving northeast along the beach for perhaps a kilometer enables us to observe in cliff exposure the entire Moscow shale – basal Rhinestreet shale interval (Fig. 40). Note the diagenetic carbonate intervals within the Cashaqua shale. The heavily fractured Rhinestreet black shale at the top of the exposure is easily recognized from the beach. The upper several meters of the Cashaqua and lower few meters of the Rhinestreet shale carry the older NS-trending joints, similar to what we saw on Pike Creek and likely for the same reasons. NW- and younger ENE-trending joints, most densely formed in the basal organic-rich interval of the Rhinestreet shale, formed as a consequence of catagenesis as these source rocks approached and finally reached their modeled maximum burial depth of ~3.1 km (see Fig. 9). This exposure demonstrates the strong stratigraphic control on jointing in these rocks. The NW joints probably formed soon after the rocks had entered the oil window and at that time that the basin was feeling the same remote stress field responsible for emplacement of the Blue Ridge-Piedmont Megathrust. The younger ENE-trending joints propagated after the Rhinestreet shale had been in the oil window for a protracted period of time, thus their greater density. The stress field under which these joints formed reflected the oblique convergence of Laurentia and Africa and had been responsible for the generation of the early (perhaps Chesterian) ENE-trending coal cleat throughout the Appalachian Basin and similarly oriented joints in the deeply buried Frasnian black shales of the Finger Lakes District (Engelder and Whitaker, 2006).

Total Miles	Miles from Last Point	Route Description
17.5	0.0	Head east on Lake Shore Road.
19.4	1.9	Turn right (south) onto South Creek Road.
19.7	0.3	Intersection of South Creek Road and Route 5. Remain on South Creek Road.
21.2	1.5	Cross Versailles Plank Road.
21.5	0.3	Intersection of South Creek Road and Route 20 (Southwestern Boulevard). Remain on South Creek Road.
21.6	0.1	Park along the shoulder of South Creek Road at the North Evans Cemetery. Cross South Creek Road and walk down the gravel path to Eighteenmile Creek heading toward the white Route 20 bridge.

STOP 2: EIGHTEENMILE CREEK/ROUTE 20 BRIDGE

Significance: I) approximately the lower half of the Rhinestreet shale can be observed in the east wall of the creek exposure immediately south of the Route 20 bridge over Eighteenmile Creek;
II) the presence of NS joints at the contact of the Cashaqua gray shale and overlying Rhinestreet shale, and the high density of NW- and younger ENE-trending joints at the organic-rich base of the Rhinestreet shale;
III) basin modeling of measured vitrinite reflectance and the possibility of vitrinite suppression;
IV) recognition of the Cashaqua-Rhinestreet contact in the subsurface – gamma-ray vs. bulk density logs.

I. A bit less than the lower half of the Rhinestreet shale is visible in the cliff exposure immediately north of the Route 20 bridge. We will have an opportunity to look at the concretion horizon – the middle concretion horizon – visible at the top of the cliff – ~34 m above the bottom contact (see Fig. 1) - at Stop 4. The Rhinestreet section displayed in the cliff contains two intervals of gray shale and thin-bedded black shale at ~10.0-12.5 m and 27.5-29.3 m above the contact (Fig. 1). These horizons, which probably represent eustatic oscillations, can be observed in gamma-ray and bulk density logs (Fig. 44). Moreover, the lower gray shale interval contains a lenticular limestone bed as thick as 20 cm, one of Luther's (1903) scraggy layers and our lower scraggy layer (see Fig. 1). The geochemistry of the scraggy layers, discussed above, confirms their early diagenetic origin and suggests that they formed in a manner similar to the concretions. The scraggy layers, which are consistently found associated with organic-lean gray shale intervals (see Fig. 1), are recognized in gamma-ray logs as a sharp deflection to the left (Fig. 44) and may serve as markers for basinwide correlation. The fact that the limestone deflection on the gamma-ray logs cannot always be correlated among neighboring offset wells demonstrates the discontinuous nature of these deposits, which is evident in field exposures.

II. The Cashaqua shale-Rhinestreet shale contact is exposed beneath the bridge. The Cashaqua shale along Eighteenmile Creek comprises ~11.5 m of silty, organic lean (TOC < 0.7%) gray shale and nodular limestone horizons. The overlying Rhinestreet shale is composed of ~54 m of organic-rich black shale, much lesser gray shale, sparse thin siltstone beds, and several carbonate concretion horizons and nodular limestone intervals, the latter termed "scraggy layers" by Luther (1903; see Fig. 1 for the measured Rhinestreet section along Eighteenmile Creek). TOC content of the Rhinestreet shale at its base is 8.09%, diminishing upward (Fig. 1; Lash and Blood, 2004b; Lash et al., 2004). Vitrinite reflectance of the Rhinestreet shale at its base along Eighteenmile Creek is 0.74%, indicating that these rocks entered the oil window during their burial history (Lash, 2006c).

Joints of the three dominant sets carried by the Rhinestreet shale are represented at this exposure. Several long NS-trending joints, can be seen at the contact. We contend that uplift of the Upper Devonian sedimentary succession in response to passage of a peripheral bulge at the onset of the Alleghanian orogeny, and resultant thermoelastic loading, induced the NS joints in the high modulus carbonate. These joints propagated upward under the influence of moderately overpressured pore fluid at the top of the Cashaqua shale. Elevated pore pressure may have resulted from a combination of disequilibrium compaction and the tight, impermeable nature of the organic-rich Rhinestreet shale. The lack of bioturbation in the finely laminated organic-rich sediment preserved a strong strength anisotropy that precluded easy vertical movement of fluid. The squeezing of ductile organic matter into voids and pore throats further reduced the permeability of the black shale (Lash, 2006b).

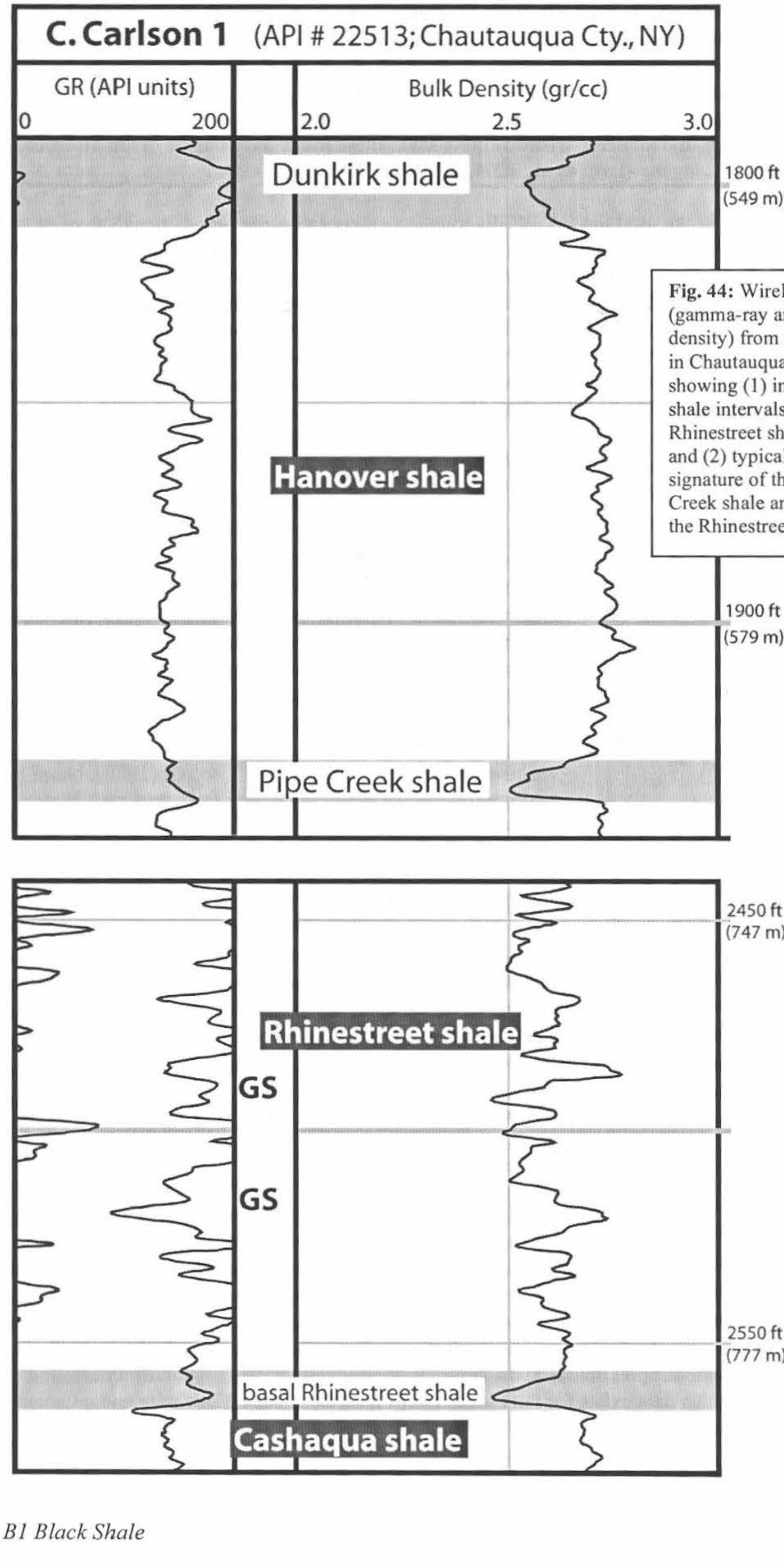


Fig. 44: Wireline logs (gamma-ray and bulk density) from a gas well in Chautauqua County showing (1) inferred gray shale intervals in the Rhinestreet shale (**GS**) and (2) typical wireline signature of the Pipe Creek shale and base of the Rhinestreet shale.

NW- and ENE-trending joints are densely formed in the black shale immediately above the contact with the Cashaqua shale (station EMC 13 and 16, Fig. 38). A similar relationship of joint density and TOC has been observed in the Dunkirk shale along the Lake Erie shoreline as well as in the Geneseo shale of the Finger Lakes District (Loewy, 1995; Lash et al., 2004). The correlation of joint density with TOC provides compelling evidence that the joints formed as natural hydraulic fractures as a consequence of the transformation of organic matter to hydrocarbons when these rocks were buried to the oil window during the Alleghanian orogeny (Lash et al., 2004; Lash and Engelder, 2005).

III. As noted earlier, we used the EASY%R_o kinetic model of vitrinite reflectance (Sweeney and Burnham, 1990) to model the burial/thermal history of the Rhinestreet black shale. Thermal modeling by use of the EASY%Ro algorithm requires knowledge of (1) the age(s) of the unit(s) of interest (the base of the Rhinestreet shale), (2) at least a partial thickness of the local stratigraphic sequence, and (3) the measured vitrinite reflectance of the unit(s) of interest (vitrinite reflectance of the base of the Rhinestreet shale = 0.74%). Our model for the Rhinestreet shale, which includes passage of the early Alleghanian peripheral bulge (see Fig. 9), requires ~1.5 km of Permian strata in this area of the basin, an amount close to that modeled by Beaumont et al. (1988) (Fig. 45).

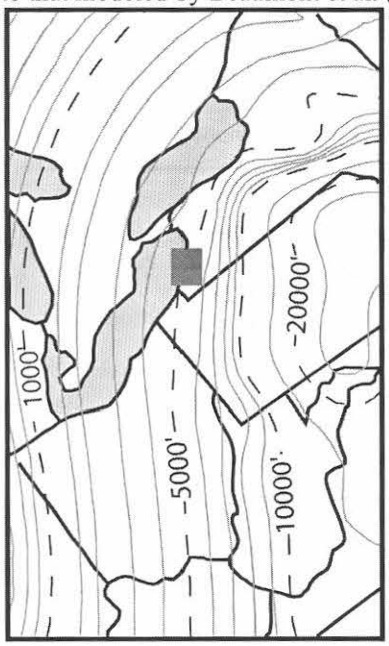


Fig. 45: Modeled thickness of Permian strata in the Appalachian Basin (modified after Beaumont et al., 1987).

Our method of modeling the burial/thermal history of the Rhinestreet shale depends on accurate %R_o values. However, several workers (e.g., Nuccio and Hatch, 1996; Rimmer and Cantrell, 1988; Rimmer et al., 1993; Rowan et al., 2004) have recognized the likely occurrence of vitrinite suppression in Devonian-Mississippian black shales, including those of the Appalachian Bain. Suppression of vitrinite leads to an underestimation of thermal maturity and, therefore, an underestimated burial depth. Suppression is most readily recognized by anomalously low %R_o on maturity profiles of wells or stratigraphic columns (e.g., Price and Barker, 1985). There are several lines of evidence suggesting that Devonian black shales of the Appalachian Basin have been affected to some extent by vitrinite suppression. Rimmer and Cantrell (1988), based on a vitrinite reflectance gradient derived from coal rank trends, claim that measured %R_o values of the Cleveland Member of the Ohio shale in eastern Kentucky are suppressed by as much as 0.3 – 0.5%R_o. More recently, Rowan et al. (2004), in the course of their burial/thermal modeling of a cross-section from central West Virginia to northeast Ohio, demonstrated that the predicted %R_o of Devonian shale is locally twice the measured %R_o. Rowan et al. (2004) constrained their burial/thermal model with measured %R_o of Pennsylvanian coals and color alteration index values of conodonts recovered from Ordovician and Devonian limestones.

We suggest that Middle and Upper Devonian black shale of the western New York state and northwest Pennsylvania region of the Appalachian Basin was also been affected by suppression. Our data base includes published thermal/maturity values (Weary et al., 2000; Repetski et al., 2002) and %R_o and Rock-Eval data from our work on the Upper Devonian shale succession of the Lake Erie District. We limited our data to the western New York state and northwest Pennsylvania region of the basin in order to assume a common thermal history for these deposits. A plot of Rock-Eval HI versus %R_o of Appalachian Basin black shale samples, principally from the Marcellus shale, reveals a trend similar to that observed by Nuccio and Hatch (1996) for the Upper Devonian-Mississippian New Albany shale, which has been affected by vitrinite suppression (Fig. 46). The higher HI samples, including samples collected from the Dunkirk and Rhinestreet shales, fall along a relatively narrow %Ro trend; however, at an HI of ~125 mg/g, measured %R_o increases rapidly from ~ 0.75 to > 2.5 as HI drops to near zero (Fig. 46). This plot appears to show suppressed values in the lower to mid oil window (%R_o = ~0.7 to 1.0), consistent with Lo's (1993) findings regarding vitrinite suppression of hydrogen-rich organic matter. An especially interesting subpopulation of Marcellus data exists for the Finger Lakes District of New York state and adjacent Potter and Tioga counties, Pennsylvania (Fig. 46). These data are defined by %R_o ranging from 0.5 to 1.0 and HI

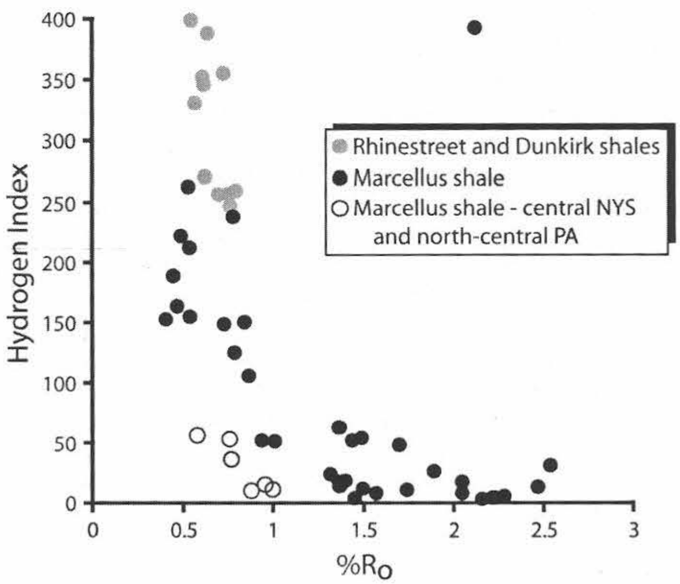


Fig. 46: Rock-Eval HI versus $\%R_o$ for Middle and Upper Devonian shale samples collected from the western New York state and northwest Pennsylvania region of the Appalachian Basin.

< 55 (Fig. 46). These samples, which appear severely suppressed, were collected from that part of the basin affected by emplacement of Cretaceous ultramafic intrusions (Kay et al., 1983).

Upper Devonian black shales exposed along the Lake Erie shoreline show a good amount of variation in measured $\%R_o$ over a relatively small distance (see Fig. 7). Measured %Ro in a well cutting sample collected from the base of the Rhinestreet shale in northern Chautauqua County is 0.51%. The $\%R_o$ of the same stratigraphic interval ~40-45 km to the northeast is 0.73 – 0.77 based on the analysis of four outcrop samples. Finally, a sample collected from the base of the Rhinestreet 7.25 km to the east on Cazenovia Creek has a $\%R_o$ of 0.59. We recognize similar erratic variations in the measured $\%R_o$ for Dunkirk shale samples. Rowan et al. (2004) suggest that variations in the type of organic matter in marine sediments may account for erratic variations in $\%R_o$ (e.g., Ujiié et al., 2004).

Vitrinite suppression has also been attributed to the overpressuring of rocks (McTavish, 1978 1998; Carr, 2000). Pervasive natural hydraulic fractures in the Rhinestreet shale provides evidence that these rocks were overpressured during their burial history, which could have contributed to suppression of virinite reflectance. Regardless of its mode of origin, suppression of vitrinite reflectance is a phenomenon that should be taken into account in any exploration and production considerations of Devonian black shales.

IV. The contact of the Cashaqua gray shale and overlying Rhinestreet black shale is readily defined in the field; both color contrast and relative weathering profile provide easy means for making this recognition. The contact is rather obvious in gamma-ray logs, which provide a measure of the intensity of natural gamma-rays (uranium concentration) emitted by a formation (Fig. 44). Still, the initial deflection marking the contact is not what one might expect based on the abrupt contact between the low-TOC (<0.7%) Cashaqua shale and the carbonaceous (TOC>8%) Rhinestreet shale. However, the Cashaqua-Rhinestreet contact is very obvious in bulk density logs (Fig. 44), no doubt a reflection of the low density of organic matter relative to shale minerals (Smith and Young, 1964; Schmoker, 1979). Schmoker (1981) claimed that the bulk density log is a more accurate proxy for organic matter content than the gamma-ray API value. Indeed, the strong inverse relationship of bulk density and TOC in Upper Devonian black shale of western New York state has been confirmed by a proprietary data set we have had the opportunity to study (Lash, 2006c). It is interesting to note that the organic-rich Pipe Creek black shale (TOC = 3.2 - 7.9%) is not easily recognized on the gamma-ray log, but is readily picked on the bulk density log (Fig. 44).

On the other hand, the base of the overlying Dunkirk shale is readily recognized on both types of logs (Fig. 44).

Lüning and Kolonic (2003) have pointed out that many black shale deposits show a strong correlation of uranium, manifested by gamma-radiation in API units, and TOC. However, they also note those examples where the correlation is weak at best. The basal Rhinestreet shale and the Pipe Creek shale provide examples of the latter phenomenon. Indeed, a spectral gamma-ray log from a well in northeast Chautauqua County reveals the base of the Rhinestreet shale to be marked by an increase in uranium from 6 to 10 ppm; the Pipe Creek shale is defined by an increase of only 2 ppm. Lüning and Kolonic (2003) suggest that the precipitation of uranium from seawater onto the surface of organic particles under anoxic conditions is governed by several factors, including carbonate content, sedimentation rate, and primary uranium content of the seawater. Upper Devonian shales of western New York state are defined by very low $CaCO_3$ abundances ruling out the first factor. Lüning and Kolonic (2003) point out that the longer organic matter is in contact with seawater, the greater the amount of uranium that can precipitate onto organic grains. Low sedimentation rates favor high uranimum/TOC ratios. The anomalously low API values of the most organic-rich interval of the Rhinestreet shales (as deduced by TOC analysis of field samples and bulk density values determined from wireline logs) may indicate a high sedimentation rate of this interval such that lesser amounts of uranium were fixed onto organic particles. The same can be said of the Pipe Creek shale. The organic-rich base of the Dunkirk shale (Lash et al, 2004; Lash and Engelder, 2005), however, defined by the highest API value and lowest bulk density of the unit, would have accumulated relatively slowly.

Finally, it is possible that the introduction of low uranium freshwater into the Devonian Sea, such as might be expected as a consequence of the reduction of polar ice volume, was responsible for some of the variations in gamma-ray response observed in wireline logs of the base of the Rhinestreet shale and especially the Pipe Creek shale, which has been correlated with the Lower Kellwasser Bed recognized in Europe (Over, 1997). The concentration of uranium in seawater is an order of magnitude greater than freshwater (Broecker and Peng, 1982). The presence of the diagnostic uranium signature of the Pipe Creek shale in well logs recovered from northwest Pennsylvania, western New York state, northeast Ohio, and northernmost West Virginia suggests a basinwide cause. The Pipe Creek interval is a particularly interesting horizon. McGhee (1990) argued that this interval is a significant crisis interval similar to the younger Frasnian-Famennian boundary. In fact, he suggested that the Pipe Creek interval was more critical in terms of the maximum number of species extinctions and the extinction rate. It has been suggested that the Kellwasser Beds, including the Pipe Creek shale, resulted from eustatic transgressions perhaps caused by a reduction in polar ice volume (Sandberg et al., 1988; Buggisch, 1991; Joachimski and Buggisch, 1993). It may be that the same global mechanism that created the Pipe Creek crisis – rapid climatic warming – also accounts for the unique wireline log signature of the Pipe Creek shale.

Total Miles	**Miles from Last Point**	**Route Description**
21.6	0.0	Head north on South Creek Road.
21.7	0.1	Intersection of South Creek Road and Route 20. Remain on South Creek Road.
23.8	2.1	Turn left onto Lake Shore Road.
25.8	2.0	Intersection of Sweetland Road and Lake Shore Road.
27.0	1.2	Turn right onto Sturgeon Point Road.
27.9	0.9	Follow Sturgeon Point Road into the Sturgeon Point Marina. Continue through the marina (slowly) to the lower (east) parking area.

STOP 3: **STURGEON POINT**

Significance: I) comparison of the degree of jointing in black shale versus gray shale and the preferential generation of NS systematic joints in gray shale at contacts with overlying black shale;
II) carbonate concretions in black shale and analysis of compaction strain;
III) variation in the thickness of the bottom gray shale interval of the Rhinestreet shale.

I. Depart the parking area walking northeast toward the bedrock cliff a few hundred meters distant. We will follow this exposure of gray and black shale of the Rhinestreet shale for perhaps 0.75 km. The 2.75-m-thick gray shale interval exposed at the west end of the exposure is the equivalent of the lower gray shale interval of the Rhinestreet shale on Eighteenmile Creek (see Fig. 1; 9.9-12.5 m). The gray shale includes several horizons of high aspect ratio (lenticular) carbonate concretions and thin grayish-black shale layers. One can readily observe differences in the degree (density) of jointing between black and gray shale in this exposure. Black shale above and below the gray shale carries NW joints and younger ENE-trending systematic joints (Fig. 47). NW and ENE joints can be observed in the gray shale, but these deposits also contain the older NS-trending systematic joints (Fig. 47). The black shale clearly is more densely jointed than the gray shale, a recurring theme throughout the Appalachian Plateau (Fig. 48; Loewy, 1995; Lash et al., 2004). Moreover, ENE joints are more densely formed (i.e., closer to saturation) than are the older NW joints. This latter observation can also be made of ENE and NW joints in the gray shale (Fig. 48). NS-trending systematic joints in the gray shale are more densely formed than are the younger NW joints. As described earlier, the NS joints are interpreted to have originated in the higher modulus carbonate concretions as a consequence of thermoelastic loading caused by passage of a peripheral bulge across the Appalachian Basin at the onset of the Alleghanian orogeny. Indeed, it is more common for a NS joint to penetrate or cut a higher modulus carbonate concretion than it is for ENE and NW joints to cleave concretions. Note also that very few NS joints penetrate more than a few tens of cm into the overlying black shale; more often than not, they are arrested at the base of the organic-rich deposits. This exposure presents a nice example of the stratigraphic control on jointing in these rocks.

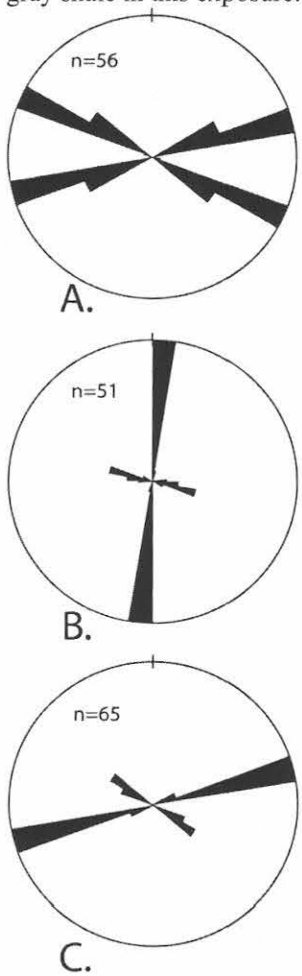


Fig. 47: Rose plots of joints in the (A) upper black shale, (B) gray shale interval, and (C) lower black shale at Stop 3.

Continue NE along the beach toward a large drainage pipe in the cliff. Careful observations of the gray shale horizon while working along the beach to this location would have revealed a gradual decrease in thickness of this interval. The thickness of the gray shale immediately east of the drainage pipe (determined by digging through beach sand to black shale) is ~ 80 cm, markedly reduced from what is was at the start of this trek along the beach. Note also that the gray shale is not as complexly jointed here; indeed, the median spacing of NS joints is 50% greater than at the western end of the exposure (Fig. 48). Finally, the thickness of the gray shale interval at the NE end of the exposure is at least 1.5 m. NW- and NS-trending joints are carried by the gray shale in this area. Further, NS joints are more densely formed at this location of the exposure than elsewhere along the beach (Fig. 48).

There is a school of thought that holds that all jointing in the exposed Devonian succession of the Appalachian Plateau of western New York state is a consequence of post-Alleghanian (perhaps even post-glacial) uplift of the Appalachian Basin. This view fails to take into account the

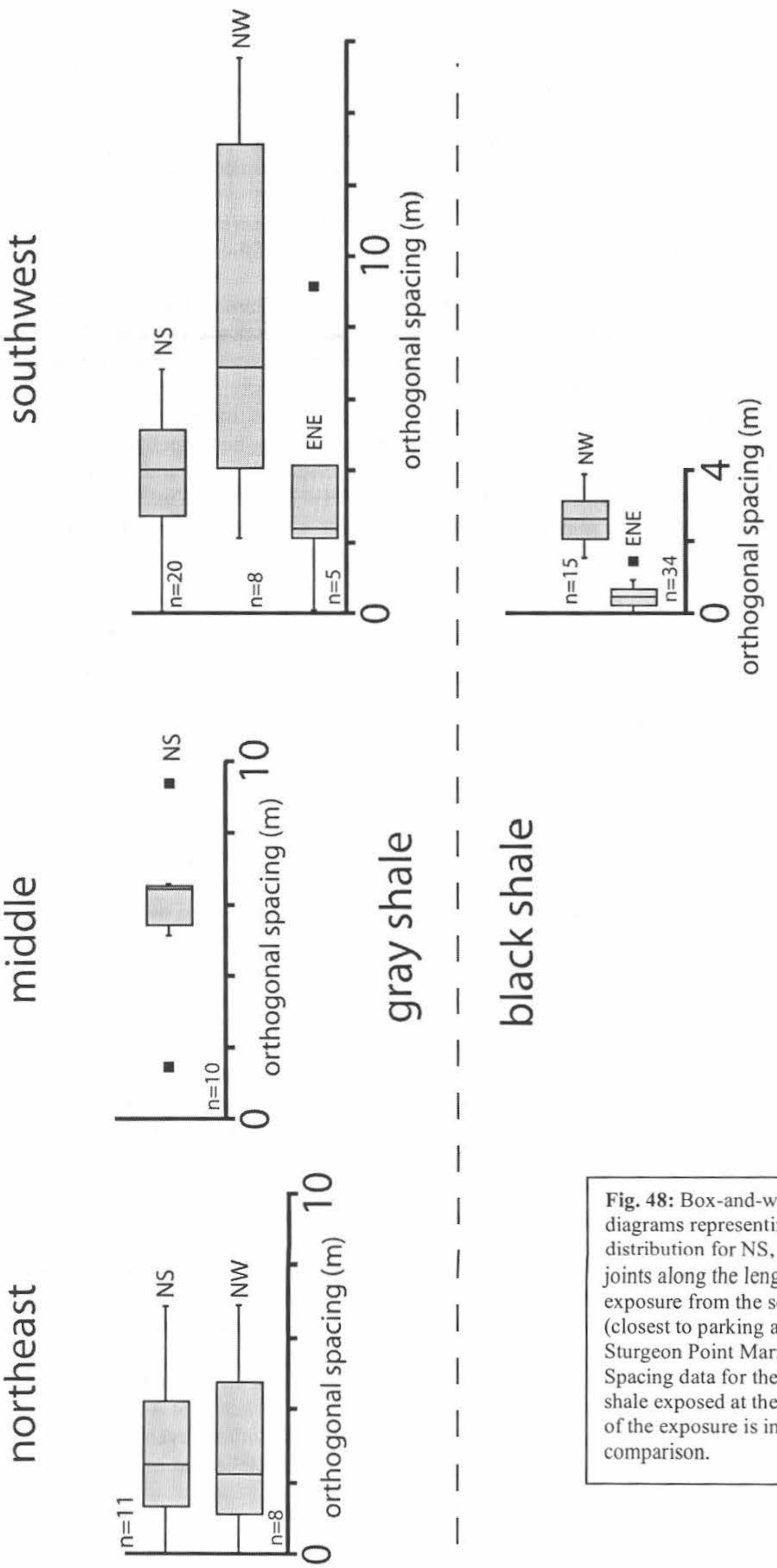


Fig. 48: Box-and-whisker diagrams representing joint spacing distribution for NS, NW, and ENE joints along the length of the Stop 3 exposure from the southwest (closest to parking area at the Sturgeon Point Marina) to the east. Spacing data for the lower black shale exposed at the southwest end of the exposure is included for comparison.

well-defined chronology of the joints (Engelder and Geiser, 1980; Lash et al., 2004) and the fact that the joints provide information regarding the orientation of the remote or lithospheric (Engelder and Whitaker, 2006) stress field that was responsible for their formation. There are likely subtle differences in rock properties between the black and gray shale that perhaps played some ancillary role in the jointing of these rocks. We suggest that the black shale, principally by virtue of its greater amount of ductile organic matter, was not as stiff (i.e., lower modulus) as the gray shale, a view at odds with the more common thinking regarding heavily fractured black shale. Accordingly, gray shale, as a consequence of its higher porosity and locally greater abundance of calcite cement (e.g., Lash, 2006b) would have had the higher modulus. Joints driven by fluid decompression, natural hydraulic fractures, form at great burial depth and, therefore, under high confining pressure. All else being equal, it would have been more difficult to initiate open mode cracks in higher modulus organic-lean gray shale under these conditions. Moreover, movement of the black shale into the oil window and resultant transformation of kerogen to hydrocarbons would more readily create a negative effective stress necessary for the propagation of natural hydraulic fractures, their growth direction being controlled by the remote stress field. NW-trending joints were produced by an Alleghanian remote stress field that was responsible for emplacement of the Blue Ridge-Piedmont Megathrust sheet; ENE joints, however, reflect a longer-standing lithospheric scale stress field related to continued oblique approach of the African portion of Gondwana against the Laurentian plate boundary (Engelder and Whitaker, 2006).

The remaining question, then, relates to the preferential formation of NS joints in the gray shale as observed at this exposure. Passage of the peripheral bulge across the Appalachian Basin at the onset of the Alleghanian orogeny and related uplift of the Upper Devonian shale-dominated succession of the Lake Erie District subjected these rocks to thermoelastic contraction. The higher modulus rock - the gray shale – would have carried a higher tensile stress and, therefore, would have been closer to the negative effective stress necessary to initiate joints. The inferred contrast in modulus between black and gray shale may explain why the NS joints terminate at gray shale-black shale contacts; i.e., the lower modulus black shale carried a lower tensile stress than the gray shale. However, the highest modulus rocks in the Rhinestreet shale were the carbonate concretions. As described earlier, concretions remained in compression during catagenesis so that fluid-driven NW and ENE joints were unable to penetrate them any more than a cm or so. However, during thermoelastic contraction, the concretions carried a high level of tensile stress (higher than that of the gray shale and certainly much higher than that of the black shale) making them more likely to fail under true tension. Initiation of the NS joints in the diagenetic carbonate was followed by their propagation through the host gray shale aided by a modest pore pressure.

In sum, this exposure, like others in this region of the Lake Erie District, provide evidence that systematic joints carried by the Upper Devonian shale succession resulted from two different loading mechanisms before post-Alleghanian uplift of the Appalachian foreland. However, we are investigating the intriguing possibility that an ENE-trending set of joints, some of which cut diagenetic carbonate, formed as a consequence of the uplift of the Appalachian Plateau. These joints appear to have been driven by thermoelastic contraction under a remote stress field perhaps induced by spreading at the Mid-Atlantic Ridge; thus their ENE strike.

II. This exposure also allows for the differentiation of concretionary carbonate found in gray shale from concretions common to black shale (spectacular examples of this latter type will be seen at Stop 4). The former are defined by high aspect ratios (they may better be described as lenticular diagenetic limestone) and lack obvious internal structures, principally layering. The scraggy layers are an example of this type of diagenetic carbonate. Carbonate concretions in the black shale have lower aspect ratios and display nicely preserved layers of the host organic-rich sediment. The above distinctions is readily made at this exposure.

Numerous small, internally laminated concretions can be observed in black shale within a meter or so of the upper contact of the gray shale interval, especially along the eastern stretch of the exposure. These concretions lend themselves to compaction strain analysis as described earlier (Fig. 49). Compaction strain is determined by measuring a stratigraphic interval within the concretion, T_i, and that same interval traced into the host shale but far enough beyond the strain shadow to provide a full measure of compaction strain, T_o (Fig. 49). The following equation is used to calculate compaction strain, ε_3:

$$\varepsilon_3 = \frac{T_i - T_o}{T_i}.$$

Estimated compaction strain of six concretions at this exposure is 60.8%, ~ 9% higher than the average for the Rhinestreet shale as discussed earlier. We believe that this reflects concretionary growth in a stratigraphic horizon that had not compacted to the extent that most of the unit did before the onset of normal compaction (see earlier discussion).

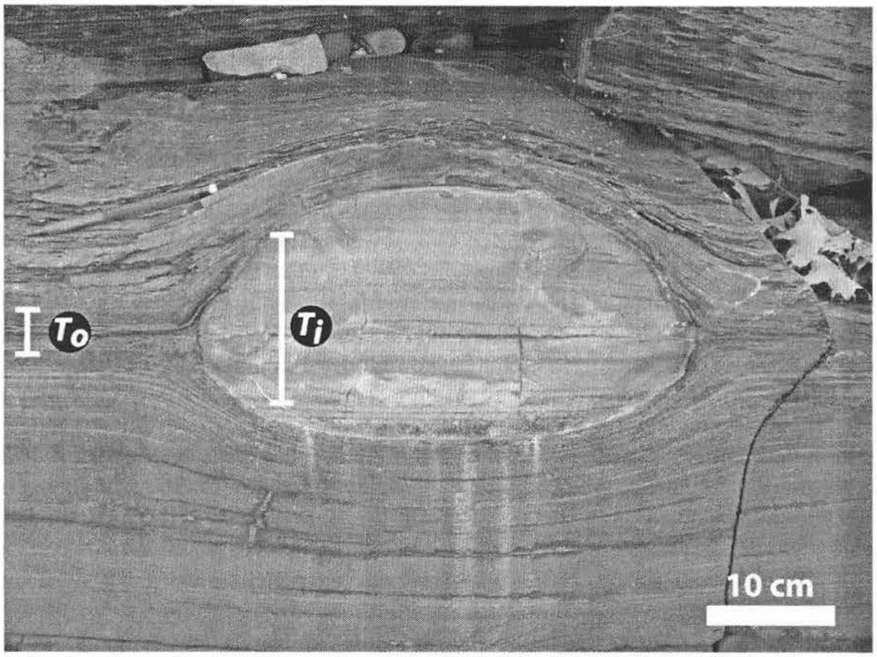


Fig. 49: Ellipsoidal carbonate concretion in Rhinestreet black shale at Stop 3. Refer to text for discussion.

III. Moving SW to NE along the beach, we observed the thickness of the gray shale interval diminish from ~2.75 m to ~ 0.8 m; farther along, though the thickness increases to at least 1.5 m. We observed no evidence of faulting that might account for the described change in thickness. Perhaps we are looking at the effects of differential compaction (e.g., Conybeare, 1967; Finely, 1990; Ryder, 1996). Indeed, the local presence of flame-like structures at the top of the gray shale interval suggests that these deposits were rather mobile and perhaps characterized by moderately high pore pressures, as suggested by the NS joints. Indeed, the relatively low density of NS systematic joints in the middle interval of this exposure where the gray shale is thinnest may reflect the effects of differential dewatering and resultant differential compaction. Preferential dewatering of the gray shale could account for (1) the reduced thickness of the gray shale at this location of the exposure and (2) a reduction in pore pressure, which would not have favored the generation of NS-trending joints in these rocks. Clearly, much more work is needed to elucidate this hypothesis.

Total Miles	Miles from Last Point	Route Description
27.9	0.0	Depart the marina onto Sturgeon Point Road.
28.8	0.9	Turn left (east) onto Lake Shore Road.
30.9	2.1	Intersection of Sweetland Road and Lake Shore Road.

31.0	0.1	Cross Pike Creek.
32.9	1.9	Turn right onto South Creek Road.
33.2	0.3	Intersection of South Creek Road and Route 5. Remain on South Creek Road.
34.1	0.9	Pass under railroad bridge.
35.0	0.9	Intersection of South Creek Road and Route 20. Turn left (east) onto Route 20.
36	1.0	Turn right (south) onto Lakeview Road.
36.4	0.4	Crossing over the New York State Thruway.
37.1	0.7	Turn right onto an unnamed road.
37.2	0.1	Turn right (west) at the stop sign onto North Creek Road.
37.8	0.6	Pull off onto the wide shoulder on the right. Cross North Creek Road and take the path down to Eighteenmile Creek.

STOP 4: **EIGHTEENMILE CREEK (main branch)**

Significance: I) the contact of the Rhinestreet shale and overlying Angola gray shale;
II) general characteristics of the carbonate concretions.

I. Follow the creek upstream (to the left) to where the east branch of Eighteenmile Creek enters the main creek. Cross the east branch just upstream from this point and follow a trail through the woods for roughly several hundred meters or until the path approaches the main branch. Cross the creek and walk to where you can view the cliff exposure across the creek. The middle concretion horizon will be visible a meter or so above the waterline. The upper ~21 m of the Rhinestreet shale and lower few meters of the Angola shale are exposed in the cliff (see Fig. 1). The Rhinestreet-Angola contact is transitional and marked by an increase in the frequency of less resistant (to the effects of weathering) gray shale layers, which can be observed both in the field at this stop and in well logs (Figs. 1 and 50). Luther (1903) suggested that the contact be placed at an easily recognizable discontinuous nodular limestone bed locally as thick as ~ 30 cm, the upper scraggy layer (see Fig. 1), which can be seen near the top of the exposure. The upper scraggy layer can also be recognized in some gamma-ray logs by a sharp deflection to the left (e.g., Fig. 20).

II) Carbonate concretionary horizons like the one exposed here are conspicuous in Middle and Upper Devonian black shale units of central and western New York state (Dix and Mullins, 1987; Lash and Blood, 2004a,b), southwestern Ontario (Daly, 1900; Coniglio and Cameron, 1990) and Ohio (Clifton, 1957; Criss et al., 1988) suggesting that diagenetic conditions favorable to the formation of these features existed regionally. Concretions are especially useful in the analysis of the early post-depositional history of encapsulating shale, for not only can they preserve the depositional texture of the host sediment (e.g., Woodland, 1984; Duck, 1990), but they carry a record of the evolution of pore fluids carried by the sediment (e.g., Hudson, 1978; Coniglio and Cameron, 1990). Details of our work on the carbonate concretions of the Rhinestreet shale can be found in Lash and Blood (2004a,b) and are presented in an abbreviated format above.

The concretions visible here are among the most spectacular in the world. They are similar to those described from such well known units as the Kimmeridge Clay (Astin and Scotchman, 1988) and the Jet Rock (Hallam, 1962). Close inspection of the exposed Rhinestreet concretions reveals their laminated

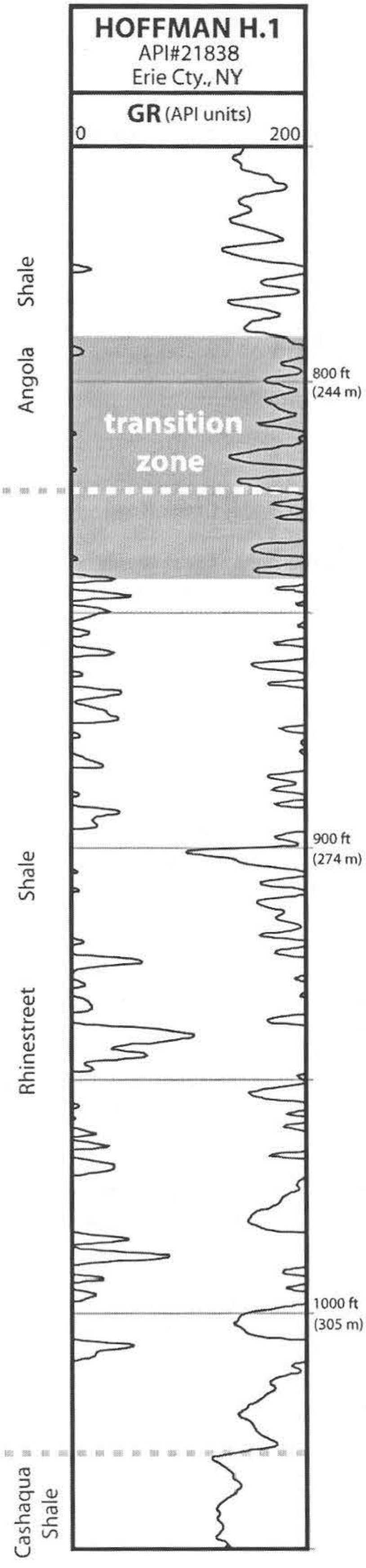


Fig. 50: Gamma-ray log from a gas well in Erie County showing the transitional contact of the Rhinestreet shale and overlying Angola shale. Note the frequency of gray (organic lean) shale-black shale alternations in the transition zone. The contact (dashed line) is based on bulk density (the upper contact of the Rhinestreet is arbitrarily set at that level where bulk density > 2.55 gr/cm^3; Lash, 2006a).

nature and presence of septaria. Our principal interest in the concretions has been their use as compaction strain indicators. However, before they could be used in this way, we had to understand their mode of origin and depth of growth, which was considered earlier. Most carbonate concretions of the Rhinestreet shale are found in three stratigraphically confined but laterally persistent horizons (see Fig. 1) – a lower horizon ~ 6 m above the base of the unit, a middle interval slightly more than halfway up the Rhinestreet (Stop 4), and an upper, less well defined, horizon ~ 43 m above the base of the unit. Most concretions are oblate ellipsoids with maximum diameters and thicknesses ranging up to 2.7 m and 1.1 m, respectively (Lash and Blood, 2004b); coalesced concretions have also been observed and are locally abundant. We carried out a detailed analysis of the dimensions and general shapes of concretions of the lower and middle horizons (Fig. 51). The dimensional data for both horizons reveals an especially intriguing feature; the maximum vertical dimension of the bulk of the concretions is ~ 80 cm (Fig. 51). However, the scatter of data from the middle horizon suggests that some concretions stopped growing in the vertical dimension at ~ 80 cm but continued to form laterally (Fig. 51). This suggests that the middle concretion horizon remained in the zone of anaerobic methane oxidation (AMO) longer than the lower concretion horizon.

Crucial to precipitation of diagenetic carbonate by the AMO mechanism discussed earlier is a marked reduction in sedimentation/burial, which would stabilize the sediment in the AMO zone thereby enabling protracted carbonate precipitation in that horizon (Raiswell, 1987, 1988). Two observations considered earlier provide evidence of a marked reduction in burial rate during concretion growth: (1) little or no change in carbonate volume percentage outward from the centers of studied concretions and (2) negligible variations in laminae thickness across concretions (Lash and Blood, 2004b).

The middle and lower concretion horizons differ in a fundamental way that yields indirect information regarding such parameters as the *magnitude* and *duration* of the change in sedimentation/burial rate. The middle horizon, 1 to 2.5 m-thick, contains concretions distinctly larger in the horizontal dimension than are concretions of the roughly 2.5-m-thick lower horizon (Fig. 51). The thickness of a concretion horizon is interpreted to be proportional to the change in sedimentation/burial rate; i.e., a marked reduction in burial rate yields the minimum vertical range of carbonate precipitation because the zone of AMO in the sediment stabilized (Raiswell, 1987, 1988). The more oblate, occasionally flat-topped, concretions in the middle horizon, then, likely reflect the restriction of carbonate precipitation to a relatively narrow, fixed zone defined by the upper limit of methane and CO_3^{2-} diffusion and the maximum depth to which seawater sulfate can diffuse (Raiswell, 1988). Thus, the generally larger horizontal dimensions of concretions of the middle horizon indicate that the break in sedimentation persisted longer at this time than during formation of the lower concretionary horizon.

The middle Rhinestreet concretion horizon, by virtue of its limited thickness and the geometry of its concretions, records a reduction in sedimentation/burial rate of greater magnitude than that which led to formation of the thicker lower concretionary interval (Lash and Blood, 2004b). However, the thickness and geometry of the scraggy layers likely reflects an even greater duration of the break in sedimentation/burial and, therefore, a protracted period of time that isotopically light methane and ^{13}C-enriched dissolved carbonate were delivered to the stationary AMO zone (Raiswell, 1987, 1988). That is, the near cessation of sedimentation and, more importantly, subsidence, resulted in confinement of carbonate precipitation to a narrow, fixed zone defined by the upper limit of methane diffusion and the maximum depth to which seawater sulfate could diffuse (Raiswell, 1988). Because the events that contributed to the reduction in sedimentation/burial rate may have been basinwide, concretionary horizons may serve as time lines in establishing regional time-stratigraphic relationships.

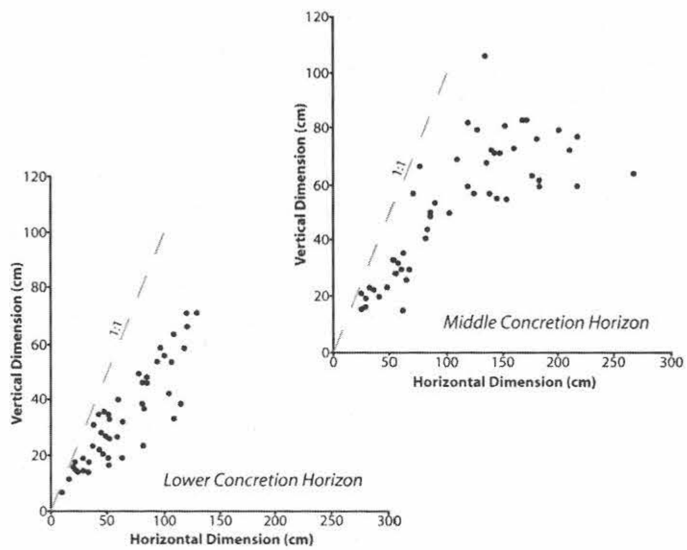


Fig. 51: Horizontal versus vertical dimensions of carbonate concretions of the lower (n = 50) and middle (n = 52) concretion horizons of the Rhinestreet shale.

END OF FIELDTRIP – return to vehicles

Total Miles	Miles from Last Point	Route Description
37.8	0.0	Continue along North Creek Road to Route 20.
39.0	1.2	Turn left (west) onto Route 20.
39.4	0.4	Turn right onto South Creek Road.
41.2	1.8	Turn right onto Route 5 and return to Adams Mark in Buffalo.

SILURIAN SEQUENCE STRATIGRAPHY, EVENTS, AND PALEOENVIRONMENTS ALONG THE CRATONIC MARGIN OF THE APPALACHIAN FORELAND

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INTRODUCTION

Silurian rocks of the Appalachian Basin and the North American mid-continent platform provide an excellent suite of strata for application of sequence, and event stratigraphic approaches. The strata are well exposed and display marked vertical changes in facies, commonly associated with distinctive condensed beds and/or discontinuities. The Niagara Escarpment in western New York and the Niagara Peninsula of Ontario represents a key reference area for the North American Silurian System. Indeed, the term "Niagaran," which has been variably applied to the middle or lower and middle portions of the Silurian, is still commonly used in North America. These rocks have been well documented by many researchers, beginning in the mid 1800s (Hall, 1852; Grabau, 1901; Foerste, 1906, 1935;Williams, 1919; Gillette, 1947; Bolton, 1957; Zenger, 1965, 1971; Sanford, 1969; Martini, 1971; Rickard, 1975; Duke, 1987; Duke and Brusse, 1987; Duke and Fawsett, 1987; Brett, 1983 a, b, 1985; Brett et al., 1990, 1995, 1998; Goodman and Brett, 1994; LoDuca and Brett, 1994).

Previous reports on the Silurian in the Niagara region have focused on identification of depositional sequences and depositional/faunal events. A theme of this guide is the documentation and interpretation of Silurian facies, events, and sequences along the northwestern rim of the Appalachian Basin. As such, the discussion is not restricted to the sequences and facies of western New York. A remaining research agenda is to trace these sequences into other regions, especially the mid-continent. Ongoing research along the Cincinnati Arch, in Ohio, Indiana, and Kentucky has revealed striking similarities of sequences and events among these seemingly disparate areas, some 500 to 700 km southwest of the main Niagara Escarpment (Fig. 1). In this paper we review and update information on depositional sequences and events in western New York and Ontario and then discuss their lateral relationships with those of the Cincinnati Arch. A companion field guide will highlight the Silurian strata of the Cincinnati Arch region.
The large-scale ("third order") sequences described herein are sharply bounded stratal packages, ranging from less than a meter (where partially truncated or condensed) to about 50 m in thickness. In each case, the sharp basal contacts, when traced in an upramp direction, change from facies dislocations (abrupt shallow over deeper facies change) to erosional surfaces that demonstrably truncate underlying strata. Most sequences recognized herein, display a generally deepening-shallowing
pattern.

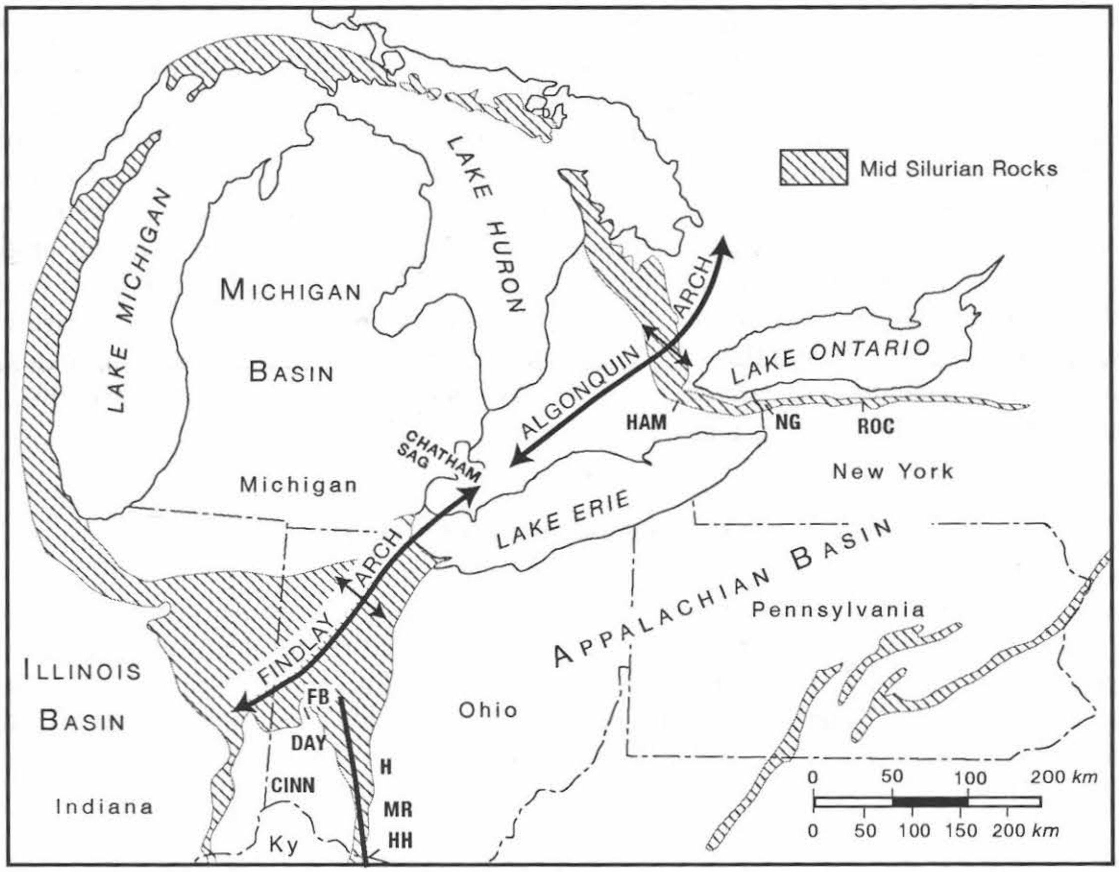


Figure 1. Geographic setting of Silurian in eastern North America. A) Map showing geomorphic features of eastern North America and outcrop belt of the Silurian; abbreviations: DAY: Dayton, Ohio; CINN: Cincinnati, Ohio; FB: FairbornQuarry near Dayton, Ohio; HAM: Hamilton, ONT; H: roadcut on Rte. 62 at Hillsboro, Ohio; HH: Cut on AA Highway at Herron Hill, Kentucky; MR: cut on US Rte. 32 at Measley Ridge, near Peebles, Ohio; NG: Niagara Gorge, NY, ONT; ROCH: Rochester, NY. Base map modified from Telford (1978).

However, in detail, the larger-scale sequences are composites, divisible into smaller, fourth-order sequences.

With its renewed emphasis on through-going discontinuities and condensed beds, the sequence stratigraphic approach, has encouraged a broader, more regional view of stratigraphy, and an attempt to understand the genetic significance of particular beds and surfaces. To some degree it vindicates the earlier "layer cake stratigraphy" approach. Sequence stratigraphy, originally developed from remote seismic studies of passive margin sediment wedges (Vail et al., 1977, 1991; Wilgus et al., 1988; Coe, 2003; Catuneanu, 2002), is now being applied at the outcrop scale to diverse depositional settings such as the Taconic foreland basin (Brett et al., 1990 a, b; Witzke et al., 1996; Brett et al., 1998; Brett and Ray, 2006). Many distinct horizons in the local stratigraphic record are interpretable as sequence boundaries, drowning surfaces, and forced regression surfaces. Moreover, a number of phenomena, which occur non-randomly in the geologic record, from phosphatic nodule horizons to reefs, fit in predictable ways into depositional sequences.

A second theme of this article is the recognition of widespread events, such as storm deposits, (tempestites), rapidly buried fossil horizons (obrution deposits), and even seismically deformed beds (seismites). The Silurian Period has typically been considered a tectonically quiescent time; however, growing evidence from eastern North America suggests the Silurian was more dynamic then previously considered. For example, sedimentology and stratigraphic mapping of Silurian deformed beds in eastern North America demonstrate that these event beds are extremely widespread and that their component sediment layers were not deformed during initial deposition, but slightly later, during shallow burial, likely the result of shaking induced by large-scale earthquakes (McLaughlin and Brett, 2006). The distribution of deformed beds, together with increased subsidence, clastic influx, and K-bentonite horizons, provides a meter of intensity and timing of pulses of Silurian orogenesis. A variety of such catastrophic events is recorded in the Silurian rocks of the Niagara region and will be examined on this trip.

REGIONAL GEOLOGICAL SETTING

Silurian strata of the Niagara Peninsula-western New York and Cincinnati Arch were deposited along the carbonate (cratonic) margin of the Taconic and Salinic foreland basins (Figs. 1, 2). Early Silurian Strata of the Niagara Peninsula were deposited in generally shallow, subtropical epeiric seas, situated 25-30° south of the paleoequator (Witzke, 1990; Fig. 1B). The axis of the Cincinnati Arch occupied a similar latitude and was oriented roughly parallel to the equator, resulting in broad, strike-parallel facies belts.

Siliciclastic sediments were derived from eastern and, possibly, northeastern (present-day orientations) source terrains that were uplifted during the Late Ordovician to medial Silurian orogenies. In addition to the major Ordovician Blountian and Taconian (Vermontian) tectophases, Ettensohn and Brett (2002) identified a third, Early Silurian (Median) tectophase of the Taconic Orogeny.

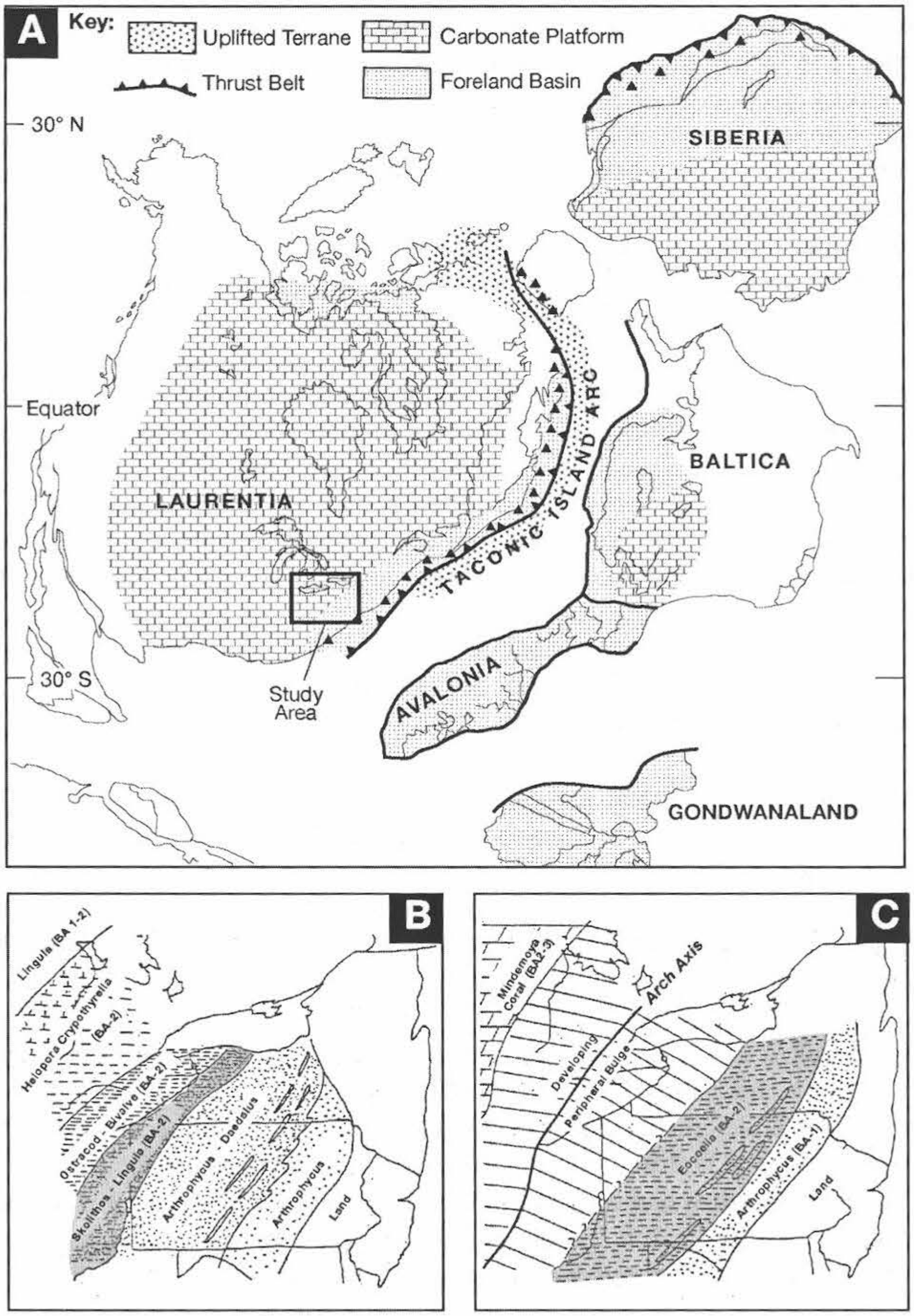


Figure 2. Paleogeography for Early Silurian. A) Reconstruction of Laurentia (ancestral North America) and adjacent paleocontinents during Early Silurian time. Note position of study area, shown with box and of the Taconic Arch and peripheral foreland basin. Modified from Scotese (1990). B,C) Paleogeographic reconstructions of New York and adjacent areas for two different times within the Early Silurian. B) Rhudannian; highstand of Sequence I. Medina Group; Castanea Member of Tuscarora Formation in Pennsylvania; note position of basin center in southern Ontario. C) Mid Telychian; highstand of sequence III; Sauquoit Shale-Otsquago red sandstone in central New York; upper Rose Hill Shale in Pennsylvania; lower Crab Orchard Shale in Ohio-Kentucky. Note development of regional uplift (forebulge) along the Algonquin Arch.

This phase of uplift produced a new episode of molasse deposition: the Tuscarora-Medina clastic wedge, centered in the southern and central Appalachians. This tectonism increased the gradients of source streams, producing an influx of coarse, quartz rich gravels and sands perhaps from reworking of vein quartz and older sandstones in the rejuvenated Taconic orogen. Renewed uplift of tectonic terranes may have occurred during the medial- and Late Silurian in the Salinic Disturbance (see Ettensohn and Brett, 1998). These tectophases produced an initial pulse of quartz sands, upper Shawangunk-Keefer-Herkimer conglomerates and sandstones in the proximal foreland basin, followed by a major influx of primarily fine-grained siliciclastics of the Vernon-Bloomsburg "delta", predominantly red molasse.

During the Early Silurian, Medina Group siliciclastics accumulated in non-marine to shallow marine environments in western New York and marine shales extended with little or no break into the region of the Michigan Basin and northeastern Ohio (Fig. 2B). However, by medial Silurian time (middle Llandovery), a narrow carbonate platform appears to have existed in the area around Hamilton, Ontario northwest into the Bruce Peninsula and extending southwest into central and southern Ohio and Kentucky (Algonquin-Findlay Arch; Fig. 2C). This platform was a region of shallow, epeiric seas, with lesser siliciclastic input than the foreland basin to the east punctuated by relatively thin packages of skeletal sand and chemical sediments. This arch formed a partial to nearly complete barrier between the Appalachian and Michigan Basins during the late Llandovery and Wenlock time (Fig. 2C).

The arch was probably emergent at times during the late Llandovery and mid Wenlock when relative base level drops produced major unconformities within the mixed carbonate-siliciclastic succession. During late Wenlock to Ludlow time, however, the Algonquin Arch, appears to have subsided such that the area between Hamilton, Ontario and the northern Bruce Peninsula was the locus of <u>deeper</u> water environments than areas to the southeast or northwest. During this time reefy carbonates were widespread in the Appalachian Basin.

As a broad generalization, Silurian strata in the northern Appalachian basin display a trend from a) siliciclastic dominated units in the Early Silurian (Alexandrian or early Llandovery) Medina Group, to b) mixed siliciclastics and carbonates in the medial Silurian (lower Niagaran, upper Llandovery-Wenlock) Clinton Group, c) dolomitic carbonates in the Lockport Group (upper Niagaran; Wenlock-Ludlow), and d) mudrocks, dolostones and evaporites in the Upper Silurian (Cayugan, upper Ludlow-Pridoli) Salina-Bertie Groups (Fig. 3). These patterns reflect the two major tectonic pulses: late Taconic Media tectophase in earliest Silurian and the medial Silurian Salinic Orogeny; these times were dominated by prograding clastic wedges and westward migration of the foreland basin depocenter (Fig. 3; see also Fig. 15; Goodman and Brett, 1994; Ettensohn and Brett, 1998). Intervening times of quiescence were marked by predominant carbonate deposition and eastward backstepping of the basin depocenters (Ettensohn and Brett, 1998).

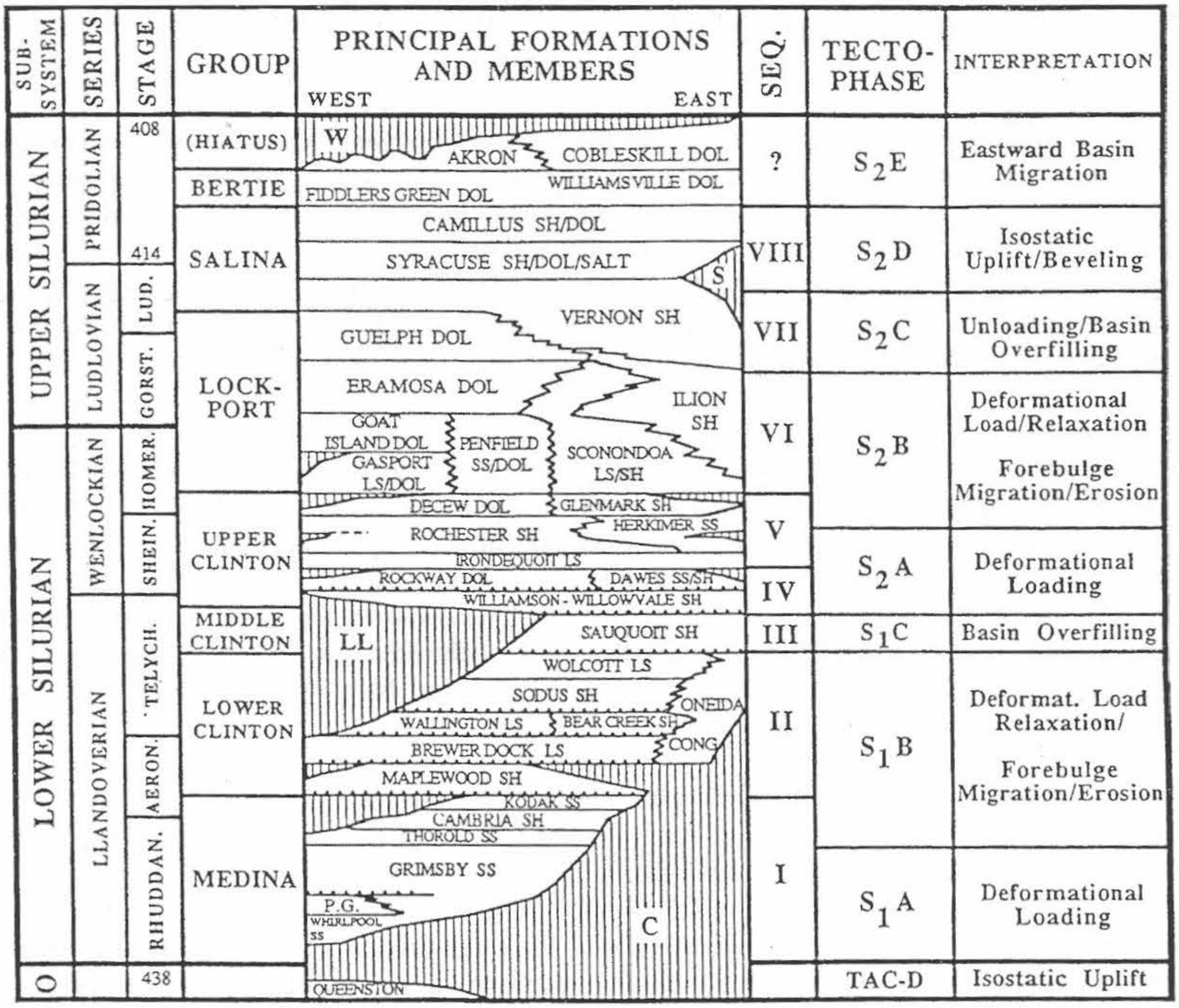


Figure 3. Generalized chronostratigraphic chart for the Silurian System along the east-west outcropbelt from Niagara County to Utica, New York; middle column shows subdivision of stratigraphy into depositional sequences(I-VIII) each marked by a basal sequence bounding unconformity; right two columns show inferred tectophases of latest Taconic (S1) and Salinic (S2) orogenies. Dots indicate phosphatic/ironstone rich condensed beds. Major unconformities marked by letters: C: Cherokee; LL: Late Llandovery; S: Salinic; W: Wallbridge.

SEQUENCE STRATIGRAPHY AND LATERAL CHANGES OF SILURIAN SEQUENCES IN SOUTHERN ONTARIO-WESTERN NEW YORK, COMPARSIONS TO OHIO AND KENTUCKY

The Silurian strata of western New York and the adjacent Niagara Peninsula of Ontario have been broadly subdivided into groups that correspond in part to large-scale (third order of sequence stratigraphers, see Van Wagoner et al., 1988; Vail et al. 1991; Catuneanu, 2002) depositional sequences (Figs. 3, 4); the Clinton Group, however, is divisible into at least three sequences, which correspond roughly to Gillette's (1947) "lower", "middle", and "upper" Clinton divisions. The larger sequences are also divisible into smaller (fourth-order) sequence-like units; we have termed "subsequences" (see Brett et al., 1990, 1995, 1998).

Silurian sequences are bounded by unconformities, three of which, the I-II, II-IV, and V-VI boundaries, are regionally angular (Figs. 3, 4). The magnitude of these three unconformities (i.e., extent of beveling on the erosion surface) increases westward along the Niagara Escarpment. These surfaces appear to have been accentuated by uplift along the "Algonquin Arch", probably an intermittently active forebulge (Figs. 3, 8). The I-II and II-IV boundaries are merged west of St. Catharines, Ontario, forming a compound unconformity (Figs. 3, 4). A very similar pattern is seen to the southwest along the flank of the Findlay Arch near Dayton, Ohio, where unconformities at the top of the Brassfield Formation (base of Plum Creek Shale and below the Dayton Dolostone merge westward (see Fig. 9).

Conversely, the basal sequence I unconformity (Cherokee unconformity and base of Silurian system) decreases westward (Fig. 4). In east central New York State the entire Queenston Formation is truncated and in places basal Silurian sandstones rest unconformably on Maysvillian or Edenian shales of the Frankfort or Schenectady formations with the magnitude of unconformity increasing progressively eastward. This regionally angular, eastward-opening unconformity apparently represents late Taconic or perhaps isostatic uplift of proximal flysch-to-molasse successions (Goodman and Brett, 1995; Ettensohn and Brett, 2002).

Varying east to west facies changes within each of the sequences along the Niagara Escarpment reflect differential subsidence and elevation of the Algonquin-Findley Arch and adjoining basins, as noted in the following sections. In each, the stratal unit name is followed by series/stage assignment based on biostratigraphy (see Brett et al. 1990, 1998, for details).

Sequence I: Medina Group, Lower Llandovery (Rhuddanian)

The stratigraphically lowest Silurian interval (S-I) is the predominantly siliciclastic Medina Group (Cataract Group of some Canadian authors). In west central New York the Medina Group contains, in ascending order: the Whirlpool Sandstone (2.5-4.5 m)

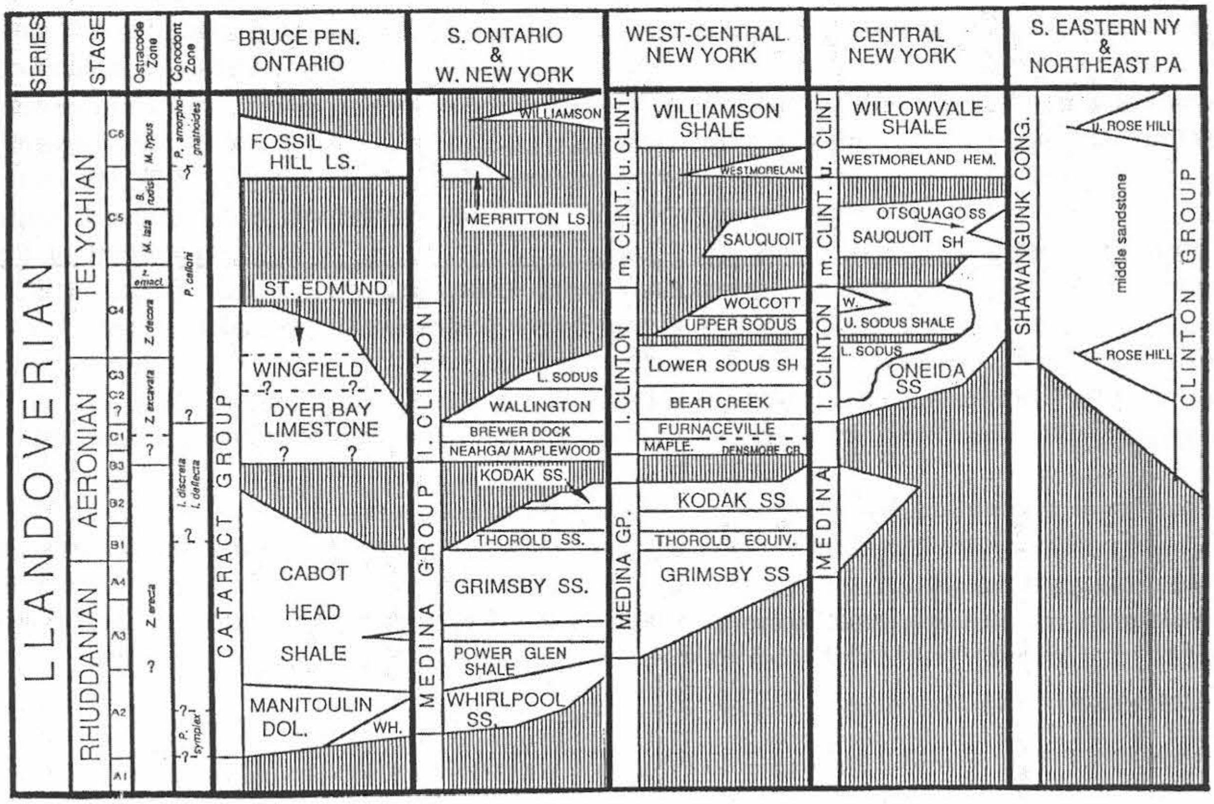


Figure 4. Correlation chart Lower Silurian (Llandovery) stratigraphic units in Ontario and New York State; columns arranged approximately west- to-east. Abbreviations: BC: Bear Creek Shale; Dol: Dolostone; Sh: Shale; SS: Sandstone; WH: Whirlpool Sandstone; WO: Wolcott Limestone.

whitish gray, trough cross bedded quartz arenite, Power Glen Shale (10-15 m) dark gray shale with thin sandstones and dolomitic limestones, Devils Hole Sandstone (2-3 m) whitish gray, phosphatic quartz arenite, Grimsby Formation, (12-15 m) maroon to green shale and reddish and white mottled sandstone, Thorold Sandstone (2-3 m) reddish to whitish gray, bioturbated quartz arenite with greenish gray sandy, bioturbated mudstones, Cambria Shale (0-3 m maroon shales and muddy sandstones, and Kodak Sandstone (0-3 m) whitish gray sandstones and greenish to maroon shales (Figs. 4, 5). The Medina represents a large scale depositional sequence with lowstand (non-marine) to transgressive (foreshore to shoreface) Whirlpool Sandstone (Middleton et al., 1991), overlain by maximally highstand Power Glen Shale (offshore marine muds), and later highstand (progradational shoreface and tidal flat) Devils Hole through Kodak strata. However, the interval is also divisible into smaller (4th order) subsequences at the bases of the Whirlpool, Devils Hole, Thorold, and Kodak transgressive quartz arenites (Figs. 4, 5).

The Medina Group exhibits relatively constant thickness along the Niagara Escarpment, but all of its component units show westward changes in facies corresponding to increasingly open, fully marine conditions (Figs. 4, 5). Thus, the lower, fluviatile (braided stream) Whirlpool Sandstone (Middleton et al. 1991) apparently pinches out (or changes facies to marine) to the north of Georgetown, Ontario (Rutka et al., 1991). The upper Whirlpool thickens and displays evidence of shoreface to shallow, sandy shelf deposition near Hamilton, and may be replaced laterally by the Manitoulin Formation open shelf carbonates (Fig. 5). Dark gray, sandy, sparsely fossiliferous Power Glen Shale in the Niagara region, grades westward near Hamilton into greenish gray shales with abundant bryzoan-rich carbonates, indicating open marine conditions. The Grimsby interval changes northwestward from red sandy mudstones and tidally influenced sandstones with a lingulid (BA-1) biofacies at Niagara Gorge (Martini, 1973; Duke and Fawsett, 1987) to red and green mudstone with only thin sandstones and hematitic bryzoan-rich limestones (BA-2-3) in the upper Cabot Head Formation (Duke and Brusse, 1987). Conversely, the Grimsby interval shows a return to marginal marine or non-marine red beds northward along the Bruce Peninsula (Fig. 2B). Thus, the Hamilton area was close to the deepest part of the foreland basin during Medina deposition (Fig. 2B; Ettensohn and Brett, 1998).

Westward erosion of the upper Medina Group below Sequence II beveled an upper fourth order cycle (Cambria-Kodak) between Rochester and Lockport, New York and culminated with removal of the Thorold Sandstone near Hamilton (Fig. 5). This pattern suggests an inversion of topography, with minor broad uplift on the Algonquin Arch region following Medina deposition.

The Medina is traceable in the subsurface of western New York, northwestern Pennsylvania and Ohio where its units are relatively well known as a result of oil and gas exploration. As a result the continuity of sandstone tongues, notably the Whirlpool, Devils Hole, and Thorold is demonstrable. As in Ontario he Power Glen and Grimsby appear to pass southwestward into a greenish gray to dusky reddish, shale-dominated succession of the Cabot Head Formation.

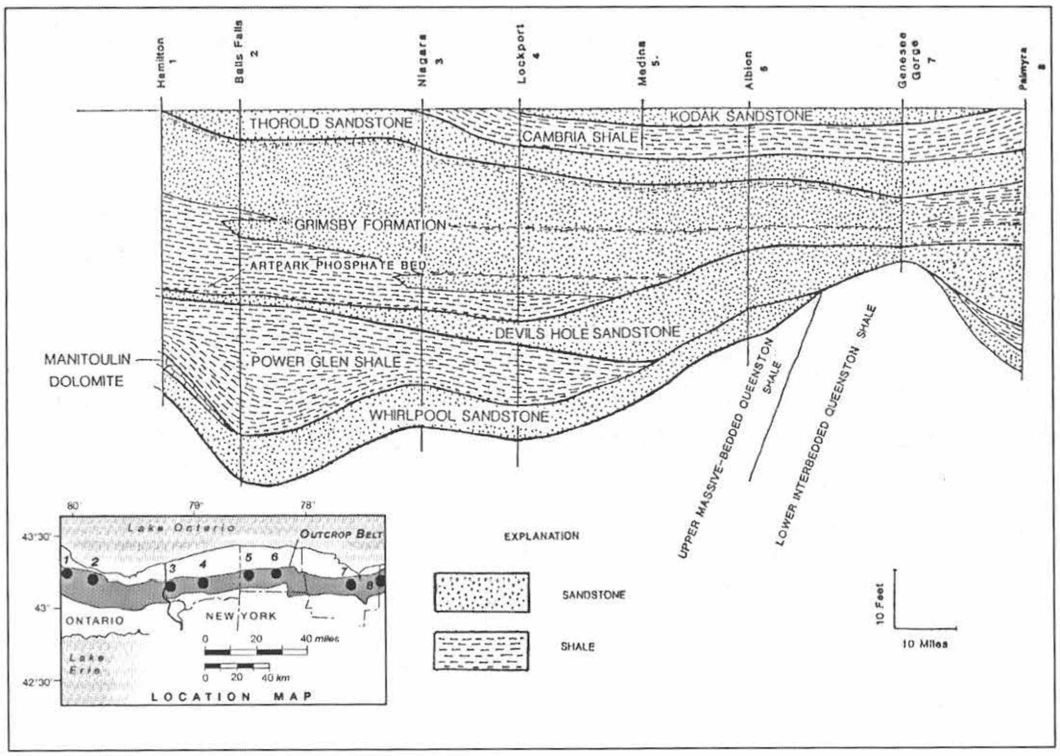


Figure 5. Regional cross section of Medina Group (Sequence 1) in western New York and southern Ontario.

In the outcrop area of southwestern Ohio the Medina interval is represented by the dolomitic carbonate-shale succession of the Brassfield Formation. This interval is divisible into informal members. A basal silty-sandy dolostone of the Belfast Member (0.5 to 2 m thick) may represent the same first 4[th] order sequence as the Whirlpool Formation of the western New York. A basal bioturbated, glauconitic sandy dolostone bed rests sharply on greenish to reddish silty mudstones of the Preachersville Member, Drakes Formation, representing distal marine facies of the Queenston formation. Hence, this contact represents the major Cherokee unconformity. In terms of the overall Medina sequence, this unit represents lowstand to early transgressive conditions. An overlying massive cherty unit carries much the same benthic fauna as the Manitoulin Dolostone in Ontario and probably represents the TST of the second major 4[th] order sequence. A higher shaly zone with localized bioherms in central Ohio may occupy the position of the Cabot Head Shale of Ontario. As noted above, there appears to be an association of biohermal development and transgression as the mounds protrude upward from an interval of skeletal grainstone and are surrounded by fine grained argillaceous carbonate and shale. As with the upper Cabot Head this mud is reddish suggests correlation with the lower Grimsby Formation. If so, the persistence of red coloration even into the carbonate dominated facies of central Ohio raises the intriguing issue of the significance of this coloration. Color in this case crosscuts facies as it definitely does in the Grimsby to Cabot Head transition in Ontario. Does the red Grimsby-upper Cabot Head reflect a distinctive weathering regime that produced large volumes of oxidized iron which were distributed over paraliac to offshore marine environments?

Clinton Group, Middle Llandovery (Aeronian) to Middle Wenlock

The Clinton Group consists of mixed carbonates and shales, representing offshore storm-influenced shelf environments, and was informally subdivided into lower, middle, and upper Clinton by Gillette (1947). This convention is adopted herein because two of the three divisions correspond to depositional sequences (Figs. 3, 4).

Sequence II: Lower Clinton Group, Middle Llandovery (Aeronian)

The lower Clinton (Sequence II) is very incomplete in western New York, consisting only of the Neahga Shale (0-2 m of greenish gray shale marked at the base by a phosphates dolostone) and Reynales Limestone (0-3 m of calcisiltite, nodular packstone and bryzoan-brachiopod-echinoderm grainstone, and minor shale) (Figs. 4, 6). To the northwest, in the Bruce Peninsula, up to 20 m of possibly correlative strata, the, Wingfield, St. Edmund formations, and lower Fossil Hill sharply overlie the Cabot Head Shale (Stott and Von Bitter, 1999; Fig. 4).
So little remains of Sequence II in the Niagara Peninsula that it is difficult to determine facies trends. However, facies changes in the Neahga and Reynales formations in western New York suggest westward deepening patterns (LoDuca and Brett, 1994).

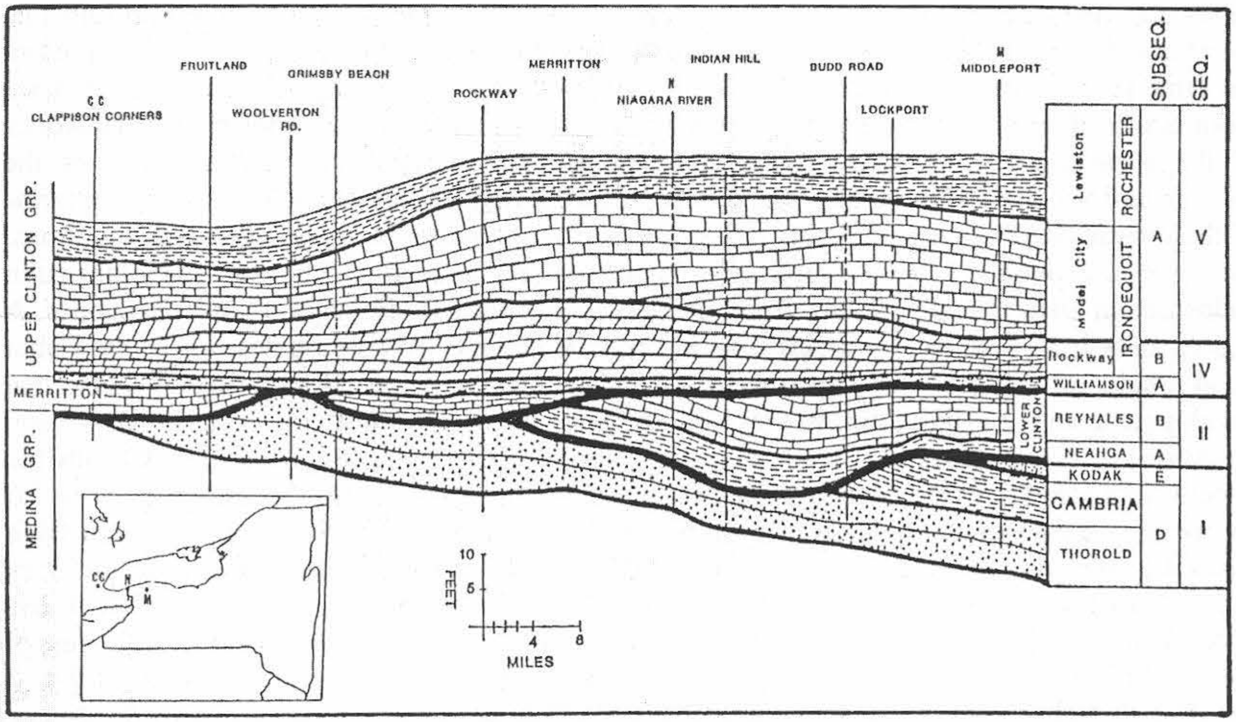


Figure 6. Regional cross section of upper Medina (Sequence I), lower Clinton (Sequence II), and upper Clinton (Sequences IV and lower part of V) through western New York and Ontario (see locations on inset map). Modified from Kilgour (1963). Strong vertical exaggeration to highlight sequence truncations.

In south central Ohio and into northern Kentucky, the position and approximate biostratigraphic equivalent of the Neahga Shale is occupied by the Plum Creek Shale (Figs. 7-10). This thin (0.5 to 2 m) package of greenish gray shale and thin limestones is locally highly fossiliferous with a fauna of tabulate corals, small solitary rugosans and atrypid brachiopods. The Plum Creek appears to be a more distal, offshore facies of the Neahga. A widespread oolitic/skeletal hematite bed (Rose Run ironstone of Foerste, 1906) marks the base of the Plum Creek may correlate with the Densmore Creek phosphate of the New York succession (Figs. 7, 10).

The Plum Creek is overlain, in turn by fossiliferous and locally hematitic limestone of the Oldham Formation, which is a correlative of the Reynales. In turn, the Oldham is overlain by up to 4 m of greenish gray slightly fossiliferous shale, the Lulbegrud Shale, which may be co-extensive with the Lower Sodus Shale of west central New York (Figs. 7-10).

Middle Clinton Unconformity: Upper Llandovery (Telychian)

Middle Clinton Group strata (Sequence III) are absent in the Niagara region, as in southern Ohio, and a major regionally angular unconformity separates the lower Clinton Reynales Formation from the overlying Sequence IV (Figs. 8, 10). A major change in depositional topography of the Appalachian foreland occurred during the mid Llandovery; throughout west-central New York State and Ontario the middle Clinton Group is missing and an erosion surface beneath late Llandovery (Telychian) strata truncates lower Clinton units in a westward direction along the outcrop belt (Lin and Brett 1988; Ettensohn and Brett, 1998; Fig. 2C, 8). This substantial

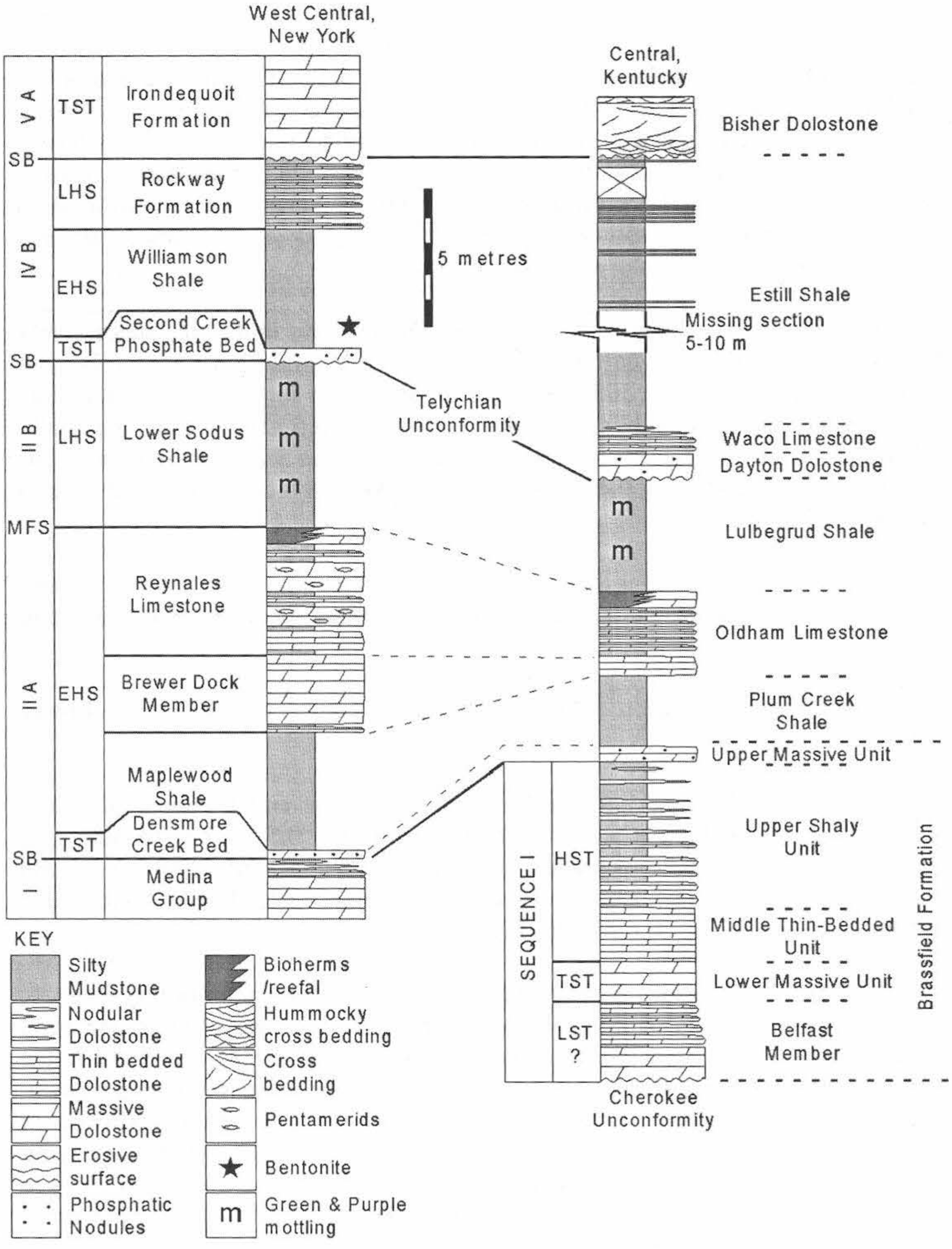


Figure 7. Comparison of Llandovery lithostratigraphic succession in west central New York (vicinity of Rochester, New York) and north-central Kentucky. Sequence stratigraphic abbreviations: EHS: early highstand systems tract; LHS: late highstand (or falling stage) systems tract ; SB: sequence boundary; TST: transgressive systems tract.

regionally angular unconformity suggests another period of broad regional uplift centered on the Algonquin Arch (Fig. 8). Stott and Von Bitter (1999) documented a complex pattern of local uplift/erosion and minor basins in the area of the southern Georgian Bay during this interval, suggesting reactivation of basement fault blocks in response to lithospheric flexure. Development of the unconformity also coincides with a shift in basin axis migration from eastward (Medina-middle Clinton) to westward (upper Clinton); it may signal renewed tectonic activity in the eastern hinterland (Ettensohn and Brett, 1998).

As in western New York, the Lulbegrud, Oldham and Plum Creek units are, successively truncated to the northwest onto the Findlay Arch in west-central Ohio, by an upper Telychian erosion surface (Horvath, 1964; Lukasik, 1987; Fig. 9). This truncation surface can be traced into central and southern Kentucky (McLaughlin, unpublished data) and reflects the same regionally angular unconformity as noted in western New York and Ontario. It indicates contemporaneous uplift along at least a 700 km trend along the Algonquin Arch. We suggest that this trend reflects activation of a forebulge.

It is still not entirely clear whether this feature formed in response to deformational load relaxation as interpreted in Figure 3,or actually reflects the initiation of a new tectophase associated with the Salinic Orogeny.

Sequence IV: Upper Middle Clinton Group, Upper Llandovery (Telychian)-Lower Wenlock

In western Ontario Sequence IV comprises the Merritton Dolostone and its apparent lateral equivalent, the upper Fossil Hill Formation in the Bruce Peninsula (0.5 to 5 m of dolomitic limestone with many corals and pentamerid brachiopods; Stott and Von Bitter, 1999), a very thin tongue of Williamson Shale (0-20 cm), and the Rockway Dolostone-Lions Head Member (3-4 m) of the upper Clinton Group (Figs. 3, 6;see also Fig. 22).

The thin, condensed Merritton Dolostone overlies the major mid Clinton unconformity (Figs. 6, 9); it is unknown in western New York State, although it is roughly equivalent in age (mid Telychian) to the Westmoreland Hematite of central New York (Fig. 8). In general, the Merritton displays a slight westward shallowing trend from glauconite-rich wackestones at St. Catharines to pentamerid- and coral-rich packstone (upper Fossil Hill Formation) northwest of Hamilton. A similar, though very gradual, westward facies change is seen in the overlying Rockway Dolostone, which becomes increasingly carbonate-rich from west central New York to southern Ontario. Biofacies change from *Clorinda*-dominated (BA-5) to *Costistricklandia*-dominated associations (BA-4) also suggest gradual westward shallowing. Locally, in Ontario a feather edge of the dark gray Williamson Shale, as demonstrated by acritarch assemblages (M. Miller, personal comm., 1986) intervenes between the Merritton and Rockway Formations. The Williamson and laterally equivalent Willowvale Formations thicken eastward to a maximum of over 30 m in central New York.

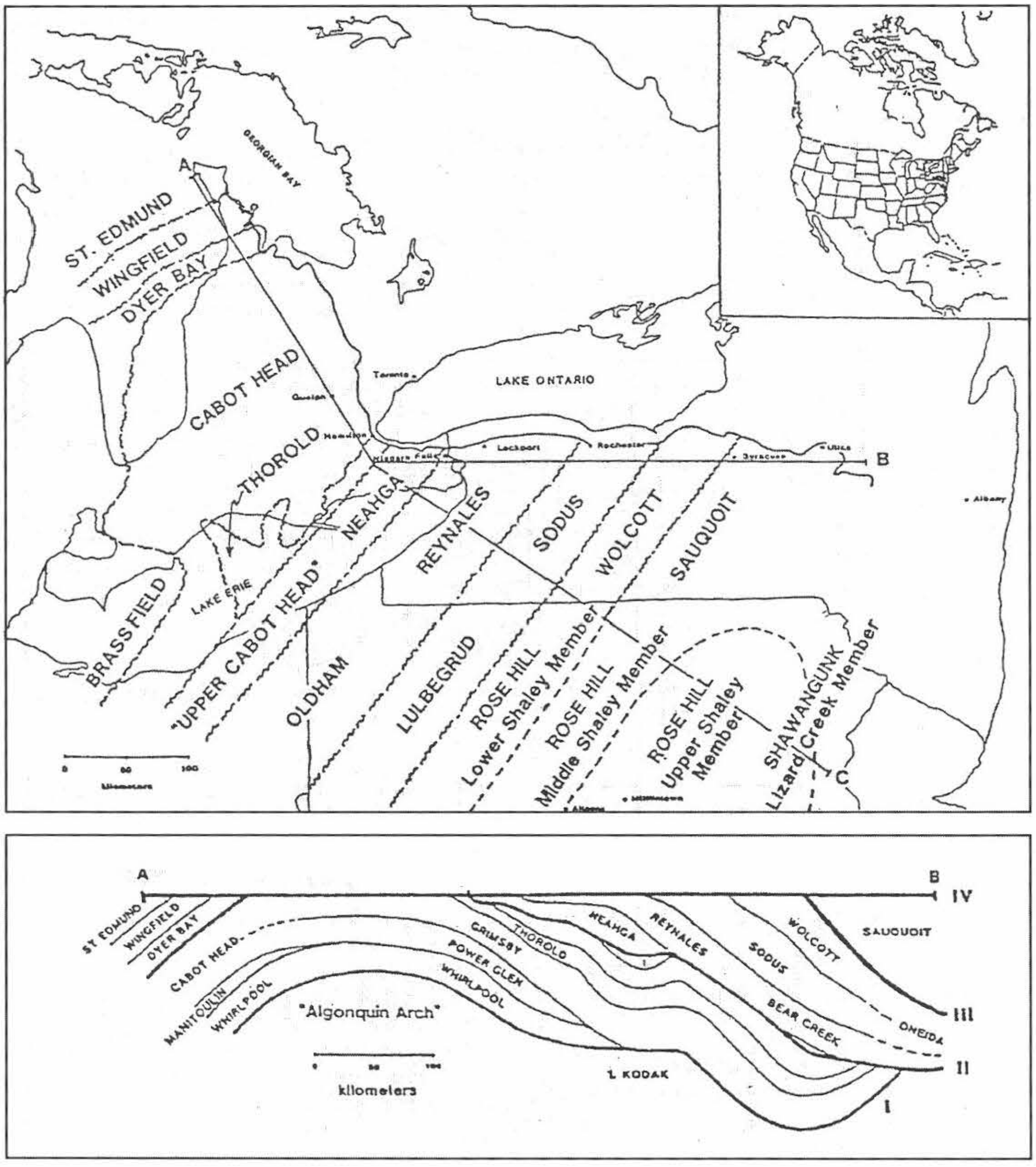


Figure 8. A) Subcrop map of strata beneath late Llandovery (S-IV) unconformity; inset shows general location. B) Hypothetical, vertically exaggerated cross section along line A-A' showing truncation of units along crest of Algonquin Arch and the reappearance of equivalent strata to the northwest.

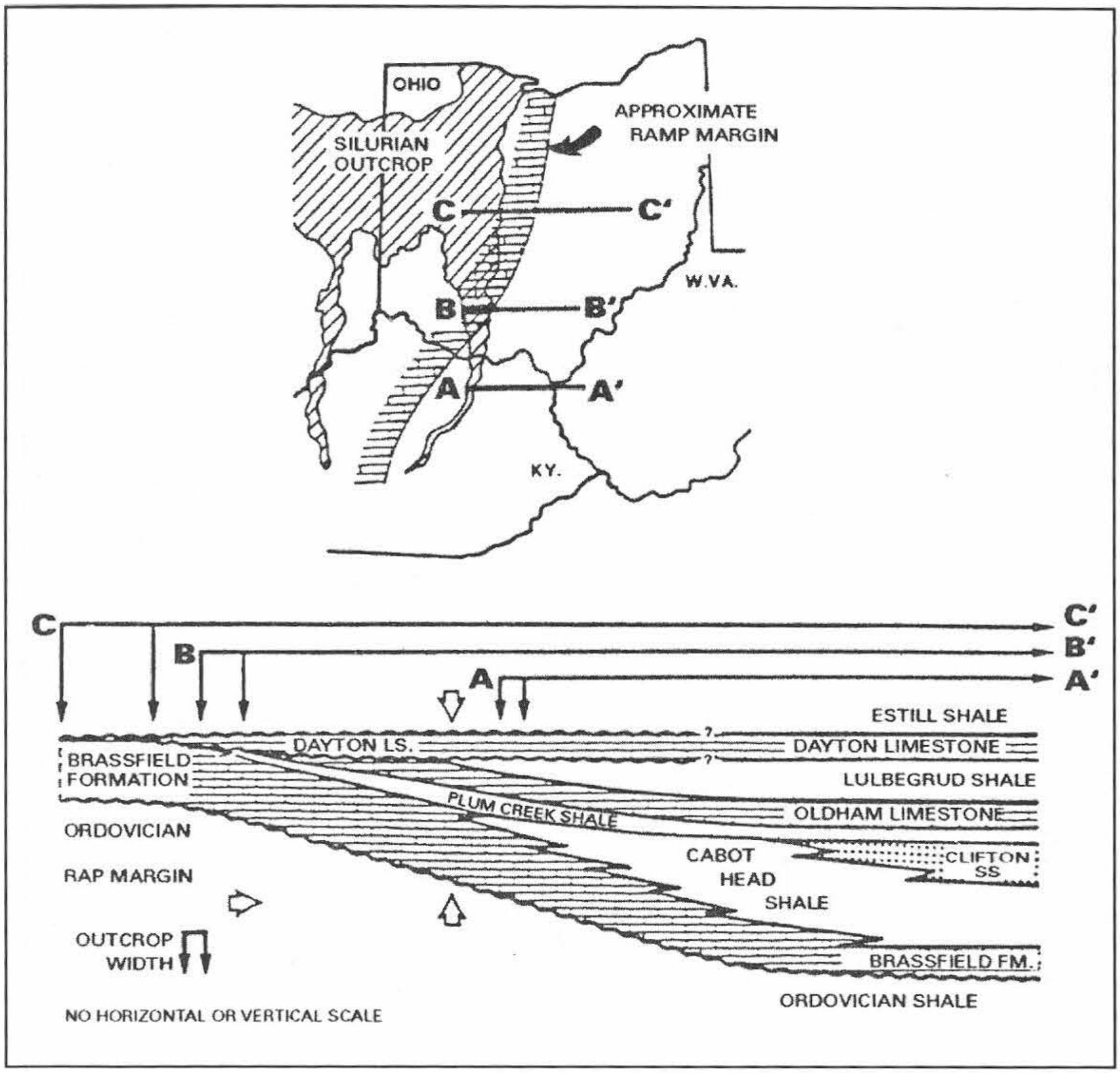


Figure 9. Regional cross sections of Silurian strata through south-central Ohio and northern Kentucky. Note the regional truncation of units along a proto-Findlay Arch (northwest or left side of cross section) below a major unconformity beneath the Dayton Limestone. Adapted from Lukasik (1988).

In southern Ohio a unit lithologically similar to the Merritton, termed Dayton Limestone also overlaps the regionally angular medial Silurian unconformity (Lukasik, 1988; Fig. 9).The Dayton consists of pale gray, glauconitic and heavily burrow-mottled limestone/dolostone. It passes upward into gray green Estill Shale up to 60 m thick, which is correlated with the Williamson-Willowvale interval on the basis of *amorphognathoides* Zone conodonts, as well as graptolites. The upper contact of the Estill is sharp at the base of the Bisher Formation, interpreted as a forced regression surface. The lower 1.0 to 2.0 m of the is composed of nearly barren dolomitic calcarenites to calcisiltites in 5 to 10 cm amalgamated and semi-tabular beds (silty carbonate facies of Bowman, 1956), greatly resembling the age-equivalent Rockway Dolostone of New York.

Sequence V: Upper Clinton Group, Lower to Middle Wenlock (Sheinwoodian)

The Irondequoit Limestone (crinoidal grainstone; 3-5 m), Rochester Shale (gray, calcareous mudstone with interbedded calcisiltites and bryzoan-brachiopod-pelmatozoan packstone storm beds; 0.5-20 m), and DeCew Dolostone (argillaceous laminated and typically heavily deformed dolostone; 3-4 m) together, form another genetically related sequence (Sequence V) in the upper Clinton Group (Figs. 3, 10, 11). The lateral equivalents of these units in the Bruce Peninsula are assigned to the Amabel Formation of the Albemarle Group (Fig. 8; Bolton, 1957; Armstrong and Goodman, 1990).

The basal Irondequoit Disconformity (Sequence IV-V boundary) is nearly planar with little evidence for regional truncation (Figs. 10, 11). However, this contact becomes increasingly sharp westward from Rochester, New York (probable basin center) and the basal Irondequoit contains clasts of the underlying Rockway dolostones. The Irondequoit, changes from thin bedded skeletal wackestones and packstone in the basin center to massive, amalgamated grainstone to the northwest. The unit represents open shelf to crinoids shoal (BA-3) environments. The upper contact of the Irondequoit is an abrupt, but conformable flooding surface. Locally, in Niagara County, small thrombolitic-bryzoan bioherms occur at the Irondequoit-Rochester transition zone; the mounding apparently associated with rapid deepening (Sarle, 1901; Cuffey and Hewitt, 1989).

In southern Ohio the interval corresponding to the Irondequoit Formation is represented by a unit of the Bisher Dolostone, informally designated "*Cryptothyrella*" beds, consists of 0.5 to 1 m of medium, thick and massive bedded fossiliferous dolostone. It tends to weather buff to pale orange in color and is rather porous owing to the weathering of crinoidal ossicles and other fossil debris. Internally, beds may display slightly irregular, darkly stained surfaces, apparently hardgrounds, especially near the top. The topmost 20-50 cm ledge is typically most heavily fossiliferous and comprised mainly of crinoidal pack- to grainstone. The sharp basal contact of this unit is interpreted as the sequence boundary of Sequence V.

Near Dayton, Ohio the Bisher is referred to as the "Laurel Dolostone". This is apparently based on erroneous correlations with sections in southeastern Indiana (Kovach, 1974). Detailed measurements indicate instead that this unit is correlative with the middle Osgood Limestone of Foerste (1897), whereas the true Laurel of Indiana correlates with a thicker succession of crinoidal dolostones identified as Euphemia-Springfield in Green County, Ohio.

The basal Bisher has been sampled and identified as belonging to the *K. ranuliformis* Zone; underlying upper Estill Shale has been identified as upper *P. amorphognathoides* zone (Kleffner, 1989. 1990; pers. comm. 2000); as such, the interval appears to be time correlative with the Irondequoit Limestone in New York and Ontario, and with the Keefer Sandstone of the central Appalachians (Figs. 10, 12). All of these units also share a similar two-fold subdivision with middle shaly intervals, and an abundance of the brachiopod *Whitfieldella*. All are interpreted as parts of the transgressive systems tract of Sequence V.

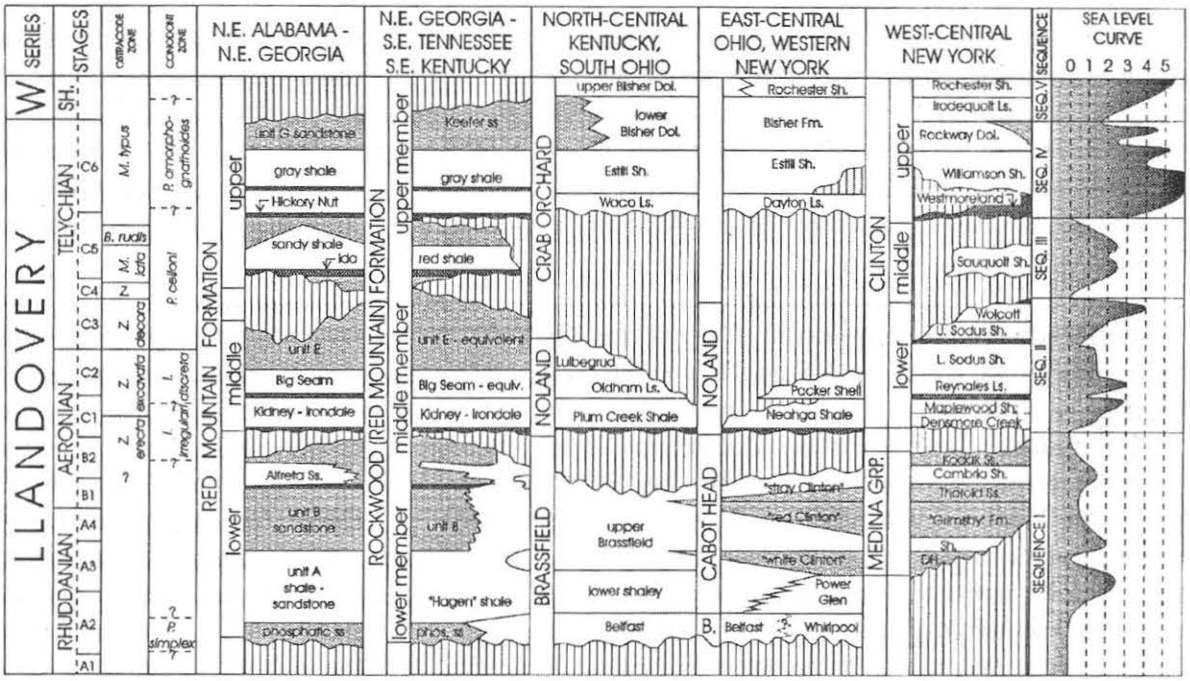


Figure 10. Correlation of Lower Silurian sequences in eastern USA; note particularly the comparisons of Kentucky, Ohio, and New York State. Curve on right side of diagram shows relative sea level curve for central New York State calibrated to benthic assemblages (BA-: shoreline, BA-2 above wave base; BA-3: average storm wave base; BA-4 deep storm wavebase; see Brett et al. (1993) for discussion of depths of these assemblages. Adapted from Brett et al. (1998).

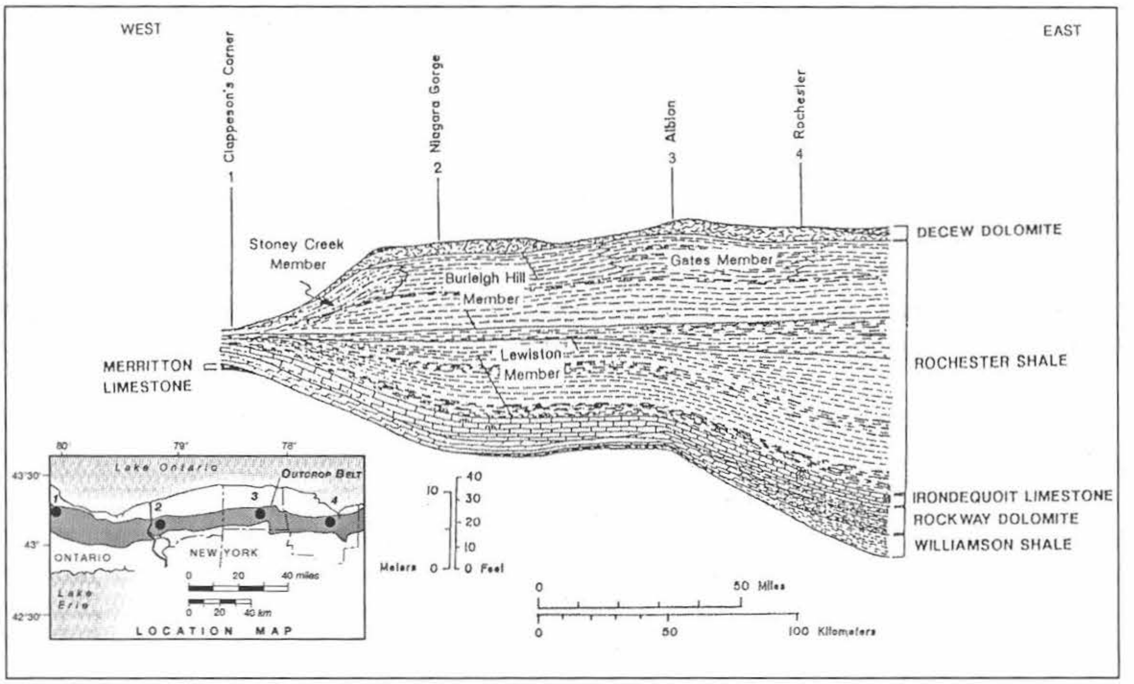


Figure 11. Regional stratigraphic cross section of Sequence V (upper Clinton Group) between Clappison Corners, Ontario and Rochester, New York. Datum is contact of Lewiston and Burleigh Hill Members of the Rochester Shale. From Brett et al. (1995).

To the west, in southeastern Indiana, the "*Cryptothyrella* bed" is represented by a 1 to 2 m thick compact dolomitic crinoidal pack to grainstone, generally mapped as the middle member of the Osgood Formation (*sensu* Foerste, 1897). It shows a sharp but cryptic basal contact. Its upper surface is also a sharp contact with the upper Osgood (=Massie) Shale. Locally, as at Napolean, Indiana, small fistuliporoid bioherms very comparable to those of the upper Irondequoit in western New York protrude upward from this flooding surface into the overlying upper Osgood Shale. This situation is precisely analogous to the small bioherms of the upper Irondequoit and underscores the lateral persistence of mounding events associated with this transgression (Fig. 11).

The Rochester Shale of western New York is divided into two members (Brett, 1982,1983a, b; Fig. 11). The lower-Lewiston Member- is highly fossiliferous along most of the Niagara Escarpment, with over 200 species of bryozoans, brachiopods, mollusks, crinoids, blastozoans, trilobites, and graptolites; bryzoan-brachiopod rich limestone beds occur near its base and top. This facies represents deeper, storm-influenced shelf environments (BA 3-4). However, to the south of the main outcrop, as in southern Niagara Gorge, the Lewiston becomes dark gray shale with sparse brachiopod-trilobite (BA-4 to 5) assemblages, indicating a southward dipping ramp (Brett, 1982, 1983 a, b). The upper Rochester shows two distinct facies: the Burleigh Hill Member-dark gray, sparsely fossiliferous shale-east of Grimsby, and the Stony Creek Member - banded dolomitic mudstone and argillaceous dolostone-to the west (Fig. 11; Brett, 1983). The Rochester Shale becomes increasingly carbonate-rich and thins dramatically to a feather edge from Niagara to Hamilton (Fig. 11). Thinning represents both condensation and erosion below the bases of the Stony Creek Member, and overlying DeCew and Gasport formations. Rochester-equivalent strata may reappear to the northwest in the lower Amabel Dolostone of the Bruce Peninsula (Bolton, 1957). Together, these observations indicate that during the late Llandovery to late Wenlock interval, the topographic center of the foreland basin lay to the southeast and that Hamilton was situated close to the crest of the Algonquin Arch (Fig. 11).

The Rochester Shale-equivalent interval in the Cincinnati Arch outcrop belt is best developed near Hillsboro, Ohio, where it has previously been identified simply as the dolomitic shale facies of the Bisher Formation (Bowman, 1956; Fig. 10). In this vicinity the interval is nearly 3 m thick and comprised primarily of medium dark gray, dolomitic shale with minor fossiliferous, dolomitic packstones in the lower half meter. These beds yield varied brachiopods, especially *Atrypa*, *Coolinia*, and small *Whitfieldella*, as well as ramose and fenestrate bryozoans, the trilobites *Dalmanites* and *Trimerus*, and pelmatozoan ossicles. The upper portion of this unit is dominantly barren dolomitic shale and thin laminated calcisiltites.

A similar, but slightly thinner (190-200 cm) dark gray shale occurs in the corresponding position in the Dayton area where it has been termed the Massie Shale (Foerste, 1935; Fig. 9). Again, the lower half-meter contains scattered fossils and shows a relatively abrupt contact with overlying barren dark gray shale.

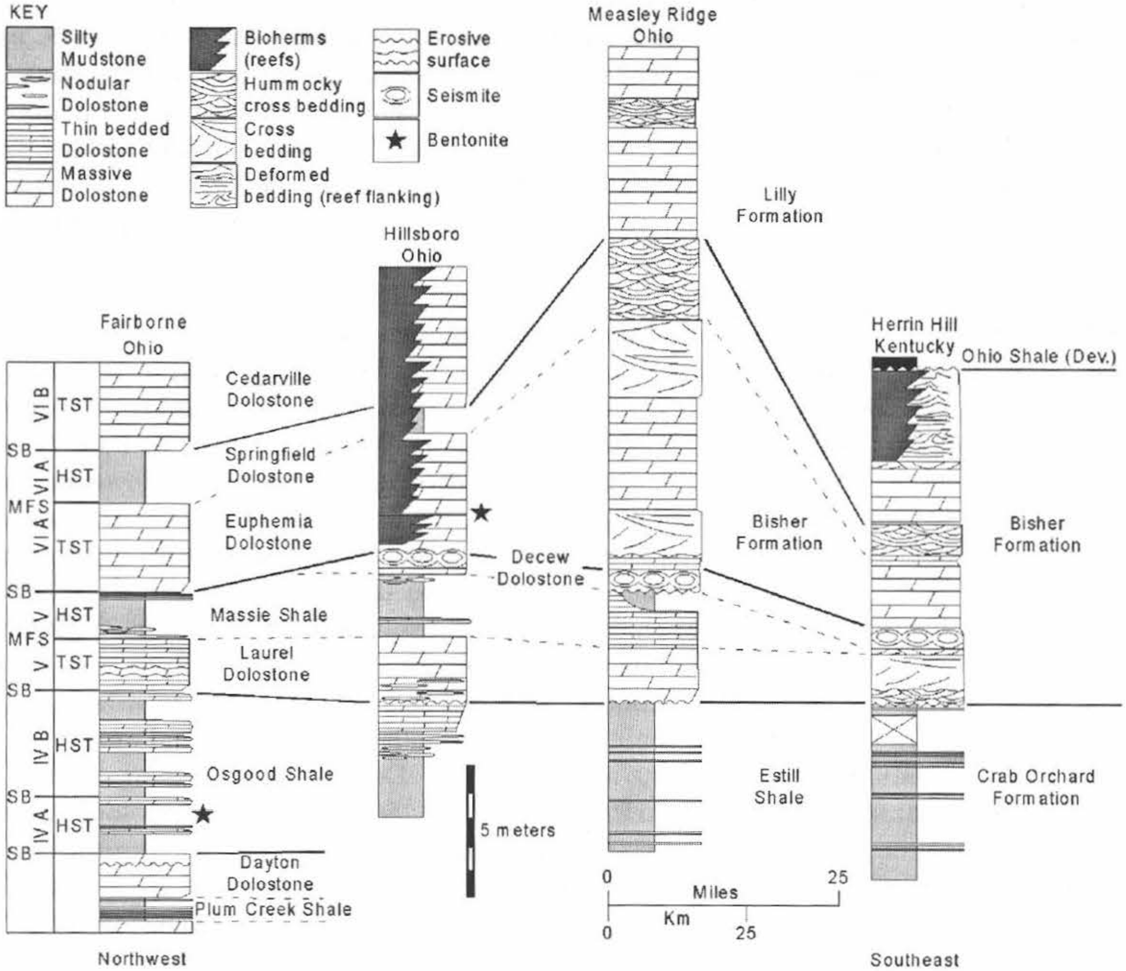


Figure 12. Correlated stratigraphic columns along NW to SE cross section from Fairborn, SE of Dayton, Ohio to Herron Hill, Lewis County, Kentucky. Note comparison to western New York-Ontario terminology. Sequence stratigraphic abbreviations as in figure 7.

To the southeast of Hillsboro, Ohio the interval becomes substantially thinner and more dolomitic but retains its general fine-grained aspect. A crinoidal packstone, about 50 cm above the base of this interval may represent the *Cryptothyrella* bed. The upper two thirds of the unit consists of thin-bedded, argillaceous dolostone and thin dark gray shales. This interval is evidently truncated at the top along a wavy contact with the overlying tan weathering hummocky laminated dolostones .

Correlation suggests that the dolomitic shale facies of the lower Bisher Formation is laterally equivalent to the Massie Shale. The Massie, in the Hillsboro to Dayton area closely resembles the Rochester Shale of the northern Appalachian Basin in both lithology and stratigraphic position. Moreover, near Dayton it has yielded a few fossils that elsewhere are found only in the Rochester. Nine of the 54 species found in the Massie have only otherwise been reported from the Rochester Shale (Busch, 1939). From these observations and the general conodont biostratigraphy we conclude that the Massie Shale is probably coeval with and coextensive with Rochester Shale. Indeed, the division of the Massie into a lower fossiliferous unit and upper barren dolomitic shale is consistent with division of the Rochester into lower fossiliferous Lewiston Member and upper sparsely fossiliferous Burleigh Hill Member (Brett et al., 1995; Figs. 10, 14). Finally, by outcrop to outcrop tracing, the Massie appears to correlate with the upper shale member of the Osgood Formation in Indiana. This interval has yielded a diverse fauna of brachiopods, bryozoans, and pelmatozoans that is very similar to that of the Rochester Shale in New York and Ontario (see Frest et al, 1999, p. 729, for a detailed faunal list).

All evidence indicates that the Massie Shale represents a deepening during early middle Wenlock time (Fig. 14). As such it is interpreted as the highstand systems tract of Sequence V. This same deepening, represented by the Rochester Shale, is also associated with a widespread interval of mud to sand deposition in the Appalachian foreland and may record a tectonic pulse in the Salinic Orogeny (Ettensohn and Brett, 1998). However, a similar interval of relatively deep-water mud deposition also occurs at least in Avalonia (e.g. the early mid Wenlock Coalbrookdale Formation in Britain, suggesting at least some eustatic influence (Brett et al 1990, 1998).

The DeCew comprises hummocky laminated, dolomitic calcisiltite, probably derived from storm-winnowing of carbonate shoal areas north of the present outcrop limit. A zone of extreme soft sediment deformation in the DeCew has been traced laterally along the Niagara Escarpment from east of Rochester, New York, westward to Hamilton, Ontario (Fig. 11) and into southwestern Ohio. The DeCew deformed zone represents a widespread seismite associated with a severe seismic shock and consequent slumping on the south-dipping ramp.

The middle of the Bisher Formation in southwestern Ohio, is a buff weathering planar to hummocky bedded silty dolostone ranging in thickness from 0 to 3 m. It is well developed in Adams and Highland county (Figs. 12, 14). This unit locally displays convoluted bedded, including overturned isoclinally folded laminated and intraclastic beds at several exposures in Adams County (Kallio, 1976). Thin lenses of similarly deformed fine-grained dolostone have also been recognized within the middle of the Bisher Formation at Herron Hill, Kentucky. This middle Bisher is stratigraphically equivalent to a thin bedded interval at the base of the Euphemia Dolostone in the Dayton area, originally described by Foerste (1917).

Both the lower and upper contacts of middle unit in the Bisher are sharp and probably disconformable. Locally, this dolostone exhibits an undulatory lower contact with as much as 20 cm of relief and truncates underlying shaly beds. The upper contact with crinoidal dolostones is planar.

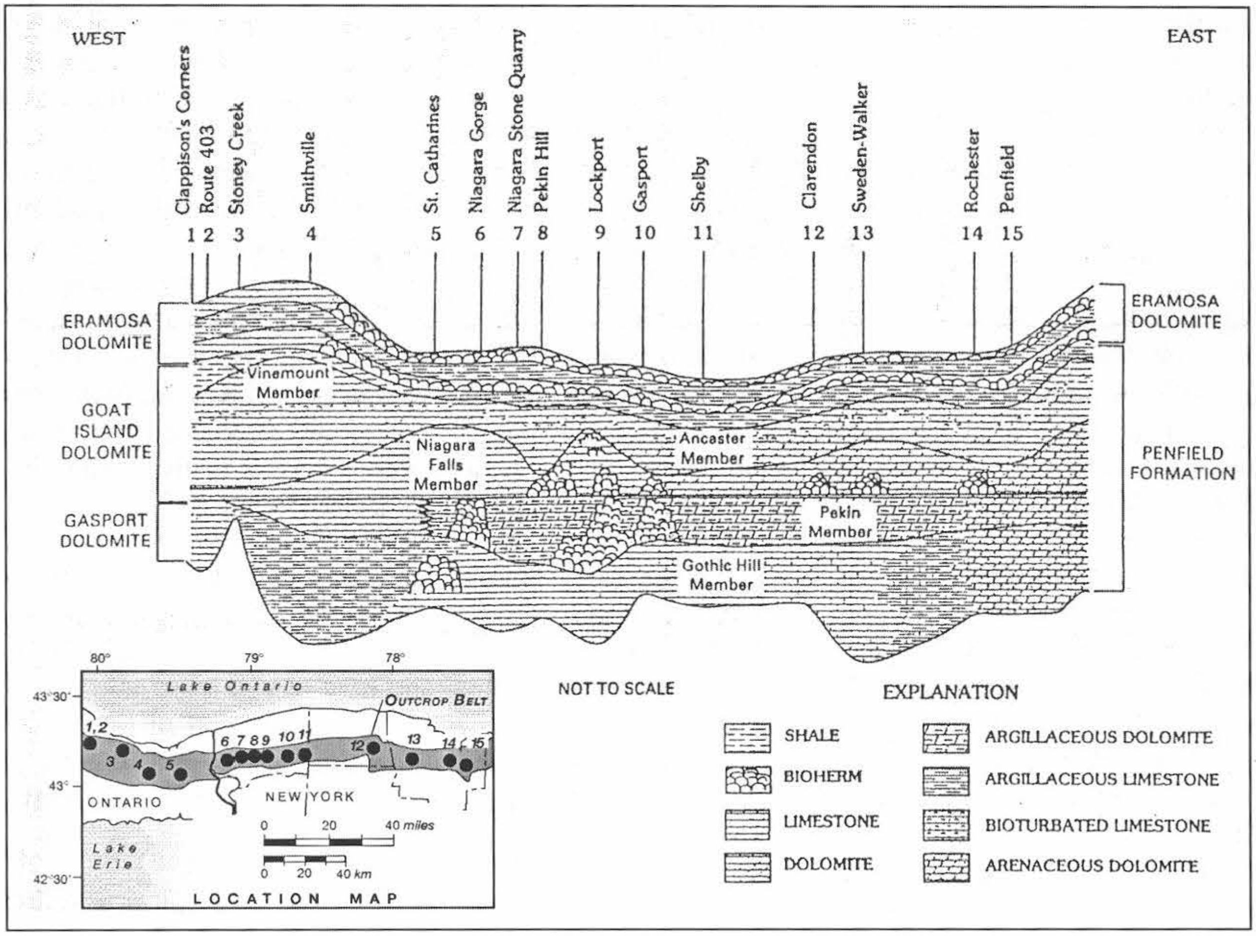


Figure 13. Regional stratigraphic cross section of Sequence VI, lower Lockport Group, between Clappison Corners, Ontario and Rochester, New York. Datum is contact of Gasport and Goat Island formations. From Brett et al. (1995).

The contorted fine-grained dolostones of the Bisher very closely resemble an interval of convolute bedded dolostone in the DeCew Formation in New York State and southern Ontario (Figs. 12, 14). This dolostone occupies an analogous and probably coeval position between the Rochester Shale and crinoidal dolostone of the overlying Gasport Formation. Consequently, we correlate unit with the DeCew Formation. The latter is considered to represent a falling stage systems tract of Sequence V. It shows a conformable to slightly erosional, channeled base and is locally truncated by the overlying sequence bounding unconformity at the base of the Lockport Group (Sequence VI).

Sequence VIA: Lower Lockport Group, Gasport Formation; middle Wenlock

The lower part of the Lockport Group (Sequence VI) comprises crinoidal pack- and grainstones, bioherms, and dolomitic wackestones near the base of the sequence (Gasport Formation), and vuggy grainstones, argillaceous, cherty wackestone and minor shales (Goat Island Formation). A clear-cut sequence boundary exists at the erosive base of the Gasport Formation (Figs. 3, 13). Tabulate-stromatoporoid bioherms typically extend upward from lower crinoidal grainstones into the argillaceous Pekin Member; this indicates that the upward growth of these reefs may have been stimulated by rising sea-level (Crowley, 1973; Brett, 1985; Brett et al., 1990). However, cap beds of fragmentary stromatoporoids suggest that the bioherms were extinguished and truncated by sea-level drop (Crowley, 1973)

Westward thinning, coarsening, and loss of argillaceous, biohermal facies in the Gasport Dolostone also suggests shallowing in that unit toward the Algonquin Arch (Fig. 13). The merging of Gasport and Irondequoit grainstones and associated total truncation of the Rochester Shale-near Hamilton also appears to mark a relative topographic high.

North of Hamilton, Ontario, where the Rochester Shale is absent, the grainstones of the Irondequoit and Gasport merge at a cryptic unconformity. In this area the succession of dolostones, including the equivalents Rockway (locally termed Lions Head Member), Irondequoit, Gasport and Goat Island are combined into the larger unit termed Amabel Formation (Armstrong and Goodman, 1990; Fig. 13). This interval presumably contains two important sequence boundaries (Sequence IV-V and V-VI). However, these key surfaces have been obscured by extensive secondary dolomitization.

The Amabel, especially the Wiarton Member, equivalent to Gasport and Goat Island formations along the northeast side of the Michigan Basin (Sanford 1969), is flat-lying and composed of locally porous, dolomitized bioclastic grainstone. Bioclasts are mostly crinoid pluricolumnals (segments up to 1 cm in width and 10 cm in length), with rhynchonellid brachiopods and stick-like bryozoans becoming the most common elements of a more diverse biota that characterizes the upper few meters. The unit ranges from less than 10 m to nearly 30 m in thickness over an along-strike distance of some 30 km (Hewitt, 1971); this is thought to reflect a primary mounded topography of the megashoal complex (Pratt and Miall, 1993). The Amabel changes in character northward, and it is lithologically more heterogeneous on the Bruce Peninsula (Bolton, 1957). These coarse crinoidal grainstones thin and grade laterally westward and southward, in the seaward direction, to thin- and wavy-bedded, argillaceous, locally cherty, dolomitized lime mudstone and wackestone and subordinate grainstone of the Lockport Group; landward equivalents have been removed by erosion, but were probably thin-bedded, argillaceous dolostones of peritidal aspect.

Throughout southern Ohio, the fine grained dolostones of Unit D of the Bisher Formation are abruptly overlain by a massive interval of cross bedded, sandy textured crinoidal dolostone. In the northwest, at Yellow Springs and Dayton, this 3-6 m interval has been identified as Euphemia Dolostone (Figs. 12, 14). It contains a distinctive interval rich in pentamerid and *Whitfieldella* brachiopods near the base. This same basic stratigraphy,

including the lower brachiopod shell beds, persists at least to Hillsboro, Ohio where the interval is simply referred to as upper Bisher Dolostone. Farther south, near Peebles, Ohio the equivalent interval thickens substantially to over 10 m and consists mainly of trough and tabular (including herringbone) cross-bedded sandy dolostone or dolomitic sandstone with thin intervals rich in crinoidal grainstone. A similar but thinner lithological unit is present at Herron Hill, Kentucky. The correlation of this interval into Indiana is somewhat uncertain, but we suggest, based on stratigraphic position that it is represented in the lower crinoidal dolostones of the Laurel Formation.

The upper Bisher resembles the Gothic Hill Member of the Gasport Formation in Ontario and western New York (Brett et al., 1995) and is probably coeval with that unit (Fig. 14). The sharp base of this unit is interpreted as the Sequence V-VI boundary, corresponding to the sharp, erosive base of the Lockport Group (base of Gothic Hill Member) in New York and Ontario (Fig. 14). Consequently, this interval is interpreted as the LST and TST of fourth order sequence VIA. Substantial lateral variations in thickness are typical of the Gothic Hill Member in western New York and Ontario as seems to be evident in the corresponding interval in southern Ohio-Kentucky. The interval may have developed as a series of sand-pelmatozoan megashoals during regional transgression (see Pratt and Miall, 1993).

A smectitic clay seam, identifiable as a K-bentonite (W. Huff, pers. comm. 1999), occurs about 1.5 m above the base of this unit a Hillsboro and a similar thin seam has been found in a small outcrop along the AA Highway near Vanceburg, Kentucky. Reconnaissance study of the Gasport Formation near Lockport, New York has also resulted in discovery of a clay bed, possibly also a K-bentonite in the Gothic Hill Member. Although these are preliminary observations, this horizon may provide a useful datum for time correlation. Its occurrence seems to corroborate sequence stratigraphic inferences.

At Hillsboro, Ohio a large (>100 m across and up to 10 m thick) mound of massive dolomicrite extends upward from within (Fig. 12). Bedding of the succession beneath this mass has been deformed into a local syncline apparently due to differential compaction beneath the mound. Although this mass was previously attributed to contorted bedding from collapse, close observation revealed the presence of encrusting bryozoans, pelmatozoan holdfasts, and tabulate corals. The mass is interpreted as a bioherm or mud mound. Along its flanks it interfingers with beds of coarse pack- and grainstone composed of large crinoid pluricolumnals. These beds show hints of graded bedding (fining upward) and they are interpreted as flank debris beds. The bioherm appears to be slightly inflected (notched) at the level of a prominent shale bed that overlies the crinoidal dolostone unit (see below). It is not clear whether the bioherm extends upward into overlying massive dolostones, or whether these beds merely drape the mound because this contact is poorly exposed near the top of the exposure.
The Hillsboro bioherm occupies a stratigraphic position that is consistent with the Gasport reef zone of western New York-southern Ontario (Crowley, 1973) and the "Lockport reef zone" in the McKenzie of Pennsylvania and West Virginia (Smosna and Patchen, 1978). Evidently, bioherms developed very widely in this interval (i.e. at the Gothic Hill-Pekin member boundary in terms of New York stratigraphy; Fig. 13). In a sequence context, these bioherms seem to have grown upward on a substrate of stabilized echinoderm sand/gravel associated with a condensed interval and maximum flooding zone during middle Wenlock time.

The massive, cross bedded crinoidal dolostones of the Bisher (or Lilly) are everywhere overlain by thin bedded, argillaceous, planar to hummocky cross laminated dolostones and shales. Near Fairborn and Yellow Springs, Ohio this interval of thinly and regularly bedded fine grained dolostones/dolomitic shales is referred to as the Springfield Dolostone (Fig.14). At the Hillsboro roadcut this zone is particularly shaly, with up to 40 cm of greenish gray dolomitic shale near the previously discussed bioherm. As noted this shale appears to interfinger with the bioherm. Near Peebles the upper is recorded as a 4-6 m finer grained hummocky laminated dolostone above the massive unit.

Overall, the thin bedded, argillaceous dolostone of the Springfield-upper Bisher closely resembles that of the Pekin Member in New York and its relationship to the bioherm at Hillsboro is similar to that seen in the Gasport bioherm zone of New York. This interval, like the coeval Pekin Member shows a vague shallowing upward trend and is interpreted as the HST of fourth order subsequence VIA.

Sequence VIB. Goat Island Formation

A second fourth order sequence boundary at the sharp, erosive base of the Goat Island Formation. Locally, as at the Rte. 427 roadcut at Pekin, New York, there is substantial erosional relief at this sequence boundary and mounds of biohermal stromatoporoid-rich dolomicrite appear to infill this erosional topography. In the Goat Island Formation, massive crinoidal grainstone shoal facies (BA-2-3) characteristic of the Niagara Gorge are replaced westward by thin-bedded, cherty wackestone (Ancaster Member; BA-3-4), which thicken to a maximum in the Hamilton area before passing laterally again into massive dolostones (Fig. 13). This westward deepening trend is also evident in the Vinemount Member, which is a slightly cherty dolowackestone to the east but is represented by dark, dolomitic shales near Hamilton, Ontario. These shales only persist northwest to near Dundas where they are replaced or pinched out against upper Amabel dolostones (Fig. 13).

The picture is not entirely straightforward, as small areas of shaly and/or cherty dolostone also occur locally in the Goat Island position in Niagara County, New York. This pattern suggests that minor fault block-controlled basins may have formed during Goat Island deposition. (Sanford et al., 1985) This irregular topography may have been associated with an abrupt westward migration of the main basin center to the Vinemount-Hamilton region (Fig. 13).

In southwestern Ohio the lower Goat Island interval appears to be represented by massive, vuggy, crinoidal dolostone (originally pack- or grainstone) identified near Yellow Springs and Dayton, as the Cedarville Dolostone and to the southeast as the Lilly Formation. This interval closely resembles the lower massive Niagara Falls Member, of the Goat Island Formation in western New York-Ontario with which it is tentatively correlated (Figs. 9-13). Both units have been dated as belonging to the *O. sagitta rhenana* Zone (Kleffner, 1990; pers. comm. 2000). This unit is interpreted as a LST and TST; its sharp, undulatory basal contact is the subsequence S-VIB boundary. Farther southeast, near Vanceburg, a small roadcut on the AA Highway, an interval of bioturbated stromatoporoid-rich dolostone sharply overlies shaly dolostones. This interval shows no trace of mounding but instead carries abundant coenostea of stromatoporoids and favositid and cladoporid corals in a bioturbated dolomitic calcisiltite (Mason et al. 1992). Upper portions of the outcrop show abundant cream colored chert nodules. This outcrop most closely resembles Ancaster facies of the lower Goat Island Formation in New York and Ontario. Stromatoporoid-coral biostromes and local mud mounds (e.g. the Pekin "Bioherm" see Brett et al., 1994) are typical of the Niagara Falls Member immediately overlying the Pekin shaly dolostone, whereas light chert nodules locally characterize the middle Ancaster Member (see Brett et al., 1995, for discussion). Together, this evidence suggests correlation with the lower Goat Island Formation and, as with that succession, the Lilly Formation is interpreted as the combined LST-TST of fourth order subsequence S-VIB (Figs. 12, 14).

Throughout south-central Ohio a 0 to 3 meter thick interval of medium gray shale and argillaceous dolostone abruptly overlies massive crinoidal dolostone of the Lilly. Several small (< 1 m high) bioherms protrude above this surface and very minor mounds with small stromatoporoids occur in association with a middle dolomitic band that splits the shale-rich beds into upper and lower portions. This contact is interpretable as a major flooding surface; it seems to correlate with the base of the Vinemount Member- a succession of shales and thin bedded argillaceous dolostones, particularly well developed near Hamilton, Ontario (see Brett et al., 1995, for formal definition). Again, the positioning of the biohermal beds is consistent with the model discussed above. Mounding is correlated with deepening and perhaps cleaning of the sea water at a maximum starvation surface. The post-Lilly shale beds and probably coeval Vinemount beds of the Goat Island Formation represent the HST of subsequence VIB.

Sequence VII: Upper Lockport-Vernon Formation, upper Wenlock to Ludlow (Ludfordian)

Biostromal to flaggy argillaceous, dolostones (Eramosa) and massive, buff, biostromal to biohermal dolostone (Guelph Formation) form the upper part of the Lockport (or Albemarle) Group (Fig. 3). The Eramosa, interpreted by Armstrong and Johnson (1990) as an inter-reefal, dysoxic environment (BA-2-3), has recently yielded assemblages of soft-bodied fossils, including algae, and unusual arthropods (Waddington and Rudkin, 1992; LoDuca, 1995, 1996; Tetreault, 1995, 1996). A disconformity at the base of the Eramosa Formation in New York is now interpreted as the boundary of a sequence (VII) not previously recognized by Brett et al. (1990). Still further westward migration of the basin axis (and final subsidence of the "Algonquin Arch") appears to have occurred during deposition of the Eramosa and Guelph Formations, in which deepest facies (BA-3) occur northwest of Hamilton, Ontario, while biostromal to stromatolitic facies (BA-2) occur in the Niagara region (Brett et al., 1995).

In southern Ohio the post-Lilly shale zone is abruptly overlain by massive vuggy, stromatoporoid, coral, and brachiopod-bearing Peebles Dolostone. This contact is thought to correlate with the sharp (sequence bounding) basal contact of the Eramosa Dolostone in Ontario and New York (Sequence VII). However, the details of this transition are presently under study and are beyond the scope of this paper.

In drill cores the Guelph can be seen to pass gradationally upward through series of interbedded shaly dolostones and dolomitic shales of the Vernon Formation (upper Ludlow; Salina Group). The Vernon Formation represents a tongue of siliciclastic sediments from the Bloomsburg-Vernon clastic wedge that was shed from tectonic regions (Salinic Orogeny) in the mid-Atlantic region. In central New York, the Vernon consists mainly of red mudstones, but in western New York and Ontario the unit consists of over 60 m of greenish gray shales and buff dolostone with interbedded anhydrite.

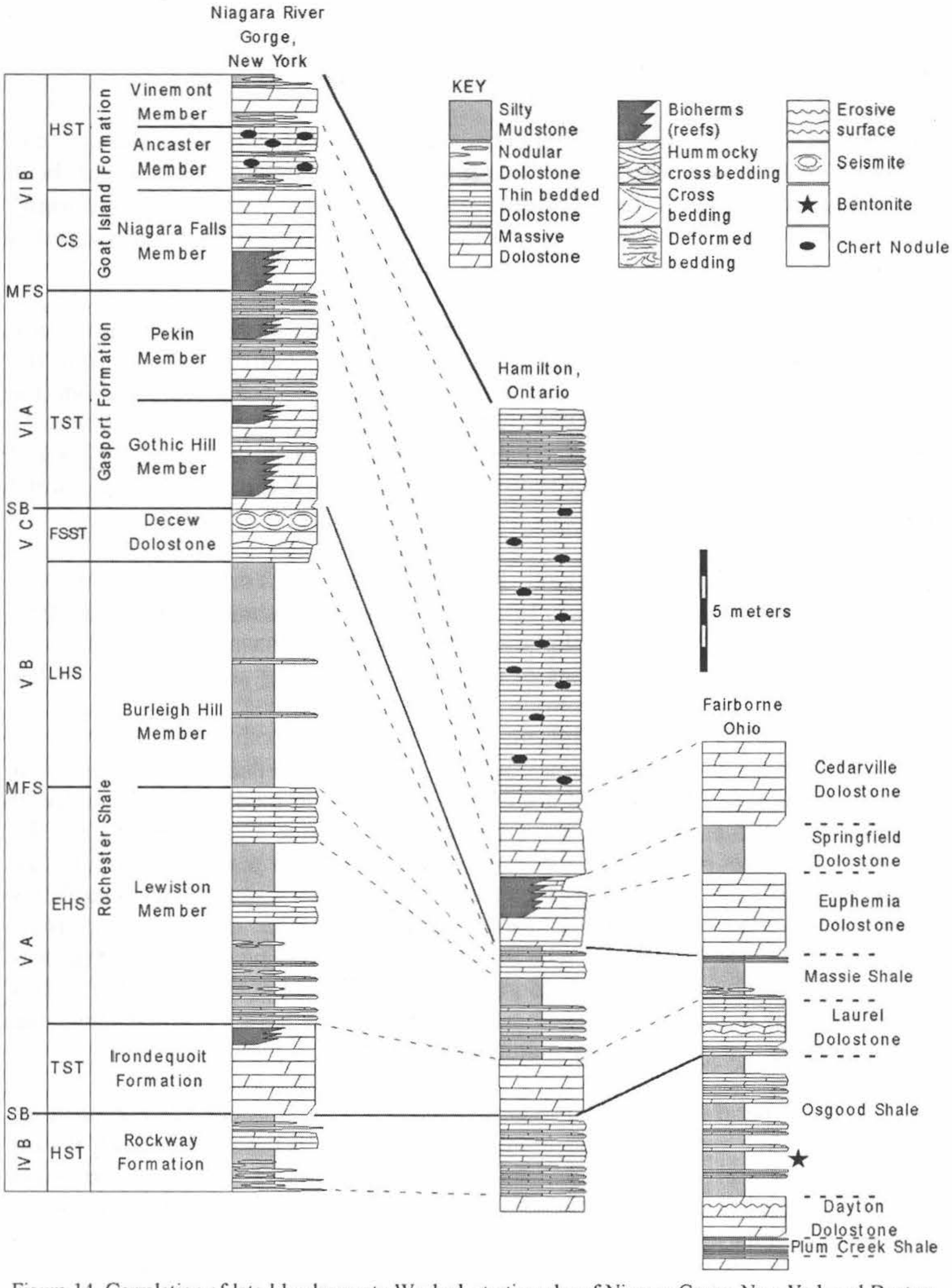


Figure 14. Correlation of late Llandovery to Wenlock stratigraphy of Niagara Gorge, New York and Dayton, Ohio, showing basic similarity of succession. Sequence stratigraphic abbreviations as in Figure 7.

Sequences VIII, IX: Upper Salina and Bertie Groups, Upper Ludlow-Pridoli

In southern Ontario the upper Salina Group comprises over 60 m of dolostones, shales, and evaporites but it is very poorly exposed. Detailed sequence stratigraphy has not been undertaken. Brett et al. (1990) noted that an erosion zone and regionally angular unconformity exist between the Vernon and overlying Syracuse Formation in central New York and suggested that a sequence boundary exists at this level within the Salina Group (Figs. 3, 15).

The Syracuse and Camillus formations, each about 30 m thick, comprise gray to green-maroon mudstones, buff dolostones and evaporites (Fig. 11). Key salt-gypsum horizons within the Syracuse have been traced in subsurface through Ontario from the Appalachian foreland into the Michigan Basin (Fig. 11; Rickard, 1969; 1975; Milne, 1992). These strata were evidently deposited under arid subtropical climates in interconnected but restricted basins. The widespread nature of the evaporite-dolostone-shale alternations indicates both that topography (e.g. on the Algonquin Arch) was subdued and that the cycles were due to eustatic-climatic effects.

The highest Silurian strata (middle-upper Pridoli) in the western New York-Ontario areas are presently assigned to the Bertie Group (Figs. 3, 15). They comprise a relatively thin (16-18 m) cyclic succession of distinctive, buff gray, slightly argillaceous dolostones ("waterlimes", so-named because of their geochemical properties of natural cement rocks) and dolomitic shales. The basal Oatka Formation is dominantly dolomitic shales and is gradational with the underlying Camillus Shale. The Fiddlers Green (5.5-8 m; 18-25') contains both massive brownish waterlimes and some thrombolitic dolomitic limestone that represents the deepest water facies of the Upper Silurian. Scajaquada Formation is a thin unit of dolomitic mudstone, apparently of sabkha origin, while the Williamsville carries a repeat of waterlime facies resembling the Fiddlers Green. Both units are noted for the occurrence of excellently preserved eurypterids, phyllocarids, and other fossils that are suspected to represent a brackish water estuarine biofacies that bordered hypersaline shallow seas. Finally, the Akron Dolostone (2.8 –2.5 m; 6-8') consists of massive burrow mottled, vuggy dolostone with molds of corals. This unit apparently records a return to somewhat more normal marine lagoonal environments. Locally, a higher (latest Silurian to earliest Devonian dolostone, the Clanbrassil Formation has been identified above the Akron; it records a return to "waterlime" deposition (Ciurca, 1990).

During deposition of the upper Salina and Bertie Groups there is a west to east displacement of depocenters (typically marked by thickest accumulations of halite in the Appalachian Basin) through central to east central New York State (Rickard, 1969). The Lower Devonian Helderberg Group was deposited in a basin the axis of which lay southeast of New York State, while western New York- Ontario were above sea level. The lateral consistency of upper Salina and Bertie Group units along the central New York-southern Ontario outcrop belt suggests also that the facies strike in this region is roughly east-west, parallel to the northern rim of the foreland basin.

Silurian-Devonian (Wallbridge) Unconformity

In western New York and southern Ontario Upper Silurian strata are unconformably overlain by upper Lower Devonian quartz arenites of the Oriskany Sandstone, and cherty, fossiliferous carbonates of the Bois Blanc and/or the Middle Devonian Onondaga Formation (Figs. 15). This second order "Wallbridge Unconformity" marks the boundary between the Tippecanoe and Kaskaskia supersequences (Sloss, 1963; Dennison and Head, 1975). It displays evidence of karst development, with irregular relief of up to 3 m. This unconformity apparently records a major late Early Devonian draw-down in sea-level which exposed older Silurian carbonates and evaporites to subaerial weathering and erosion. Sea-level rise in the late Early Devonian (late Pragian to Emsian) resulted in flooding of the irregular erosion surface. Kobluk et al. (1977) described rockground features of the *Trypanites* bored and glauconite-coated upper contact of the Silurian Akron Formation in southern Ontario (see Brett et al., 2000, for description of a similar surface in western New York). To the west, near Hagerstown, Ontario the basal fossiliferous quartz arenites of the Oriskany Formation (Pragian) rest unconformably on the Wallbridge Unconformity, but in the Ft. Erie-Port Colbourne area and in western New York the Oriskany has been removed by subsequent erosion and the basal Devonian unit is the Bois Blanc Limestone.

BASIN TECTONICS: MIGRATING DEPOCENTERS AND FOREBULGE

The ability to delineate and correlate thin sequence stratigraphic intervals also permits recognition of regional patterns that may be the result of minor tectonic adjustments and shifting depocenters within the Appalachian Foreland Basin (Goodman and Brett, 1994; Ettensohn and Brett, 1998). Within the Silurian as a whole, we recognize a large scale pattern of eastward-westward-eastward migration of the deepest water area and depocenter of the Appalachian Basin during the Early Silurian to Early Devonian time. This tectonically driven effect is superimposed on the more widespread (eustatically controlled) pattern of sea-level fluctuation manifest in the depositional sequences (Fig. 16).

In addition, minor abrupt facies changes within discontinuity-bound sequences, on the scale of a few kilometers, provide evidence for localized flexure of the crust, probably in the form of subsurface fault blocks, as described by Sanford et al. (1985). These local flexures may not be independent from the overall tectonic pattern but may record the local crustal response to migrating "waves" of compression due to episodes of tectonic loading and relaxation (Beaumont et al., 1988).

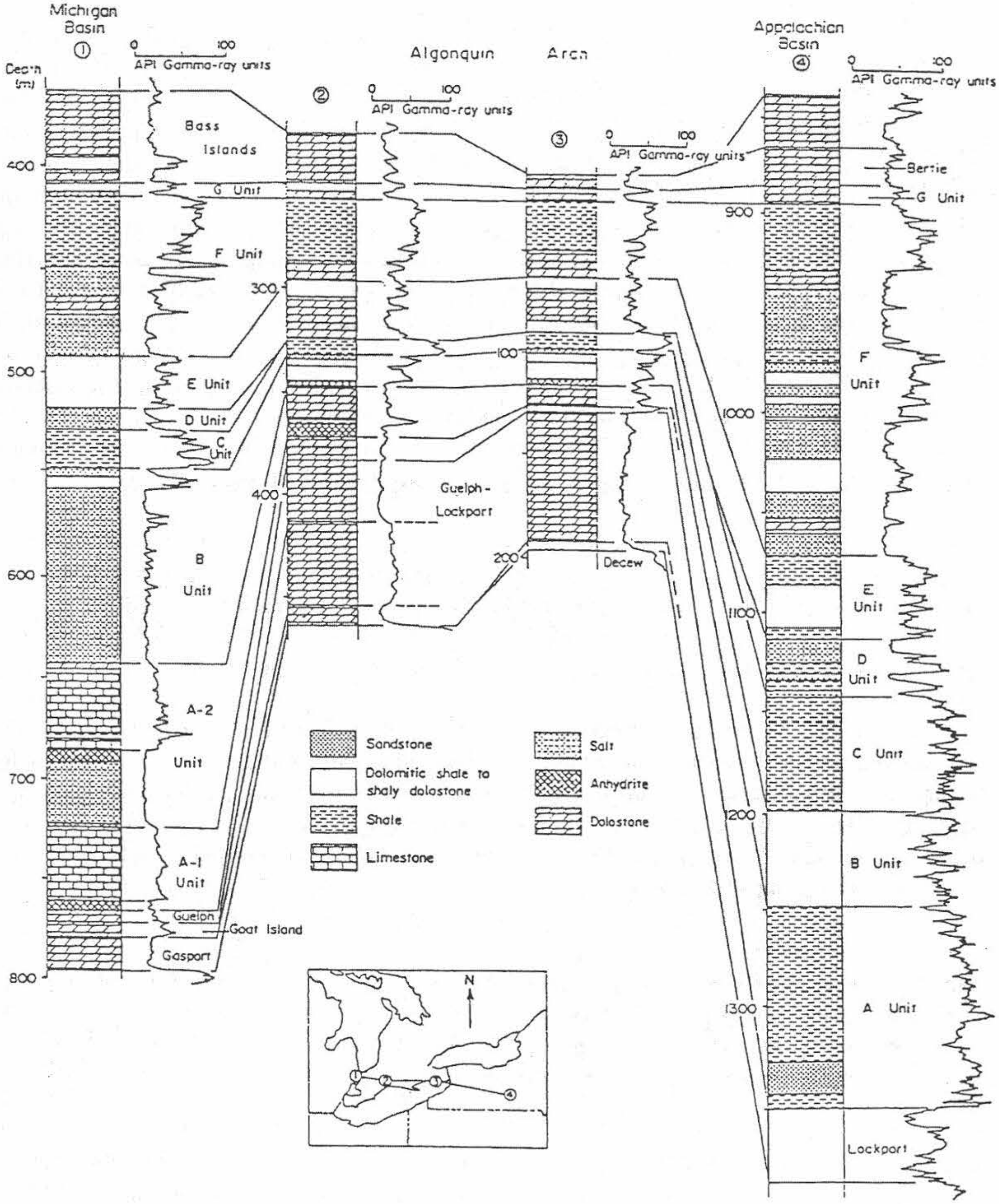


Figure 15. Correlated columns and gamma ray profiles of Upper Silurian Lockport, Salina, Bertie and Bass Islands groups in southern Ontario and south-central New York. Note persistence of salt marker units in the Salina Group and the irregular Wallbridge Unconformity beneath Devonian strata. From Thurston et al. (1992).

PALEOECOLOGY

Throughout the Early to mid Silurian marine waters in the foreland basin and mid-continent platform were of normal salinity and diverse invertebrate faunas formed a series of onshore-offshore biofacies gradients. These biofacies formed extensive belts parallel to paleoshoreline (Brett, 1999) that have been termed Benthic Assemblages (BAs; Fig. 2) by Boucot (1975). Benthic assemblages have been calibrated to absolute depths by Brett et al. (1993). Benthic Assemblage-1 (BA-1) is dominated by lingulid brachiopods, bivalves, gastropods, or, in carbonates, stromatolites/thrombolites and ostracodes deposited within just a few meters of sea level. BA-2 is dominated by low diversity brachiopod associations (especially *Eocoelia* in the Early Silurian; Fig. 3) and tabulate coral and stromatoporoid biostromes and bioherms deposited above fair-weather wave-base (<20 m). Tabulate-stromatoporoid patch reefs of BA-2 were particularly well developed during the Wenlock-early Ludlow of the Niagara region (Crowley, 1973; Armstrong and Johnson, 1990). BA-3 is characterized by diverse pelmatozoan associations and pentamerid brachiopods deposited near or just below fair-weather wave base (~20-30 m). BA-4 is typified by diverse assemblages of brachiopods, bryozoans, trilobites, mollusks, and pelmatozoan echinoderms deposited between fair-weather and storm wave-base (30-60 m). BA-5 is characterized by small brachiopods, a few bivalves, and, in some areas, graptolites deposited near storm wave-base (>60 m). These benthic assemblages have proven to be widely mappable and useful in determining the depths of the interior seas (Johnson, 1987; Brett, 1999). Events of storm-related deposition buried organisms rapidly on the sea floor leaving them largely intact, resulting in spectacular obrution deposits of crinoids, rhombiferan cystoids, asteroids, and trilobites, especially in the Cabot Head and Rochester shales (Brett and Eckert, 1982; Taylor and Brett, 1996).

During Late Silurian (Cayugan or Pridolian) epeiric seas in the Appalachian and Michigan basins became restricted by barriers of the Bloomsburg clastic wedge (to the south) and barrier reef complexes (around the Michigan Basin) and developed hypersaline conditions under which evaporites formed and few organisms, other than rare ostracodes, occupied open shelf environments. In estuarine areas, where fresh water streams mixed with the hypersaline environments, distinctive brackish water biofacies developed (Clarke and Ruedemann, 1912; Ciurca, 1990). These peritidal (BA-1 to 2) biofacies were dominated by eurypterids, and a few species of ostracodes, mollusks and algae. Only in the latest Silurian were normal marine salinities partially restored in the Appalachian Basin.

Silurian marine invertebrates and their biofacies exhibit long term concurrent evolutionary stability punctuated by abrupt intervals of extinction, immigration, evolution, and restructuring. Brett and Baird (1995) termed this pattern "coordinated stasis" and recognized distinctive stable faunas (or ecological evolutionary subunits) in the Silurian of the Appalachian Basin; these correspond very roughly to depositional sequences and were termed the: 1) Medina; 2) Lower Clinton; 3) Upper Clinton-Lockport; and 4) Salina faunas. The latest Silurian-Early Devonian E-E subunits are absent from the study area due to erosion at the Wallbridge Uncoformity.

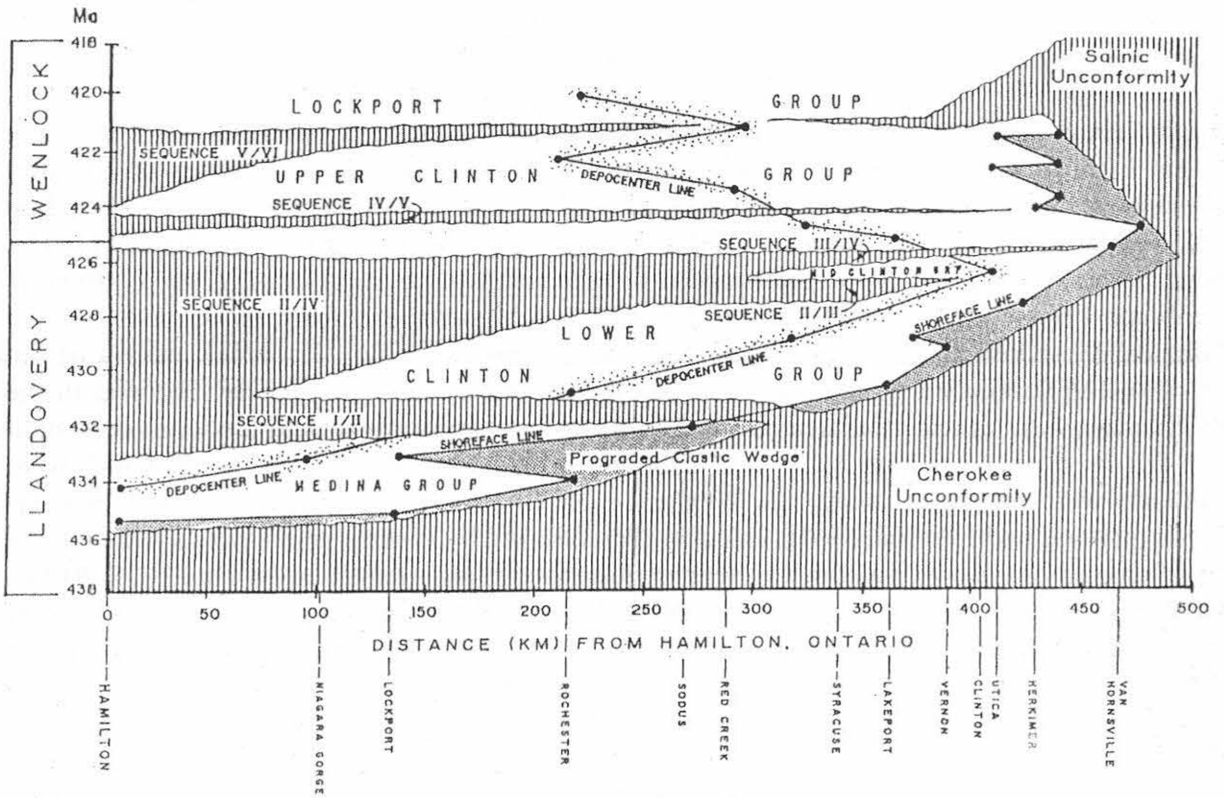


Figure 16. Migration of depocenter and shoreface facies during early to medial Silurian time relative to locations along the east-west outcrop belt (~470 km) from Hamilton, Ontario to Van Hornesville in central New York State. Note time-scale on vertical axis in millions of years. Vertical ruling indicates unconformity. From Goodman and Brett (1994).

EVENT STRATIGRAPHY

Two excellent examples of probable seismites occur in the Silurian of the Niagara area. The first consists of a ball and pillow horizon in the reddish sandstones of the upper Grimsby Formation (McLaughlin and Brett, 2006). The larger deformed masses, up to 2 m, across, show overturned folds and flame structures; The deformation is not observed everywhere but seems to be concentrated in areas of thicker sandstone beds that may represent shallow tidal channel fills (Duke and Brusse, 1987). Basal surfaces of the pillows display small-scale load casts, striations, deformed burrows and load crack casts indicating deformation of semi-plastic muds by loading. To date, this horizon has been traced from Niagara Gorge, near Lewiston, NY westward to Hamilton, Ontario.

The second example of a probable seismite horizon in the Niagara region is the well-known "enterolithic" interval in the DeCew Dolostone (Nairn, 1973; McLaughlin and Brett, 2006). The DeCew consists of buff weathering, medium dark gray, hummocky laminated dolostone (originally fine calcarenite or calcisiltite) with scattered layers of small (1-4 cm) intraclasts (Fig. 17). The lower 1 to 1.5 m of the DeCew displays extraordinary deformation that includes, ball and pillow style deformation and recumbent folds in the intraclasts beds. Overlying beds in the upper DeCew are unaffected. Preliminary study of orientations of deformed beds in the DeCew indicates that these folds do not show a consistent overturn direction (Fig. 18). This chaotic orientation suggests liquefaction and foundering, possibly accompanied by some very local submarine sliding. The bases of these beds show numerous invaginations, consistent with squeezing of liquefied sediments upward into the overlying beds. A very similar, but much less extensive, bed of deformed laminated, silty dolostone occurs in the upper Rochester Shale in the southern Niagara Gorge. The most impressive aspect of the DeCew deformed interval is its broad lateral extent. Deformation within this unit can be traced along nearly all Niagara Escarpment exposures from Penfield, east of Rochester, NY, to Hamilton, Ontario, where the DeCew has been erosionally truncated by erosion at the basal Lockport erosion surface (Sequence S-6 boundary). Similar deformation has been identified in a silty dolostone bed at the top of the Rochester Shale near Allenwood, Pennsylvania (Brett et al., 1990). Recently, a very similar zone of deformed dolostone has been identified in age-equivalent strata mapped as upper Bisher or lower Lilly formation in southern Ohio (Flanigan, 1986; Brett and Algeo, 2001). This bed also appears underlie the basal Lockport (Sequence VI) boundary at the base of the upper Bisher Formation in some outcrops. At Hillsboro, Ohio the deformed bed overlies a unit closely resembling, and probably equivalent to, the Rochester Shale (Brett and Ray, 2001, 2006). If the Pennsylvania and Ohio occurrences indeed correlate with the DeCew, this would indicate an enormous area of deformation of more than 200,000 km^2, making the DeCew one of the most widespread deformed zones yet reported. Deformed zones with similar areal distributions have been described from the Upper Ordovician of eastern North America (McLaughlin and Brett, 2004) and from the Triassic of Britain (Simms, 2003). Very large earthquakes (>7M) and bolide impacts are implicated as possibly triggering deformation (Simms, 2003).

At present there is no good indication of any one fault, which may have produced the seismic energy to cause the deformation of the DeCew sediments. However, the zone of deformation appears to lie adjacent to deep-seated basement faults, which bound the so-called Bass Island Structure in southwestern New York, northwestern Pennsylvania, and into eastern Ohio. Thrust induced loading of the cratonic margin likely resulted in reactivation of an entire network of basement faults well in to the cratonic interior, providing the triggering mechanisms to produce wide spread soft-sediment deformation.

The Silurian deformed beds display a range of sedimentary features that supply information about the environment and timing of deformation. Sedimentary structures demonstrate that these sediment layers were not deformed during initial deposition, but later, during shallow burial. In some instances there is evidence that these deformed beds were truncated while exposed at the seafloor indicating penecontemporaneous deformation. The deformed strata themselves are typically composed of thin shales, overlain by laminated silt- to fine sand-sized sediments (both carbonate and siliciclastic). The shales typically show little evidence of bioturbation; those burrows that are present are sharply defined indicating formation in firm muds. The overlying fine-grained carbonates/sands commonly contain hummocky to swaly cross bedding, suggesting lower shoreface deposition. These beds display sole marks (e.g., scratches, prods, and flutes) suggesting deposition on firm, over-compacted mud substrate that were stable, even during deposition of uneven sediment loads. The contacts between the mud and the overlying silt-sandstone are very sharp; an interface that is poised at instability given the thixotropic properties of the mud.

Observation of the deformation structures (e.g., ball-and-pillows, mudstone diapirs) suggests that mobilization of thixotropic mud caused deformation of the surrounding sediments during gel to sol transitions. In the gel state the muds were cohesive enough to record sole marks and to support the load of overlying silt or sand. However, during episodes of seismic shaking muds flowed upward as diapirs, evacuating from the lower part of a deforming interval (Fig. 17). Experimental data suggests that compact thixotropic muds will not deform unless triggered by a sudden shock, such as an earthquake wave (Brenchley and Newall, 1977). Detailed interpretation of similar occurrences in the Upper Ordovician of the Virginia-Kentucky region by Pope et al. (1997) and McLaughlin and Brett (2004) indicates that such widespread deformation over areas of a few tens of square kilometers would accompany large earthquakes with magnitudes in excess of 7 on the Richter scale. Pope et al. suggested that the ball and pillow horizons and related slumps might have been triggered by earthquakes in the Taconic Orogen or

movements of local basement faults. In any case, these dramatically deformed intervals provide excellent stratigraphic markers. They also indicate that the Appalachian foreland basin was not tectonically quiescent through the Silurian (see Ettensohn and Brett, 1998).

Lower Silurian deformed beds correspond in time with formation of shale basins and occurrence of minor K-bentonites. The timing and degree of tectonic loading of the Laurentian cratonic margin during the Silurian has been established by analyzing the distribution of dark shale basins (Goodman and Brett, 1994). The only deformed beds of the Llandovery (Grimsby and Thorold formations), for that matter since the Upper Ordovician early Maysvillian stage (~5 million years older), are coincident with these indicators of resumed tectonism. Similarly, Wenlock strata were deposited during the Salinic Orogeny (Ettensohn and Brett, 1998), a period noted for widespread formation of dark shale basins with different geometries than Taconic basins.

K-Bentonites

Altered volcanic ash beds provide some of the most valuable marker beds in the geologic record. In theory, these beds record single events of tephra deposition from explosive volcanic eruptions. Not only do they form excellent time lines but also many bentonites can be radiometrically dated. Few K-bentonites have been reported from the Silurian of the Niagara region. However, recently, thin clay layers that appear to be bentonites have been identified in the lower Lockport Group of western New York. These might correlate with probable K-bentonites that have been found in the medial Silurian on both eastern and western flanks of the Cincinnati Arch in Ohio, Kentucky, and Indiana (Brett and Algeo, 2001). Work on these beds is in very preliminary stages and will only be briefly discussed on this field trip.

Tempestites

Storms produce a distinctive suite of sedimentary deposits that range from coarse, amalgamated skeletal debris beds to small-scale hummocky laminated siltstones and sandstones, to distal mud layers. In some cases, particular conditions associated with a given storm bed may make it distinctive and usable for local or regional event correlation. Such applies to certain coarse skeletal debris layers, notably pavements or shingled shells of flattish to gently concavo-convex brachiopod shells and oriented them in edgewise clusters over a large area of seafloor; several

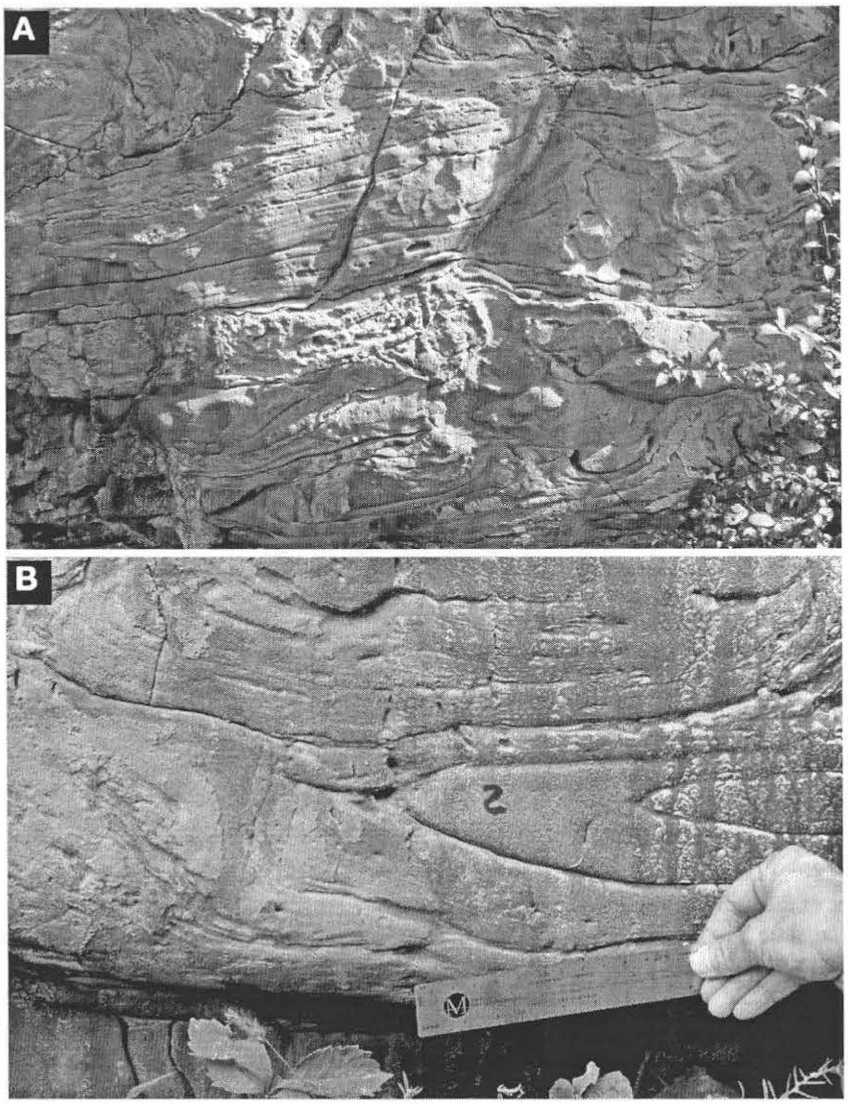
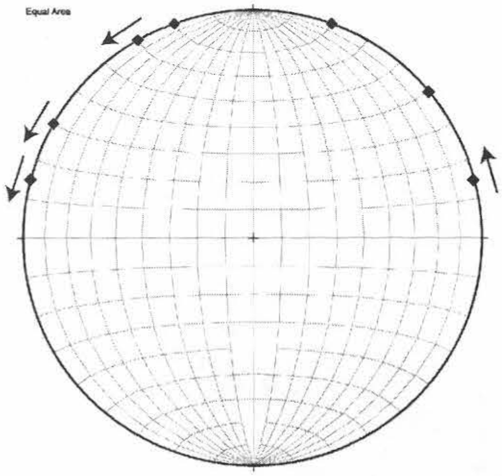
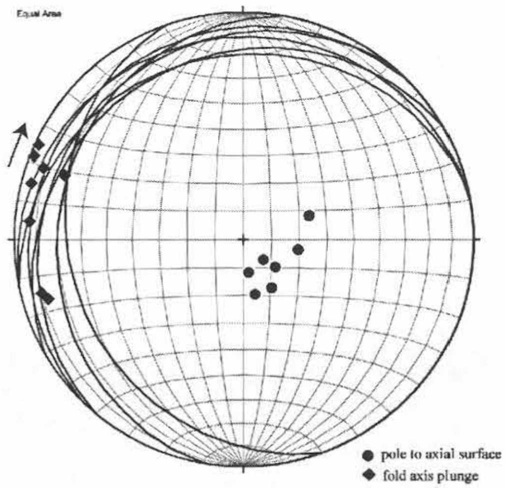


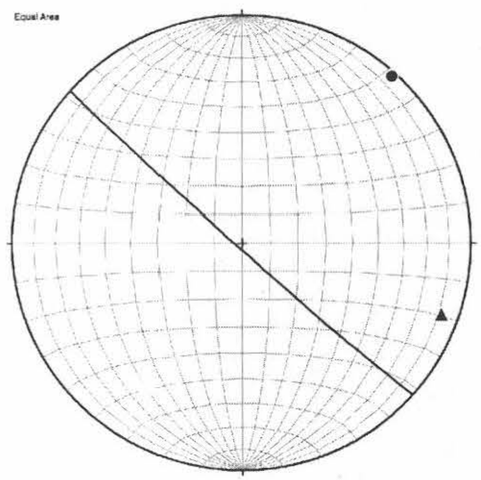
Figure 17. Outcrop of DeCew Dolostone at South Haul Road, Niagara Gorge, Lewiston, Niagara County, NY. A) General view of deformation. Note contorted beds of dolomitic shale intraclasts. B) Detail of overturned fold within lower DeCew.



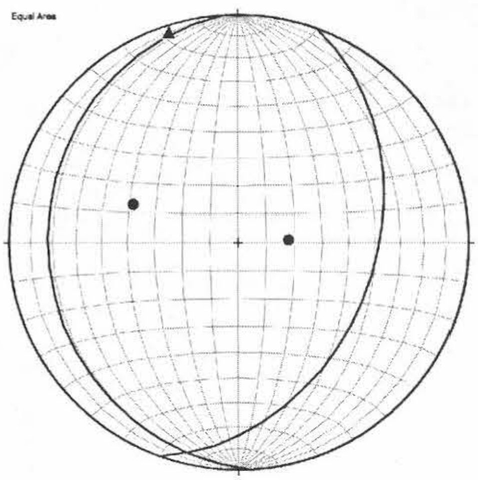
Haul Road (east side) fold axes. Axes bearing 285, 300, 330 have a NE-over-SW asymmetry (arrows indicate relative motion of upper plate).



Haul Road (west side) fold axial surfaces and axes. Arrows indicate relative motion of upper plate.



Haul Road (west side) open F2 axial surface and axis



Haul Road (west side, north end) fold axial surfaces and axes

Figure 18. Stereoplots of folds in the DeCew Member exposed along the Haul Road, Niagara Gorge. The upper left panel displays the bearing of fold axes (diamonds) observed along the east side of the Haul Road. Bearings with arrows indicate folds with asymmetry; the arrow indicates the sense of relative motion inferred from the asymmetry (relative motion of the upper plate). Note the apparently conflicting senses of motion. The upper right panel displays the disposition of axial surfaces (great circles), poles to the axial surfaces (solid circles), and bearing of fold axes (diamonds) observed along the southwest side of the Haul Road. Arrow indicates same as in the upper left panel. The lower left panel shows the axial surface of a later, open F2 fold that folds the folds in the upper panels. The lower right panel displays the disposition of axial surfaces (great circles), poles to the axial surfaces (solid circles), and bearing of fold axes (diamonds) of folds observed at the northern tip of the outcrop on the west side of the Haul Road.

such horizons have been identified in the Silurian Rochester Shale in the Niagara region. Sandstone beds full of eroded mudstone clasts and lingulid shell fragments in the Grimsby Formation may represent storm deposits in peritidal settings. The shale clasts may have been derived from banks of tidal channels by strong storm surge currents..

Among more distal events, several other types of tempestites have proved to be regionally extensive and traceable. For example, thick calcisiltites with hummocky cross stratification have been traced for tens of kilometers in the Rochester Shale. Large scale hummocky to swaly bedding locally in the DeCew Formation similarly reflect storm deposition of carbonate silts and fine sands. Bimodal cross-stratification of crinoidal sand and gravel in the Gasport Limestone may record superposition of tidal currents on storm surges (see Fig. 19).

Obrution Deposits

Another type of tempestite-related feature that provides very useful local markers are obrution or "smothered bottom", deposits. These are recognized taphonomically and may be traceable, at least locally. Excellent examples are provided by layers of beautifully preserved crinoids, trilobites, and other fossils (*Homocrinus* beds) that have been studied in detail from the lower Rochester Shale (Taylor and Brett, 1996). These rapidly buried surfaces have been correlated for tens of kilometers along outcrop strike of the Niagara Escarpment. The series of beds display much the same unusual characteristics over this area. Such evidence indicates that the smothering mud blanket was very extensive following a particular storm, resulting in mass mortality and burial on a regional scale.

LoDuca and Brett (1997) described laterally extensive horizons of extraordinarily preserved fossil green algae, annelid worms, and other non-skeletonized fossils from shaly dolostones in the base of the Goat Island Formation. Here rapid burial in organic-rich carbonate silts below an oxycline may have promoted preservation of soft bodied organisms as carbonized films.

Bioherms and Stromatolites

Organic buildups, including small coral-stromatoporoid or algal mud mounds, thrombolites, stromatolites, and larger scale reef structures, also appear to occur in very laterally extensive zones; these can be considered as "mounding events". For example, throughout western New York and Ontario, small fistuliporoid bryzoan-algal? mounds occur consistently near the top of the Irondequoit Limestone and project up to a meter into the overlying Rochester Shale (Cuffey and Hewitt, 1989). Nearly identical mounds occur at the top of the age-equivalent "Cryptothyrella bed" of the Bisher Dolostone of Ohio and its extension in the middle Osgood Limestone of Indiana (Brett and Ray, 2006).

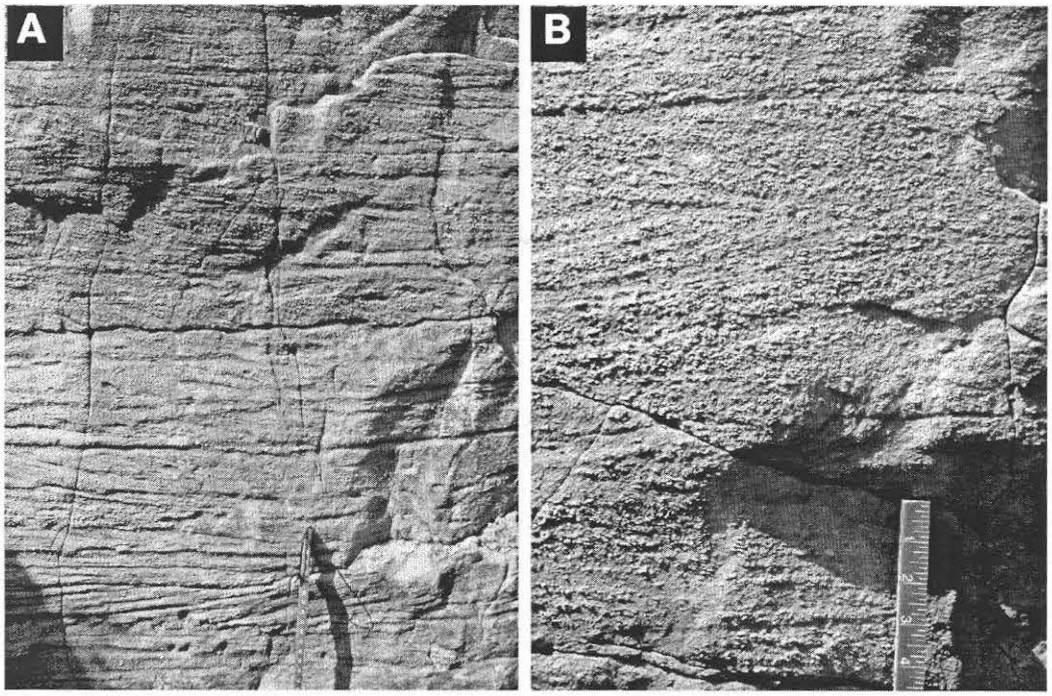


Figure 19. Outcrop of Gothic Hill Member grainstone at South Haul Road, Niagara Gorge, Lewiston, Niagara County, NY.

Larger, stromatoporoid-tabulate bioherms occur at two horizons in the Lockport Group of western New York and Ontario: the upper grainstones of the lower or Gothic Hill Member of the Gasport Limestone and extending upward up to 6m into the overlying thin-bedded argillaceous Pekin Member (Crowley, 1973; Brett et al., 1995); and in possible channel fills along an erosion surface at the top of the Gasport and extending upward into thin bedded middle Goat Island Formation (Brett et al., 1995). Not only are these horizons persistent over substantial distances in the New York-Ontario outcrop belt, similar thrombolitic mounds are present in probably correlative horizons in the McKenzie Shale in Pennsylvania to West Virginia (Brett et al., 1990). A comparable interval of bioherms has been recently recognized in the coeval upper Bisher in southern Ohio and Kentucky.

Finally, very persistent zones of large stromatolites and thrombolites (non-laminated algal mounds) occur at several horizons, especially near the base of the Guelph Formation (Brett et al., 1990, 1995) and in the lower Fiddlers Green Formation of the Bertie Group (Ciurca, 1990).

Of particular importance is the observation that mounds of varying type (stromatolites, thrombolites, coral-stromatoporoid bioherms) developed contemporaneously in different environments along an onshore to offshore gradient. What these features appear to have in common is that they are in laterally persistent horizons and occur either immediately above unconformities or on the flooding surfaces at the tops of skeletal sands. We suggest that the non-random distribution of such organic buildups in the stratigraphic record reflects the dynamic interaction of sea-level and organism growth. Mounding commonly appears to be associated with transgressing or deepening successions, in particular, during times of rapid deepening, as at maximum flooding surfaces. During these intervals rapid deepening created accommodation space and reef- or mound-forming organism may have built upward to keep pace with this increasing water depth. At the same time sequestering of sediments in coastal areas may have favored growth of algae and clonal organisms by reducing water turbidity and nutrient influx. Hence, widespread mound horizons are a signature of rising sea-level. In some cases the mounds were able to keep pace with deepening but in others they failed to keep up and were drowned. This explains the common burial of bioherms by thin, shaly sediments of deeper water facies.

SUMMARY

The Silurian strata of the Niagara Peninsula-western New York and their counterparts in the Cincinnati Arch region of southern Ohio and Kentucky, are richly fossiliferous, and display recurring depth-related benthic assemblages that aid in interpretation of relative sea-level fluctuation, as well as ecological-evolutionary history. Many fossil assemblages are exceptionally well preserved, reflecting event deposition, primarily widespread storm deposits. These fossils have also permitted relatively refined biostratigraphy.

Despite their classic status, these strata have only recently been considered from the standpoint of modern sequence and event stratigraphy (Duke and Fawcett 1987; Brett et al. 1990; Goodman and Brett, 1994; Brett et al., 1998). Newer approaches to refining our understanding of stratigraphy and facies relationships in these classic strata. These include identification and tracing of distinctive event beds, such as the DeCew seismite horizon and a hierarchy of disconformity bounded cycles of sequences. About nine large scale, 3rd-order, unconformity-bound stratal sequences, and more than 25 smaller 4th order subsequences, have been recognized within the Medina to Bertie groups (Llandovery to Ludlow Series) of the Ontario Peninsula-New York area (see Brett et al., 1990, for details. These sequences and many of their component subsequences can be correlated regionally into exposures on the east flank of the Cincinnati Arch in south-central Ohio and Kentucky (Dennison and Head, 1975; Brett et al., 1990; Brett and Ray, 2006). The persistence of similar patterns for over 700 km along the cratonic rim of the Silurian foreland basins demonstrates the existence of elongate belts of rather uniform facies parallel to the trend of the Algonquin-Findlay arch. Similarity of facies patterns was enhanced by the orientation of this belt, which during the Silurian lay approximately along the same latitude; but it also reflects allocyclic controls , including climatic and sea-level oscillations. Furthermore, some of the major events of relative sea-level fall and rise appear to be correlative with those recognized in other basins (see Johnson et al. 1985, 1998, for example), suggesting an underlying eustatic mechanism.

Nonetheless, detailed examination of regional patterns, within unconformity bound sequences permits recognition of far-field tectonics resulting from orogenesis in the Taconic mobile belt. During the Early Silurian these include a progressive eastward translation of the foreland basin axis and shoreline belts, by approximately 500 km, and the rise of the Findlay-Algonquin arch system, probably as a result of forebulge migration. This gentle differential uplift resulted in truncation of Llandovery sequences to the west-northwest in western New York-Ontario, as well as in west-central Ohio and Kentucky. During late Llandovery and Wenlock time the foreland basin and forebulge appear to roll back to the west. This abrupt reversal in basin axis migration, coupled with the influx of K-bentonites and the appearance of widespread seismites records renewed orogenic activity related to a little known medial Silurian (Salinic) tectophase.

ACKNOWLEDGMENTS

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STOP DESCRIPTIONS

Figure 20 shows stop locations for the trip on a simplified geologic base map (Rickard and Fisher, 1970). Abbreviations used in the text and figures for inferred sequence stratigraphic units are as follows: for subsequences (fourth order sequences): RLS = relative lowstands (regressive deposits); RHS = relative highstands; MFS = marine flooding surface; SDS = sea level drop surface; CI, condensed interval, and for sequences: SMT = shelf margin systems tract (lowstand deposit); TST = transgressive systems tract; CS = condensed section; EHS = early highstand; LHS = late highstand; SB = sequence boundary; TS = transgressive surface. The sequences referred to in the previous text are abbreviated as S-I through S-VIII in the following.

ROAD LOG FOR SILURIAN SEQUENCES, CYCLES, AND EVENTS

Total Mileage	Incremental Mileage	Description
0.0		I-290; take exit to Millersport Highway Junction NY 263 (Millersport Highway); turn left (northeast) onto Millersport Highway
9.5	9.5	Junction NY 78 (Transit Road); turn hard left (north) on Transit
10.0	0.5	Bridge over Tonawanda Creek; enter Niagara County
		Pass Niagara Wine Cellar
14.7	4.7	Pass Summit St/Lincoln Road these mark the approximate city of Lockport Limit and northern edge of ridges of Barre Moraine and possibly the Albion Moraine, which converges with it here
15.8	1.1	Cross Erie/NYS Barge Canal. Proceed north on North Transit crossing NY Rte 31
16.6	0.8	Turn left (west) on Outwater Drive and proceed to loop on Lockport Escarpment at Outwater Park.
17.4	0.7	Pull off in parking area adjacent to small overlook area

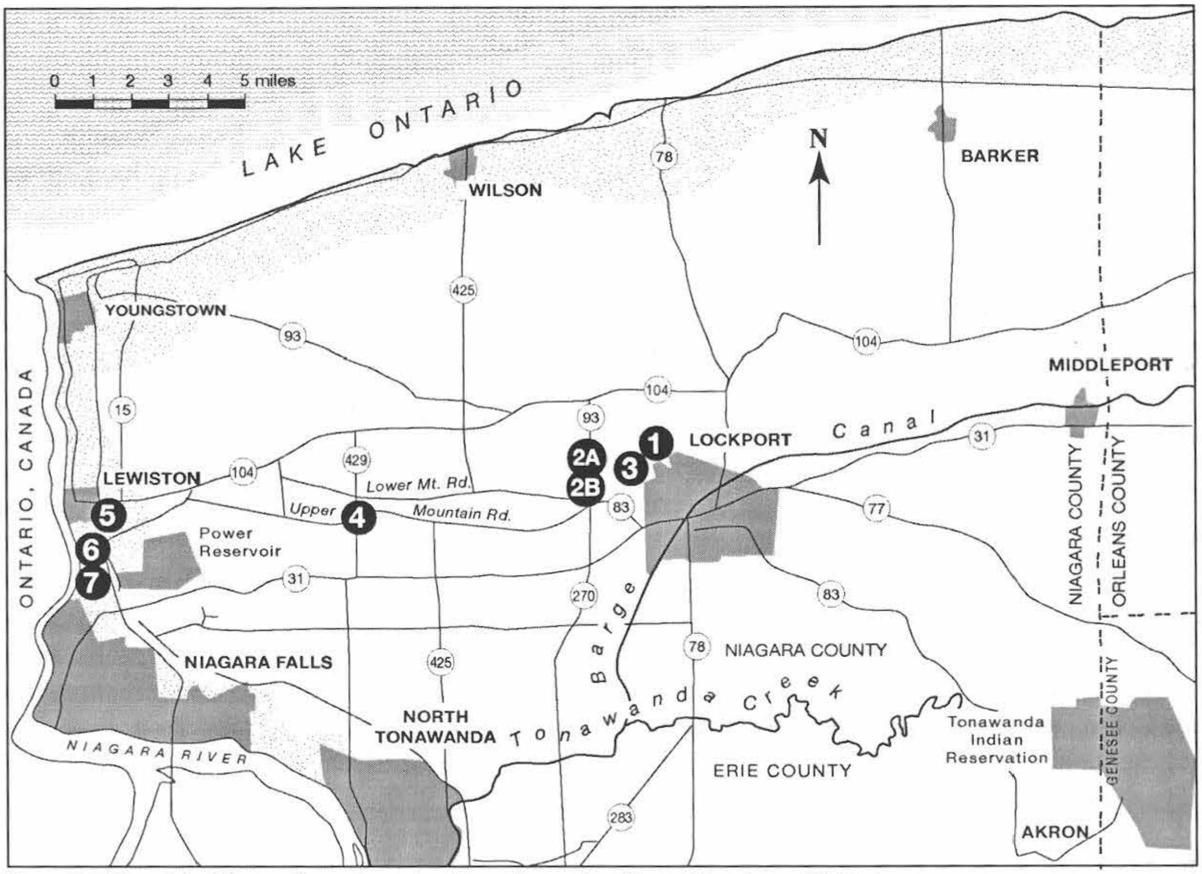

Figure 20. Map of the Niagara County area showing major roads, cities and locations of field trip stops.

OPTIONAL STOP CRAIN STREET QUARRY AND OUTWATER PARK (brief stop). View off Lockport (Niagara) Escarpment. View off Niagara Escarpment; spillways (occupied by East and West Jackson Roads) of ancient Lake Tonawanda into proglacial Lake Iroquois. Note Lake Ontario (about 18 km to north).

Walk back from overlook into old quarry at brink of escarpment to observed glacially polished surface of Pekin Member, Gasport Formation with small bioherms with well preserved stromatoporoids, favositids, and colonial rugose corals; pelmatozoan holdfasts are attached directly to bioherm surfaces. Patches of cross-bedded reef-flank debris are present at margins of bioherms.

		Retrace route to Transit Road.
18.1	0.7	Junction Transit Road; turn left (north); proceed north over escarpment.
18.2	0.1	Turn right (east) on Glenwood Drive. Adjacent hummocky deposits of cemetery may relate to Albion ice margin (moraine).

18.5 0.3 Take sharp left onto Gooding-Plank Street.

19.3 0.8 Junction West Jackson Street (on left) and Gooding-Plank Road at Lockport wastewater treatment plant. Pull off and park on Gooding near intersection.

STOP 1: ROAD CUT ON GOODING STREET BELOW SOMERSET RAILROAD VIADUCT (QUEENSTON SHALE—MEDINA GROUP)

This excellent 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' (1963) 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, Rhuddanian), siliciclastic wedge derived from tectonic source areas to the southeast (Figs. 5, 21). 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 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 transgress surface. The Whirlpool Sandstone thus is interpreted to contain either a lowstand (or shelf margin) systems tract and a transgressive deposit. A thin bed containing phosphatic pebbles and fossil grains occur with 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 upper Whirlpool into deeper shelf muds and storm sands of the Power Glen Formation, herein interpreted as highstand deposits. The Power Glen exhibits subtle 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 (2-10cm) muddy sandstones with interbedded sandy shales. Sandstone 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, the Devil's Hole Sandstone (Brett et al. 1995) 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 relative sea-level drops 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 bryzoans. 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 4,6).

Return to vehicles, turn left and proceed west on West Jackson Street.

20.7	1.4	Junction Niagara Street and West Jackson Street. Turn or continue west on Niagara Street.
22.2	1.5	Overpass over Lockport Junction Road (Route 93).
22.25	0.05	Access road to Route 93 on left; turn left onto road.
22. 3	0.05	Pull off along berm on left side of road and park. Proceed on foot directly down embankment along Route 93 to road cut.

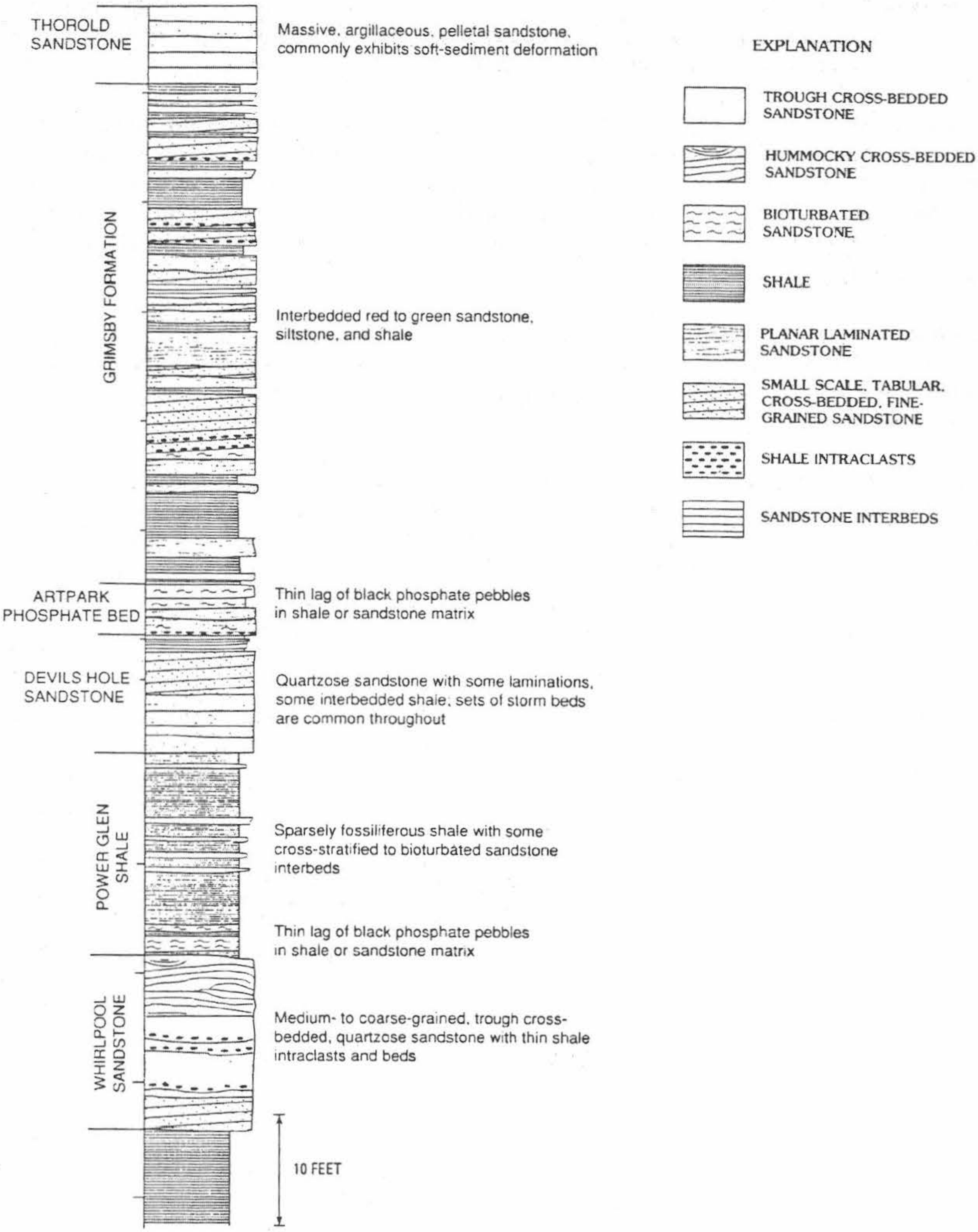

Figure 21. Generalized stratigraphic column for the Lower Silurian Medina Group in Niagara County, NY, showing revised terminology. From Brett et al. (1995).

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

This cut along Lockport Junction Road exposes upper units of the Medina Group and the lower part of the Clinton Group. 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 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 shaly unit that overlies the Thorold Sandstone and is traceable regionally at least as far as east as the Rochester area. Brett et al. (1995) termed this unit 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 actually, this is simply a leached sandy zone in the Cambria Shale. The Kodak Sandstone 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 and Fawsett, 1987).

Here the top of the Medina Sandstone is an erosion surface overlain by a thin (3 to 5 centimeter) dark gray, phosphatic sandy dolostone (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.

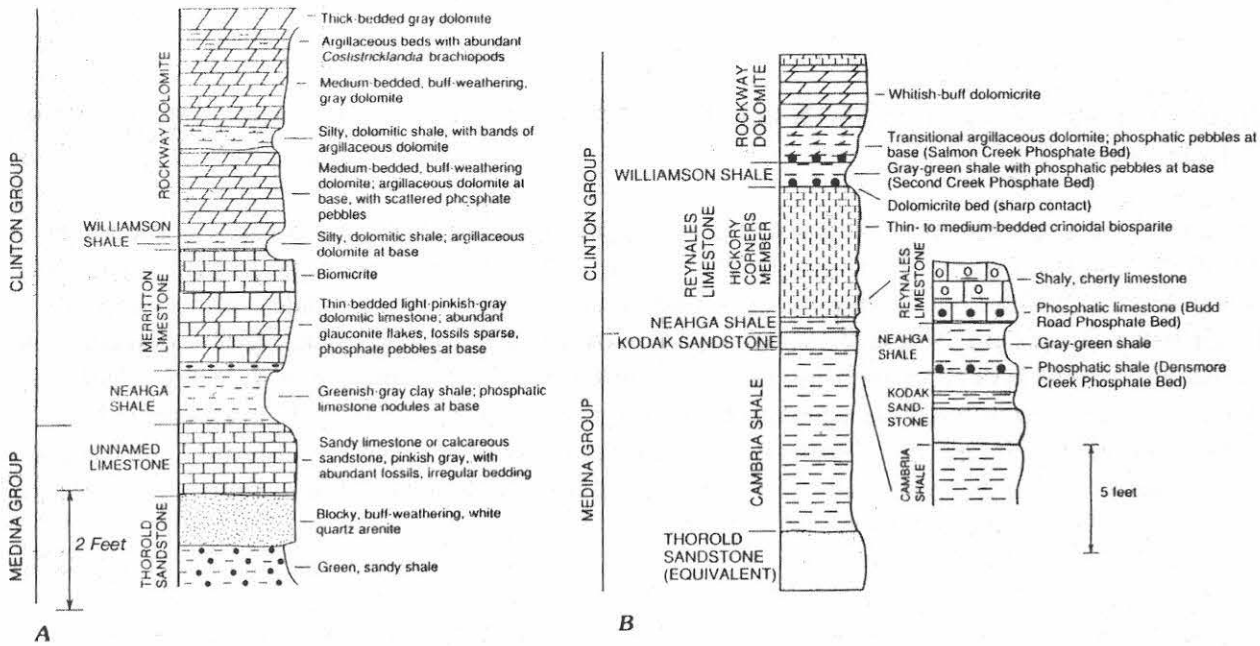

Figure 22. Detailed stratigraphy of the uppermost Medina Group (Sequence I), lower Clinton Group (Sequence II) and upper middle Clinton Group (Sequence IV) at outcrops separated by about 40 km. A) Western section; railway cut at Merritton (near St. Catharines), Ontario. B) Eastern section; Lockport, New York area. Note the loss of Cambria Shale between the two sections; also gradation of Densmore Creek phosphatic bed into unnamed limestone; truncation of the Reynales Limestone and intercalation of Merritton Limestone between Lockport and Merritton. Adapted from Brett et al. (1995).

At the base of the Hickory Corners Limestone (Reynales) is a thin (3 to 5 cm) pyritic, sandy limestone packed with black phosphatic pebbles and shell fragments (Budd Road phosphatic Bed). It is overlain by about 40 cm of alternating greenish gray shales and thin limestones capped by a 60 cm-thick ledge of nodular crinoidal pack—and wackestone at the top of the road cut. These beds contain a prolific fauna including corals, brachiopods (*Hyattidina, Dalejina, Platystrophia*) and pentameric crinoid stems belonging to a newly described disparid, *Haptocrinus*. 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

22.4	0.1	Junction with Lockport Junction (Townline) Road (Route 93); turn right (south) and proceed up escarpment
22.7	0.3	Pull into small parking area at gap in guard rail (for driveway) to left (Note; this parking is not suitable for large groups)

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

This large road cut 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 details as Stop 1 of Brett (1982 NYSGA Guidebook). 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 convoluted 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 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 bryzoan 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 270/93.

22.9	0.2	Upper end of road cut.
23.2	0.3	Junction Upper Mountain Road (Rte. 270/93). Turn right (west) on Upper Mountain Road.
23.9	0.7	Junction Thrall Road on right. Exposures of fossiliferous Rochester Shale along Thrall Road were described as Stop 2 in Brett (1982).
28.1	4.2	Junction Rt. 425.
30.1	2.0	Pekin Village; fire department on left.
30.2	0.1	Junction Old Pekin Road (one way). Turn right and proceed down

hill.

30.3	0.1	Junction Rt. 427; turn left (south) and move quickly toward right shoulder of Rt. 427
30.4	0.1	Pull off on wide area of right shoulder of Rt. 427 and park

STOP 3. "PEKIN BIOHERM" CUT ON RT. 427

Location: Cuts on both sides of NY 427, just north of underpass beneath Upper Mountain Road, Pekin, Niagara County, NY (Cambria 7.5' Quadrangle).
Description: This classic cut in the brow of Niagara Escarpment has recently been widened, and provides an excellent, but very fresh exposure of the upper Gasport (Pekin Member of Brett et al., 1995) and the lower portion of the overlying Goat Island Formation (Fig, 19). The large, massive mound exposed particularly on the west side of Rt. 427 was originally described as a Gasport bioherm and was considered to show reefal succession (Crowley and Poore, 1974). However, recent excavations have revealed that the mass of stromatoporoid bearing rock is almost entirely within the Goat Island Formation and is separated from older Gasport deposits (including a formerly exposed small bioherm, now destroyed by blasting) by a major erosion surface. The contact between the gray, thin bedded argillaceous dolostones of the Pekin Member of the Gasport and overlying massive, dolomitic crinoidal grainstone of the Goat Island (Niagara Falls Member of Brett et al., 1995), a sequence boundary, appears sharp. but horizontal in the northeastern portion of the cut. However, this contact abruptly descends to near road level just north of the Upper Mountain Road overpass. Here a mass of biohermal lithology overlies the unconformity. On the freshly blasted west side of the cut the irregular contact between the dark gray Pekin and light pinkish biohermal Goat Island is now very clear. The erosion surface separating the units has a relief of over 2 meters. The Goat Island "bioherm" appears to have developed in low areas on the unconformity. It consists of a mixture of crinoidal grainstones and light gray dolomicrite, rich in stromatoporoids, many of which are tumbled onto their sides or even inverted, indicating storm disturbance. This biohermal mass built up in "channels" along a sequence boundary during initial transgression.

		Reboard vehicles and proceed south on Rt. 429 through the underpass beneath Upper Mountain Road; prepare to turn right
30.6	0.2	Junction Grove Road; turn right
30.7	0.1	Junction Upper Mountain Road; turn left; Pekin, NY; note dolostone church on right
35.8	5.1	Dangerous curve on Upper Mountain Road at junction of Blackman Road; bear right continuing on Upper Mountain Road; view out to Lake Ontario
36.3	0.5	Leave Tuscarora Indian Nation

38.3	2.0	Junction Military Road at stoplight. Go straight onto Upper Mountain Road extension ; crossing I-190; avoid ramps to Lewiston-Queenston Bridge and Robert Moses Parkway
38.8	0.5	Exit ramp to Rte. 104 (Lewiston Road)West to Niagara Falls
39.0	0.2	Bear right onto Lewiston Road (Rte 104W); proceed south. Good view of Niagara Gorge to right
39.5	0.5	Pass forebay on left; driving over sluiceways of Robert Moses Power Plant
40.0	0.5	Pass Niagara University on left; Hyde Park Blvd at light
40.1	0.1	Junction Hyde Park Blvd. (NY Rt. 61) at light; bear left onto Hyde Park
40.3	0.2	Cross railroad track and prepare to make immediate right
		Junction South Haul Road; access for Robert Moses Power Plant; turn right
40.4	0.1	Pass by fisherman's parking area (cars can be left here; we will continue on South Haul Road to the bottom of the gorge and allow passengers to disembark and return uphill on foot); ahead through underpass beneath Robert Moses Parkway are excellent views of the Niagara Gorge and a complete stratigraphic section from the mid Silurian Lockport Group down to the unconformable contact between the Silurian Whirlpool Sandstone and the underlying Ordovician Queenston Shale
41.0	0.6	Fishermen's access at bottom of Niagara Gorge, just outside gates for Robert Moses Power Plant; passengers will disembark and proceed back up the South Haul Road; vehicles will return back up the road to park

STOP 4: NIAGARA GORGE: SOUTH HAUL ROAD

Location: Large road cuts in east wall of Niagara Gorge along South Haul access road for Robert Moses Power Plant, and ascending for about 1 km south to a tunnel beneath Robert Moses Parkway (Fig. 21). Parking is available in a fisherman's access parking lot just west of Rt. 62 (Hyde Park Boulevard) immediately south (uphill) from the tunnel. Lewiston, Niagara County, New York (USGS Lewiston 7.5' Quadrangle). Note: Access to the Haul Road exposure is strictly controlled by the Robert Moses Power Project and requires advance permission.

Description: This outstanding outcrop of Lower to mid Silurian strata shows important contrasts with sections of the comparable interval near Hamilton. The section begins near the Power Plant with about 8 m of the Upper Ordovician Queenston Shale (Fig. 21). Its sharp contact with the overlying Whirlpool Sandstone is the Cherokee Unconformity (Silurian Sequence I boundary).

The units of the Silurian succession are described in ascending order, as follows:

Medina Group (Fig. 21).

Whirlpool Sandstone: (4.5 m)- White, trough cross bedded, quartz arenite facies which record a non-marine to marine transition. Excellent profiles of channels are visible.

Power Glen Shale: (~ 8 m)- Dark gray, friable shale, with very minor sandstone interbeds

Devils Hole Sandstone: (2 m)- Pale gray, massive, quartz arenite with a distinctive, meter-thick phosphatic, sandy dolostone, Artpark Phosphate Bed near the top (IC).

Grimsby Formation: (15 m)- Greenish gray to maroon shales and mudstones with bundles of thin reddish and white mottled sandstones (Fig. 21).

Thorold Sandstone: (2 m)- White, cross-bedded quartz arenite. The Thorold has a sharp, erosive base which marks the base of the next Medina subsequence (IC). A thin (2-10 cm), sandy phosphatic bed, the Densmore Creek Bed (Brett et al., 1995) rests sharply on the Thorold (and on a Cambria Shale remnant north of the Power Plant), marking the base of the Neahga Shale. Clinton Group (Fig. 18)

Neahga Shale: (2 m)- Dark greenish gray, very friable shale (base sharp (II SB).
Reynales Formation (Hickory Corners Member): (~50 cm)- Medium gray, nodular, burrowed, bryzoan-rich wacke- to packstone. Conodonts indicate a mid Llandovery age for the Reynales (see LoDuca and Brett, 1994); this unit represents an erosional remnant of the Reynales.

Rockway Formation: (3 m)- Buff-weathering, argillaceous dolostone with thin dolomitic shales shows prominent rhythmic bands (10-50 cm) of sparsely fossiliferous argillaceous dolostone interbedded with thin gray shales. The Rockway shows a sharp upper contact(S-IV-V SB).

Irondequoit Formation: (2.5 m)- Massive, pinkish-gray, crinoidal pack-and grainstone. Clasts of fine-grained dolostone, derived from the underlying Rockway occur in the basal thin bed of the Irondequoit (Fig. 23). Its sharp upper contact (MFS) is marked by a 30 cm thick shell bed.

Rochester Shale: (18 m)- Medium dark gray mudstone with thin calcisiltites and lenticular fossil rich limestones; 1.5 m of bryzoan-rich limestone beds underlie the sharp top (MFS) of the Lewiston Member . The upper Rochester (Burleigh Hill Member) also displays a sharp contact (SSB) with the enterolithic DeCew Dolostone beds.

DeCew Formation: (3m) -Dark gray, buff weathering, laminated dolostone (calcisiltite). Here and, especially in weathered exposures in the adjacent Devils Hole Park section, the DeCew displays spectacular soft-sediment deformation with isoclinally folded beds (seismite?).

Lockport Group (Fig. 24): 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; basal Gasport shows rip-up clasts of DeCew Dolostone (Fig. 17).

Gasport Formation: (5 m)-Pinkish gray thin bedded to massive dolostone, divided into a lower pinkish gray dolomitic crinoidal grainstone (Gothic Hill Member) and a 2.5 m upper argillaceous, bioturbated dolostone (Pekin Member) Weathered surfaces of the Gothic Hill grainstones display probable bipolar cross- stratification . The sharp upper contact is a subsequence boundary.

Goat Island Formation : (~10 m)- Buff weathering dolomitic, crinoidal grainstones, buff, thin bedded dolostone with white chert, and dark brownish gray, argillaceous, banded dolostone; the basal unit (Niagara Falls Member) is massive crinoidal grainstone with scattered *Cladopora* corals and stromatoporoids; abundant vugs appear to be solution cavities in a stromatoporoid-rich zone. The Lancaster Member is poorly developed here, thin (-2.5 m) and only sparingly cherty. Argillaceous and bituminous gray dolostones of the Vinemount Member form the uppermost unit on the access road. This member is much less shaly than at its type area near Hamilton.

Return to vehicles and proceed

41.6	0.6	<u>Turn right</u> out of fisherman's parking lot
41.7	0.1	Junction Hyde Park Boulevard; again, proceed straight across onto University Road
42.4	0.7	as road curves to right; <u>turn left</u> onto unnamed access road; note jungle of high tension wires ahead
42.8	0.4	T-intersection with Robert Moses Power Vista back entrance road; <u>turn right</u>
42.9	0.1	<u>Pull off and park</u> along road cuts just before underpass beneath I-190

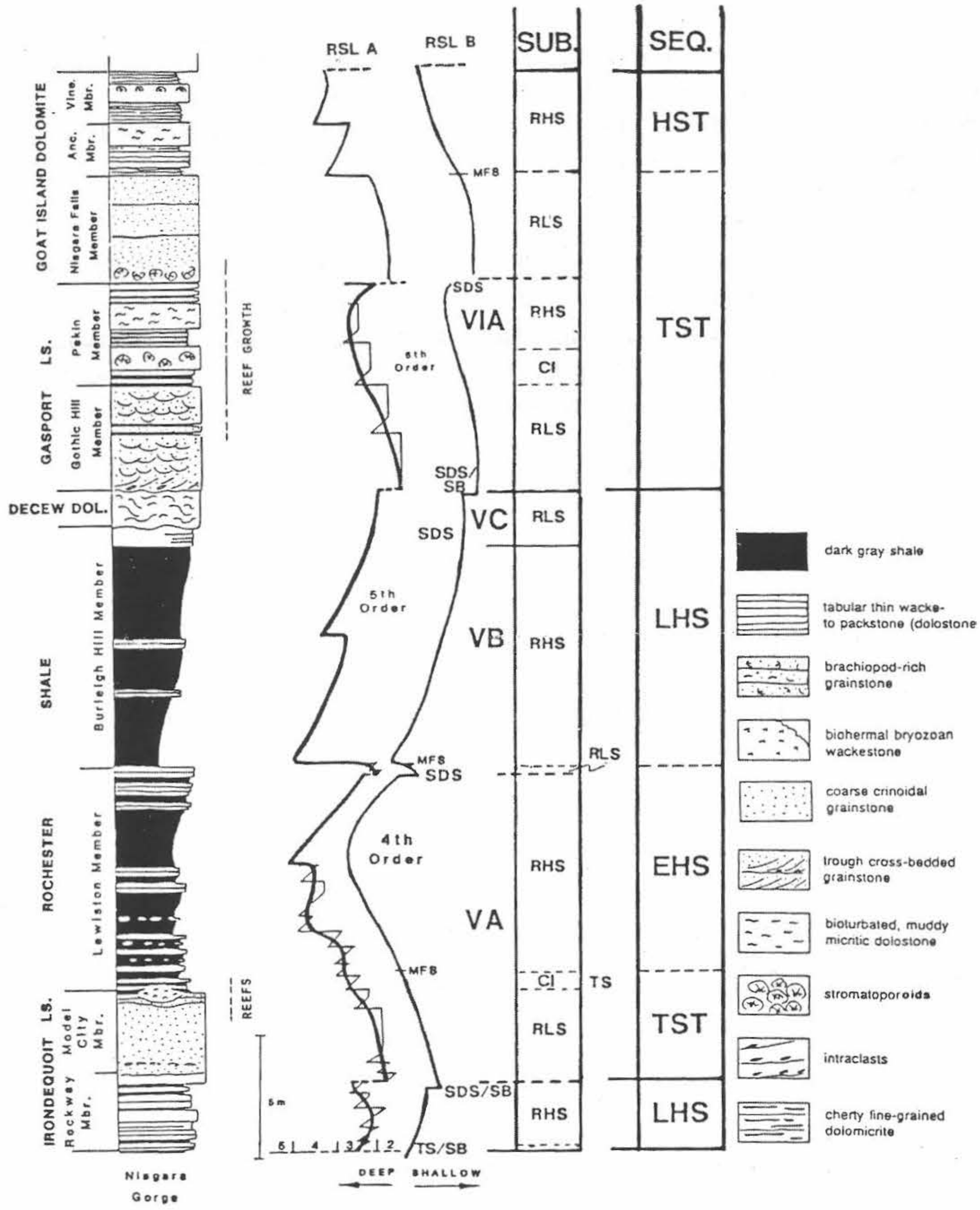

Figure 23. Stratigraphy, sequence stratigaphic interpretation and inferred sea level curve for the upper part of the Clinton Group (SequenceV) and the lower Lockport Group (Sequence VI) in Niagara County. Sequence stratigraphic abbreviations: CS: condensed section; EHS: early highstand systems tract; LHS: late highstand (or falling stage) systems tract; SB: sequence boundary; TST: transgressive systems tract. Adapted from Brett et al. (1990).

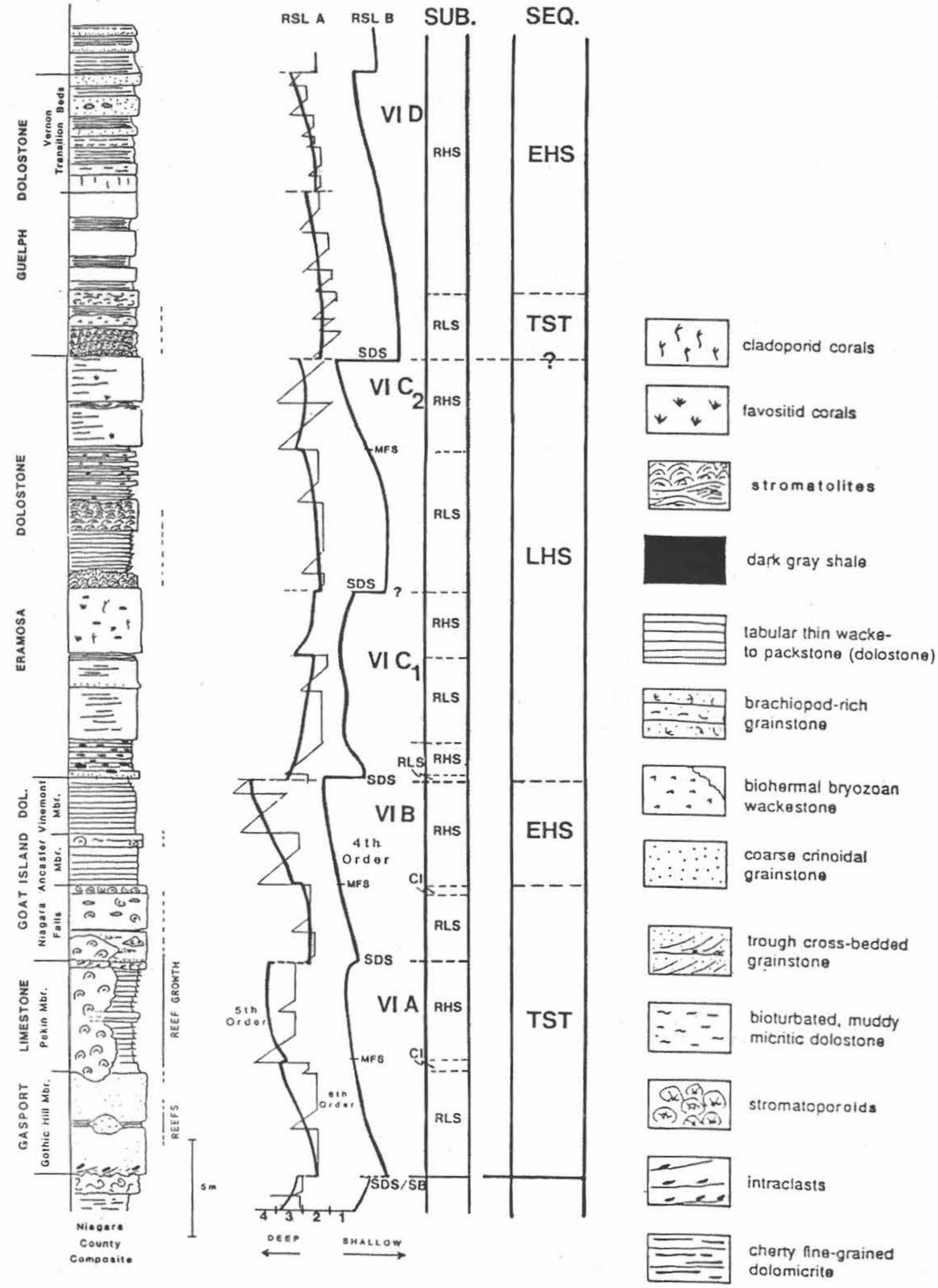

Figure 24. Stratigraphy, sequence stratigaphic interpretation and inferred sea level curve for the upper part of the Clinton Group (SequenceV) and the lower Lockport Group (Sequence VI) in Niagara County. Sequence stratigraphic abbreviations: CS: condensed section; EHS: early highstand systems tract; LHS: late highstand (or falling stage) systems tract; SB: sequence boundary; TST: transgressive systems tract. Adapted from Brett et al. (1990).

STOP 5 : ROBERT MOSES ACCESS ROAD AND FOREBAY

Location: Access road to the Robert Moses Power Plant, and adjacent forebay canal just west of Military Road and 1.2 km north of Route 31, Lewiston, Niagara Co., NY (Lewiston 7.5' Quadrangle).

Description: Higher units of the Lockport Group are visible along the forebay and consist of the upper Eramosa (new usage in New York) and lower Guelph formations (Fig. 24). Large algal bioherms characterize the uppermost units of the Eramosa Formation in the forebay. Exposures in the small road cut at the underpass of the access road beneath the lanes of I-190 show exceptionally large (2 m high) stromatolites, and some non-laminated thrombolites the stromatolitic interval at the base of the Guelph is traceable with subsurface data at least to Hamilton (Fig. 24).

Pull forward under the I-190 overpass and proceed to Military Road

43.1	0.2	Junction Military Road; turn left (north)
43.6	0.5	Crossing east end of Robert Moses forebay area; Pump Generating Power Plant to right; note small thrombolitic mounds in Eramosa Formation exposed in forebay cuts to left
44.5	0.9	Junction Upper Mountain Road at stoplight
45.2	0.7	Junction NY 104; turn right (north)
45.5	0.2	Brow of Niagara Escarpment; on clear days good views of the mouth of Niagara River and Lake Ontario to the north; cuts to right in Gasport-Goat Island formations
45.9	0.4	Skip Exit for Rt. 18F east, but stay right
46.4	0.5	Exit for Rt.104-18F west; bear right onto exit lane
46.6	0.2	Merge right onto Rt. 18F; Center Street, Lewiston, NY
47.1	0.5	Portage Road; turn right and proceed to entrance for Artpark
47.3	0.2	Entrance to Artpark Road
48.0	0.7	Parking Lot B; bear right and then turn left immediately once in parking lot; proceed up to left on small road labeled "Narrow"

near Clay Studio

48.2	0.2	Fishermen's parking area; pull in and park; proceed south on foot to entrance of Niagara Gorge

STOP 6. NIAGARA GORGE, LEWISTON (ARTPARK)

Location: Sections in east wall of Niagara Gorge along old haulage road extending from north end of Niagara Gorge at Niagara Escarpment just south of Artpark, off Fourth Street, southward to the Lewiston-Queenston bridge, Lewiston, Niagara County, NY (Lewiston 7.5' Quadrangle)

Description: The north-facing cuesta (Niagara Escarpment) stands 76 m above the adjacent Lake Ontario plain. Niagara Falls was initiated here about 12,000 BP.

Exposures of the Upper Ordovician Queenston Shale and its unconformable contact with basal Silurian Whirlpool Sandstone are visible along a short path, adjacent to the river, immediately south of the Artpark theater. Outcrops of the Lower Silurian stratigraphic units above the Whirlpool Sandstone are accessible along an old haulage road that leads southward from the Artpark Visitor Center into the gorge. At the entrance to the gorge (edge of Niagara Escarpment) an isolated "butte" of Lower Silurian strata (Power Glen-lower Grimsby; type locality of the Artpark Phosphate bed) between the path and the river, represents a remnant of a promontory in the gorge wall that was breached during excavation for the power plant. A 0.5 m layer of mottled sandstone with prominent ball-and-pillow structures, about 5.5 m above the base of the Grimsby, is well exposed in the cliffs about 200 to 700 m north of the Lewiston-Queenston Bridge. Caution is required in this exposure, as rock falls are common. Abundant fallen debris also provides an excellent look at varied lithologies of the upper Medina, as well as Clinton and Lockport units.

Reboard vehicles and retrace route to parking lot B

48.4	0.2	Bear right out of parking lot onto exit road and proceed to exit from park
49.1	0.7	Proceed straight through onto Portage Road
49.5	0.4	Junction Rt. 18F; turn right (east)
49.7	0.2	Junction Rt. 104 West/Robert Moses Parkway; turn right
49.8	0.1	Fork; bear left onto entrance for Rt. 104; IF going to Canada DO NOT go onto Robert Moses Parkway

NOTE: If time permits we will take a short side trip to optional stop for Whirlpool State Park; approximately 4 miles south of Robert Moses Parkway.

END FIELD TRIP

AN EARLY LATE DEVONIAN BONE BED-PELAGIC LIMESTONE SUCCESSION: THE NORTH EVANS-GENUNDEWA LIMESTONE STORY

In memory of Daniel B. Sass (1919-2006)

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INTRODUCTION

The Genesee Group succession in western New York represents deposition in dysoxic to near-anoxic settings in the subsiding Appalachian foreland basin (Fig. 1) that records a series of widespread sedimentary and biological events. We will focus on lower and medial Genesee divisions that record changes in the basin that range in age from latest Middle Devonian into the early Late Devonian. In particular, we are interested in the genesis of styliolinid (pelagic) limestone units and associated bone-conodont beds. This paper will examine the results of recent mapping and biostratigraphic work on a succession of units in the Genesee Group, ranging, in upward succession, from basal Geneseo black shale facies into the lower part of the West River Formation. We will also highlight the discovery of two mappable discontinuity-related bone-conodont beds in the lower part of the Genesee succession. The most important part of this project is the study of the regional character and inferred genesis of the North Evans Limestone (a famous conodont-bone lag unit above a major discontinuity) as well as that of the Genundewa Formation, a distinctive layer of pelagic, styliolinid carbonate that overlies the North Evans. We will examine closely the relationship of the Genundewa to both the North Evans lag deposit and overlying basal beds of the shaley West River in the context of inferred flexural and eustatic events within the basin. Another focus of this project is to better elucidate the complex chronostratigraphic story recorded in this stratigraphically condensed succession.

The change from Middle Devonian (Late Givetian), neritic shale facies of the Windom Member (Moscow Formation) of the Hamilton Group into black, condensed, basinal deposits of the lower Genesee Group [latest Givetian and earliest Late Devonian (basal Frasnian)] is abrupt and profound (Fig. 2). This change, across the widespread top-Windom, Taghanic disconformity in western New York was what initially lured two of us (Baird and Brett) away from fossil-rich Hamilton deposits into a mysterious parallel universe of black mud deposition and hypoxic water mass chemistry. The study of the genesis of exotic facies such as the diachronous, detrital pyrite deposits of the Leicester Pyrite along the disconformable base of the Geneseo black shale marking the base of the Genesee Group (see Fig. 2) was critical in understanding the genesis of black shale-roofed unconformities (Baird and Brett, 1986, 1991; Baird et al., 1989). Although work relating to the genesis of such contacts was directed to basinal units elsewhere in succeeding years, detailed study of the Genesee Group above the Taghanic unconformable contact has continued. Detailed biochronostratigraphic studies of Genesee divisional units are also ongoing. Application of new advances in sequence stratigraphy and recognition of the importance of basin flexural tectonics in the evolution of the Devonian foreland basin

(Ettensohn, 1987, 1994, 1998), make the condensed western New York Genesee Group succession ideal for the testing of models (see Figs. 1, 2).

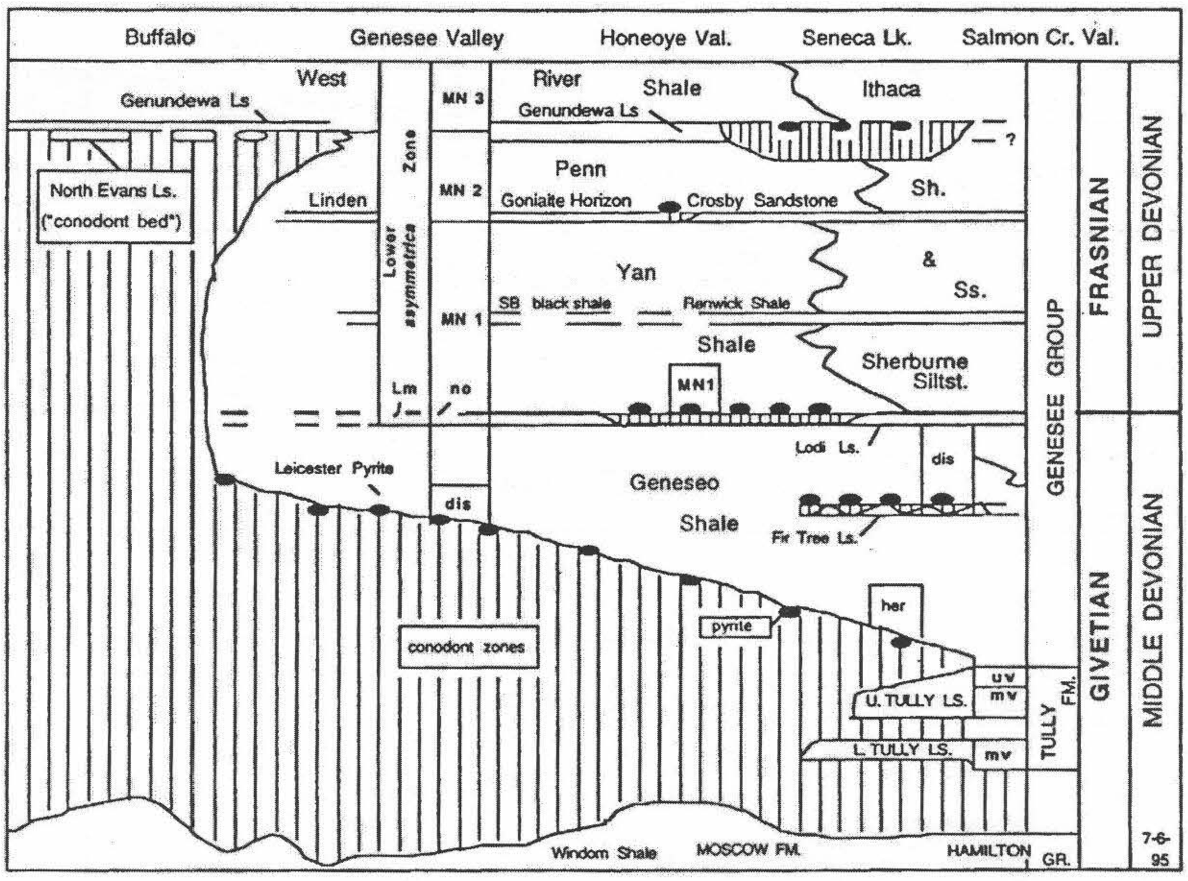
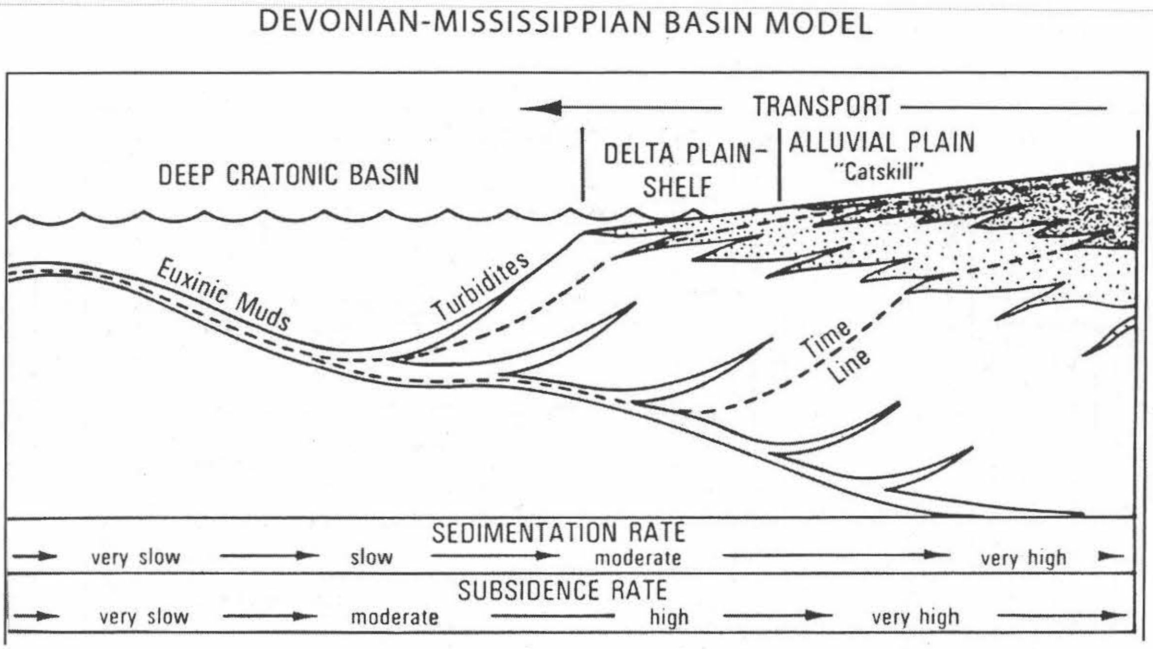


Figure 1. Generalized chronostratigraphic cross-section of lower Genesee Group and subjacent Moscow Formation (Hamilton Group, Windom Shale Member). Large hiatus below Genesee Group marks position of compound Taghanic Unconformity, Genesee onlap succession, and sub-Genundewa unconformity. The pre-Tully erosion- Hamilton erosion surface marks a major sequence and tectophase (III) boundary. The lenses of detrital Leicester Pyrite are derived from this erosion but were deposited through a long period of diachronous onlap of Genesee black muds from this discontinuity during the Taghanic transgression. The locally beveled beds with pyrite and fish debris include condensed styliolinid limestones and nodules (Fir Tree, Lodi, Abbey, Linden, Crosby, Genundewa) associated with surfaces of maximum sediment-starvation formed during pulses of sea level rise. In this report these horizons have been traced to the most highly condensed westernmost sections. The North Evans Limestone (conodont bed) in the Buffalo area in western Erie County is a lag deposit of crinoid, fish, and conodont debris that accumulated in shallow water over the peripheral bulge at the west margin of the basin where the gap of the compound unconformity is greatest. Lenses of North Evans debris [with Middle *varcus* to MN Zone 2 conodonts and Frasnian goniatites (*Koennites*)] are traceable beneath the sub-upper Genundewa disconformity as far east as the Genesee Valley. From Kirchgasser, Brett and Baird, 1997, fig. 7).



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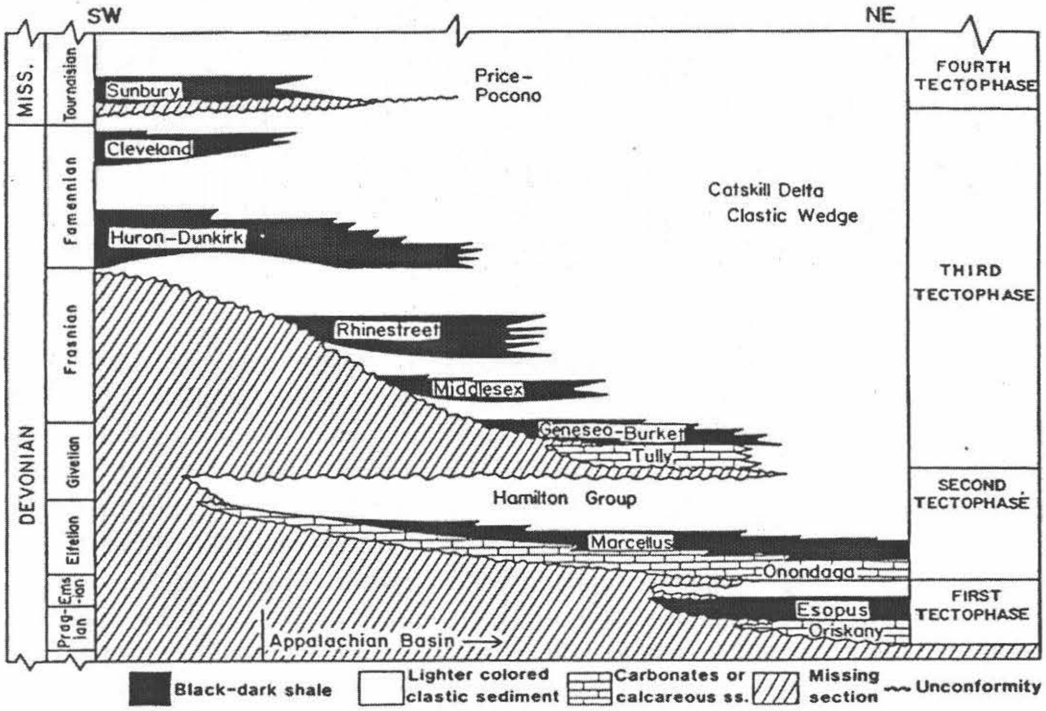

B

Figure 2. A. Idealized east-west cross-section and depositional model through the Catskill Delta complex. From Broadhead et al. (1982). **B.** Stratigraphic transect showing lithostratigraphic units in an inferred flexural foreland basin setting across the Appalachian Basin. From Ettensohn (1994).

Herein, we will examine several component units and boundaries within the lower and medial parts of the Genesee Group in Erie County and western Genesee County (Figs. 4-9). Some of these are newly recognized owing to renewed mapping work across this area. In particular, we will focus on regional aspects of, and inferred origin of the medial Genesee North Evans Limestone and overlying Genundewa Formation (Figs. 8-14). The first part of the paper summarizes the succession of divisional Genesee units based on new stratigraphic work across Erie and western Genesee counties. The second part focuses on new discoveries in the North Evans-Genundewa-basal West River shale succession that bear on the genesis of this distinctive, time-rich interval. This paper represents a tribute to the pioneering work of Dan Sass, who, 45-47 years ago, traced the Genundewa Limestone across western New York and measured many of the sections described herein as part of his Masters Thesis project at the University of Rochester (see Sass, 1951). Dan, who served in World War II, was active in the fields of Environmental Science and Limnology and was instrumental in establishing both a Geology undergraduate major and an Environmental Studies program at Alfred University. He died this past March.

GEOLOGIC SETTING

Acadian orogenic uplift in New England and the central Atlantic region was associated with progradational developmemt of the Catskill Delta Complex which filled the Devonian foreland basin from east to west (see papers in Woodrow and Sevon, 1985; Fig 1). Catskill Delta progradation began in earnest during deposition of the Middle Devonian Hamilton Group following onset of the second collisional tectophase, but accelerated significantly during the third tectophase (Ettensohn, 1998). Not only do strata above the Taghanic disconformity thicken greatly to the east, but they also grade spectrally eastward and shoreward into variably fossiliferous neritic facies that are typically much coarser (Fig. 1).

During the late Middle Devonian, the study area was in the southern hemisphere tropics or subtropics and was covered by an epicontinental sea (Scotese, 1990). Deposits seen on this fieldtrip accumulated on the northern margin of a subsiding foreland basin that periodically expanded and deepened during phases of oblique collisional overthrusting (tectophases) associated with the ongoing Acadian Orogeny (Ettensohn, 1987, 1998). The most pronounced thrust loading event of tectophase three coincided with the onset of the deposition of the Genesee Group (Fig. 1); this flexural drowning event was also largely coincident with a major rise in sea level within T-R cycle IIa of Johnson et al. (1985). In west-central New York this deepening is expressed by lithologic change from shelf carbonates of the Tully Formation into basinal black shale deposits of the Geneseo Formation (Heckel, 1973; Baird and Brett, 2003). In western New York the Tully Formation is absent due to erosional/corrosional processes and progressively younger divisions of the Genesee Group (Geneseo Formation, Penn Yan Formation, and condensed North Evans/Genundewa deposits) are observed to successively onlap the Taghanic Unconformity, a major regional disconformity, in a westward direction (Fig. 2). This disconformity and associated detrital pyritic lag debris (Leicester Pyrite Member), separating fossiliferous neritic facies of the late Middle Devonian (Middle Givetian) Windom Member of the Hamilton Group from overlying dysoxic to near anoxic Genesee deposits (see Figs. 2, 3) will be seen at STOPS 2 and 3. In west-central New York localities, the gradational transgressive change from Tully carbonate facies into black shale of the Geneseo Formation coincides with the latest Middle Devonian (Huddle, 1981; Kirchgasser et al. 1989; Figs. 2, 3).

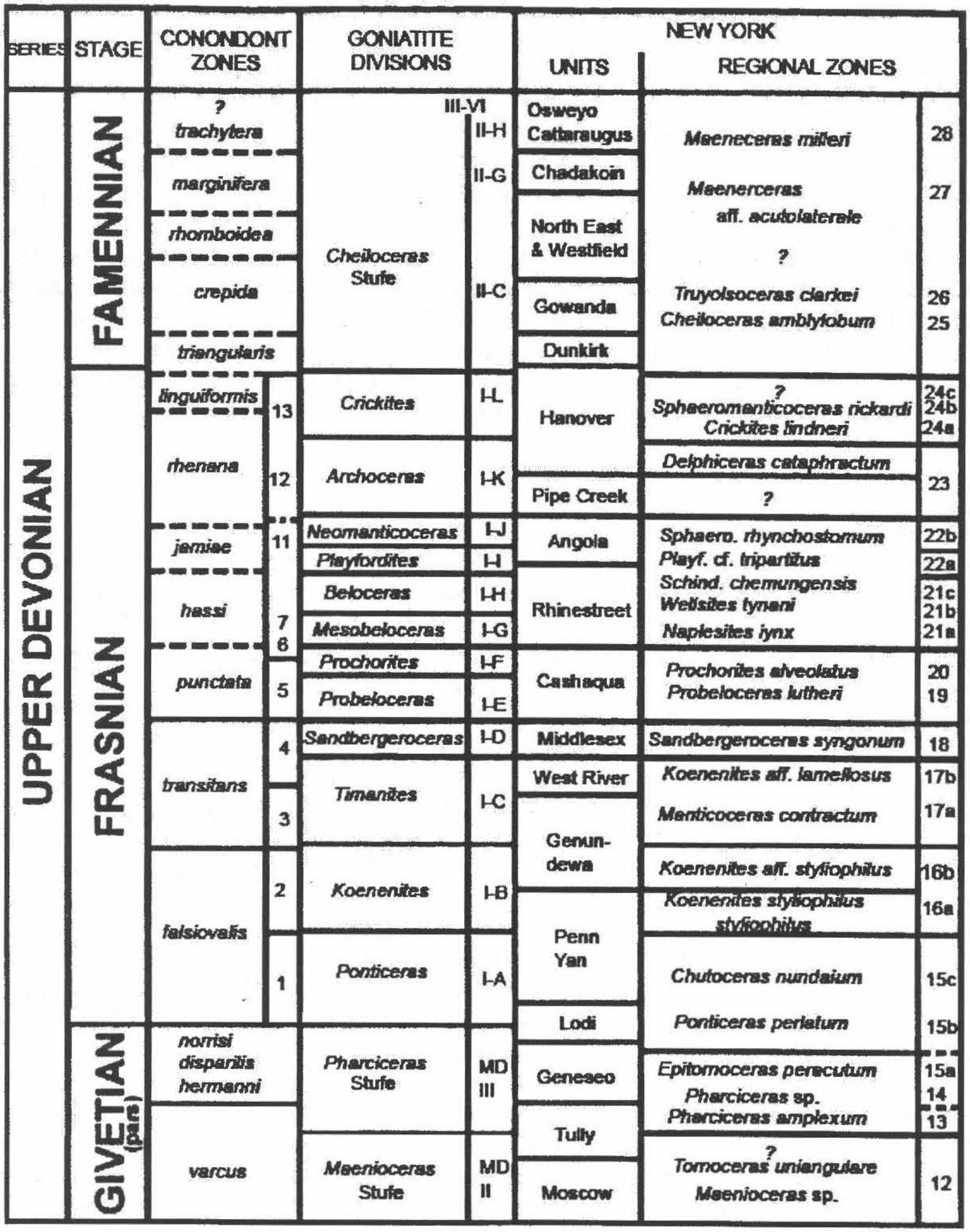

Figure 3. Late Devonian (Givetian, Frasnian and Famennian) stratigraphic succession in New York State showing alignment to international conodont zones [Standard and Montagne Noire (1-13)] and goniatite cephalopod divisions and New York regional goniatite zones (12 to 28). (Modified from House and Kirchgasser, 1993).

Proceeding westward along the Taghanic disconformity, the age of the onlapping black shale deposits become progressively younger into eastern Erie County; this reflects the regional flexural-eustatic Taghanic event (Kirchgasser et al., 1989; Baird and Brett, 1986). A younger erosion surface, associated with the North Evans conodont – bone lag below the Genundewa Limestone, oversteps the Taghanic disconformity in Erie County, thus merging the two discontinuities into a composite major unconformity (Figs. 2, 3). At STOP 1 (Eighteenmile Creek), the Late Devonian (early Lower Frasnian) North Evans Limestone rests directly on late Middle Devonian (Late Givetian) shales of the Windom Member with at least six conodont chronozones either missing or whose representative taxa are present in a range of preservations in the reworked and transported debris (Huddle, 1981, Kirchgasser et al. 1989, Kirchgasser and Koslowski, 1996, Kirchgasser, 1994, 1996, 1998, 2001, 2002, 2004). The taphonomic age of the North Evans (age when the debris came to final rest and the unit to lithification) is the upper part of Montagne Noire (MN) Zone 2, based on the occurrence of *Ancyrodella recta,* the youngest zonal indicator in the unit.

Baird and Brett (1986, 1991) discussed a variety of mechanisms capable of producing coarse traction-generated, detrital pyrite lags such as that represented by the base-Geneseo Leicester Pyrite in a deep-water, nearly anoxic regime. Processes including deep-storm wave gyre impingement, bottom current processes, and internal waves were considered as mechanisms capable of moving coarse particles at depth. We tentatively settled on a model of internal wave-shoaling against a sloped basin substrate as a possible traction mechanism; in this scenario, internal waves generated along the pycnocline, or water mass boundary, within the water column, eventually shoal against the basin margin slope resulting in erosion and sediment traction (Baird and Brett, 1991). This fits into the black shale onlap scenario in that this erosion occurs on the Taghanic disconformity slope prior to slope burial by black mud; as water deepens, owing to sea level-rise and/or flexural subsidence, the zone of pycnoclinal erosion continually migrates westward in the upslope direction ahead of black mud onlap which takes place within a lower energy, lower dysoxic substrate regime below the pycnocline (Baird and Brett, 1986, 1991). On the eastern margin of the basin, a parallel, pyrite mantled discontinuity appears in basin margin areas and dies out up-ramp into near continuity within dark gray, silty mudstone facies. Hence, erosion appears to commence earlier in the basin and migrates up-ramp in association with a rising pycnocline; such that discontinuities pass upslope to continuity beyond the highest reaches of the pycnocline during maximum highstand, a pattern opposite to unconformities generated during falling stages of sea level.

Westward flexural basin expansion during Geneseo Shale deposition would account for east-to-west slope drowning and conveyor belt-type pycnocline migration and subsequent sediment onlap across a 100 km lateral distance across western New York (Figs. 1, 2). Reworked calcareous fossils and diagenetic carbonate debris reworked from the underlying Windom Shale on the east-sloping sediment-starved Taghanic erosional ramp would start out as calcareous lag material in a shallower water wave-influenced oxygenated regime. Subsequence slope drowning with consequent overspread of dysoxic water below the rising pycnocline was postulated to explain the dissolutional transformation of the lag material to a residual placer of pyrite and other insolubles. Since the zone of pycnocline impingement was always upslope from the mud onlap limit during deposition of the Geneseo, the basal Geneseo lag would always be made up of insoluble material (Baird and Brett, 1986).

STRATIGRAPHY OF LOWER GENESEE GROUP: GENESEO AND PENN YAN FORMATION DIVISIONS

Overview

Recent mapping of the thin, stratigraphically condensed, deposits of the lower-medial Genesee Group succession (Geneseo, Penn Yan, Genundewa, and basal West River Formations) across Erie and Genesee counties has led not only to more detailed delineation of regional trends in the interval, but to discovery of new component units and erosion surfaces between units (Figs. 4-8). Generally, units thicken eastward from an extremely condensed condition in Erie County sections. Sections in southeastern Genesee County contain proportionately more gray shale relative to black shale reflecting more terrigenous influx from the east, but also reflecting better oxygenated bottom conditions than facies seen in Erie County. Westward changes from gray to dark gray shale at several levels appears to be most pronounced near Linden, on the trend of the Clarendon-Linden fault zone. Presented below is an outline description of contacts and divisions that will proceed from oldest upward.

Leicester Pyrite

The Leicester Pyrite is a detrital pyrite lag deposit that occurs along the Taghanic disconformity and, in rare instances, slightly above. This unit tracks the stages of the Taghanic onlap sequence. Typically it is an accumulation of coarse gravel-grade pyrite and pyritic steinkerns (including the goniatite *Tornoceras uniangulare)*, mostly derived from the underlying Windom Shale, that is admixed with other insoluble components such as conodonts and fish bones. Leicester debris occurs in laterally disconnected lenses along the contact; these lenses may represent erosional channels on the exposed submarine erosion surface (Figs. 4-6). Dissolution of carbonate material is usually complete owing to pervasive near-anoxia, punctuated by episodes of bottom aeration and monosulfide oxidation which periodically lowered pH levels in the vicinity of lag deposits (see Baird and Brett, 1986, 1991). We will see the Leicester Pyrite at Cazenovia Creek (Loc. 11; STOP 2) and at a tributary of Cayuga Creek (Loc. 14b; STOP 3). Localities cited in text and figures are listed in the Locality Register.

Basal black shale division of Geneseo Formation

In the Genesee and Wyoming valleys the basal 0.5-1.5 meters of the Geneseo Formation is typically hard, black, and well jointed (Fig. 6). In the Genesee Valley, several thin rusty bands in this interval may be altered volcanic ash (K-bentonitic) layers, but this is, as yet, unproven. From Linden westward, this division appears to be absent due to non-deposition prior to the Taghanic onlap (Figs. 4, 5).

Mid-Geneseo calcareous shale interval

Above the basal black shale division is a thicker interval of brown weathering, splintery to concoidal, calcareous shale that becomes more calcareous and harder towards the top of the interval (Figs. 4-6). In Genesee and Wyoming valley sections, this interval displays thicker beds near the top and is generally more calcareous overall than further west and typically holds up waterfalls in creek sections. Fossils in the interval are usually poorly preserved; key forms encountered include *Styliolina*, *Pterochaenia*, the large bivalve *Panenka*, orthoconic cephalopods, and the goniatite *Ponticeras perlatum.* The prominent limestone band with rare *P. perlatum* in about the middle of this interval in the Genesee Valley, earlier mistaken for the Lodi

Limestone (Kirchgasser, 1981), is the Genesee Limestone (Kirchgasser, Over and Brett, 1997), a unit that may be traceable to the limestone bands in the lower Geneseo in Genesee, Wyoming and Erie counties. The upper part of the calcareous shale unit is notable for pavements of a chonetid brachiopod, particularly in condensed western sections. The calcareous shale interval is capped by a regional diastemic surface that can be traced from Genesee Valley sections westward into eastern Erie County (see Figs. 4-6)). This division thins westward from approximately 8 meters in the Genesee Valley to 0.4 meter at Little Buffalo Creek (Loc. 13) in eastern Erie County. West of Linden, this interval gradually loses its calcareous aspect and becomes darker overall, suggesting more substrate dysoxia to the west.

Regional discontinuity in upper part of Geneseo Formation

Capping the calcareous shale interval from the Genesee Valley westward to Marilla, Erie County (Loc. 13), is a sharp contact that is roofed by a hard, well jointed, black shale unit (Figs. 4-6). Where this contact is well exposed, a thin, discontinuous lag of fish bones, fish scales, and conodonts can be found, suggesting that the contact is an erosion surface. The stratigraphic occurrence of this contact below the Lodi "A" and "B" beds, precludes this being part of that marker. We believe that this horizon may equate to the top-Fir Tree discontinuity within the upper part of the Geneseo Member in the central Finger Lakes region (see Baird et al., 1989; Kirchgasser et al., 1989), but this still has to be corroborated by examination of sections between Seneca Lake and the Genesee Valley as well as examination of conodonts collected from this lag surface. West of Little Buffalo Creek (Loc. 13) this contact could not be located, though a thorough search of the Buffalo and Cazenovia Creek sections was conducted.

Top-Geneseo black shale division

Overlying the upper Geneseo regional diastem is an interval of well jointed, black shale and higher interlayered black and dark gray shale deposits (Figs. 4-6). This is a division that separates the aforementioned diastem from the overlying Lodi "A" and "B" layers. If the underlying diastem is eastwardly correlative with the top-Fir Tree discontinuity, than this shale unit is correlative with the Hubbard Quarry Shale (Baird et al., 1989; Kirchgasser et al.,1989).

Lodi "A" and "B" calcareous basal divisions of the Penn Yan Formation

From the vicinity of Geneseo west to Linden at Little Tonowanda Creek (Loc. 22), two thin beds of calcareous, gray shale with small concretions in the lower bed mark the base of the Penn Yan Formation (Figs. 4-6). Baird et al. (1989) and Kirchgasser et al. (1989) mapped and delimited the Lodi Member and a coeval discontinuity within the Finger Lakes region and established a link between the Lodi in its type area at Seneca Lake and sections at and west of the Genesee Valley. Two beds are recognized in this latter area that correspond to the Lodi; a lower "A" bed of bioturbated, gray shale and irregular, calcareous nodules yielding numerous auloporid corals, small brachiopods, gastropods, the goniatite *Ponticeras perlatum* and conodonts of the late Givetian *norrisi* chronozone, and an upper "B" bed which is characterized by calcareous, bioturbated shale rich in hashy fine grained fossil debris. Both layers, separated by a thin intervening black shale unit, can be traced westward to Linden (Figs. 5-6). However, west of Linden this level is covered in sections all the way to Durkee Creek (Loc. 15) near Alden. At and west of Durkee Creek, the fossiliferous aspect of the layers is lost and the interval is a cryptic, thin unit of dark gray shale yielding *Chondrites* and *Planolites* trace fossils (Figs. 4,5).

The westward darkening of the unit accords with the darkening of the underlying mid-Geneseo calcareous interval, suggesting more basinal conditions in that direction.

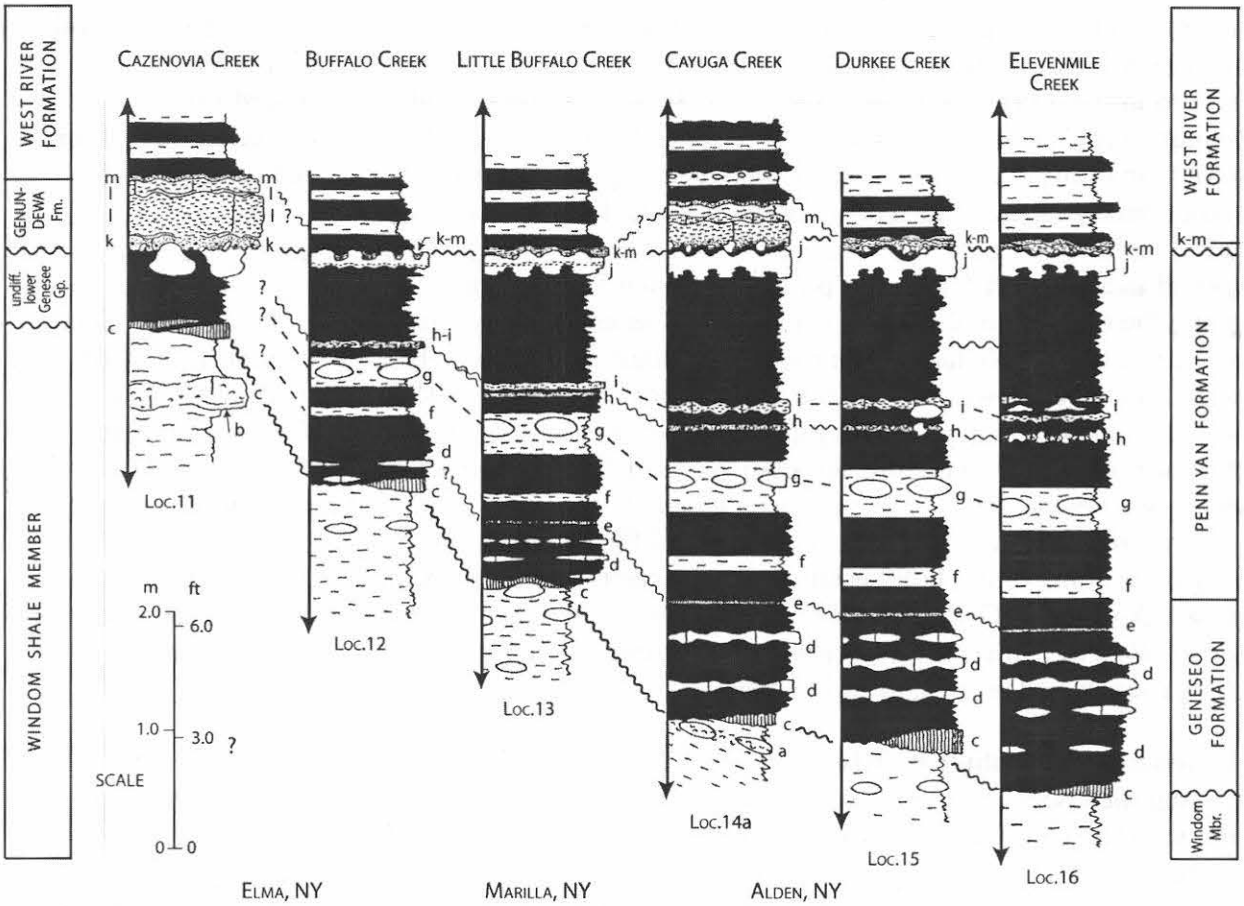

Figure 4. Lower and medial divisions of the Genesee Group; transect from Elma area, Erie County to the meridian of Darien Center. Note westward thinning of units by stratigraphic condensation to Buffalo Creek followed by accelerated westward erosional overstep of lower Genesee Group divisions by the sub-North Evans disconformity. Numbered sections are listed in Locality Register (see text).Lettered units are: a, *Pseudoatrypa – Athyris* Bed of medial Windom Member; b, Amsdell Bed of upper-middle Windom Member; c, Leicester Pyrite; d, calcareous, resistant, black-brown shale division of Geneseo Formation; e, discontinuity horizon yielding conodonts and fish bones. This may correlated eastward to the top-Fir Tree Limestone discontinuity in the central Finger Lakes region (see text); f, very dark gray, fissile shale unit probably correlative with the Lodi Limestone; g, band of large discoidal concretions in dark gray shale interval; h, discontinuity horizon yielding conodonts, fish bones, and reworked concretions. This correlates eastward to the Linden Bed (see text); i, eastwardly splaying bundle of styliolinid and micritic limestone beds and lentils herein informally designated the *Elevenmile Creek Beds*;j, zone of large, irregular, sculptured concretions herein informally designated the *grotesque concretion layer*; k, North Evans lag horizon; l, Genundewa Formation upper division ledge of styliolinid limestone; m, thin styliolinid-rich, hashy limestone bed yielding numerous conodonts and glauconite grains. This may correlate to the "Huddle Bed" horizon in the Genesee Valley (see text).

Post-Lodi black shale division

Between the Lodi "B" bed and a higher gray shale division is a black shale interval that thickens westward from the Genesee Valley into eastern Erie County. This thickening, running counter to westward overall thinning of lower Genesee units, is herein believed to reflect westward, waltherian, gray-to-black, facies change in this part of the section. In the Honeoye Lake and Canandaigua Lake region, the post-Lodi black shale division (Black Shale A) yields early forms of the conodont *Ancyrodella rotundiloba*, the taxon that marks conodont zone MN 1 and the base of the Upper Devonian Frasnian Stage (Kirchgasser et al. 1989; Kirchgasser, 1994).

Mid-Penn Yan gray shale division, concretion band, and Schumacher Bed

In the lower middle to medial Penn Yan succession, from the vicinity of Geneseo westward to Buffalo Creek in Erie County, is a persistant band of 0.3-0.5 meter diameter discoidal concretions with an associated interval of gray to dark gray, less resistant shale (Figs. 4-6). This shale unit, which largely underlies the concretion band, both thickens and becomes lighter in color eastward. It thickens from 21 cm at Buffalo Creek (Loc. 12) in Erie County, to 70 cm at Linden (Loc. 22) in the Little Tonowanda Creek Valley, and to 4.1 meters in the Genesee Valley (Figs. 4-6). From the Genesee Valley west to Linden, the top boundary of the gray shale layer with a succeeding black shale unit is marked by a styliolinid-rich and locally conodont-rich lag zone informally termed the "Schumacher Bed" (SB Bed) for a section (Loc. 26) in the Wyoming Valley (Fig. 6). In the Genesee Valley, the SB Bed yields conodonts of the early Frasnian MN Zone 1 (Kirchgasser, 1994). West of Linden no lag zone is seen at this level. From Alexander westward this contact is gradational.

Sub-Linden Bed black shale unit of Penn Yan Formation

Above the Schumacher Bed, or gray shale unit where it is absent, is a black shale interval that extends east from Erie County into the Wyoming Valley. This unit thickens eastward from 10 cm at Buffalo Creek (Loc. 12) to 85 cm at near Suicide Corners (Loc. 24) in Genesee County (Figs 4-6). East of Suicide Corners this black shale largely grades into gray shale facies in the Genesee Valley. Between the Genesee Valley and Linden this black unit grades vertically upward into a gray shale unit characterized by irregular concretions developed around a network of burrows (Figs. 5, 6). West of Linden, no gray shale or concretionary horizon can be found at this level; from Alexander (Loc. 19) west to Buffalo Creek (Loc. 12) black shale facies extends upward to the sharp base of the Linden Bed of nodules or its coeval discontinuity (Figs. 4, 5). The rapid westward loss of the gray shale unit and irregular concretions west of Linden cannot be fully explained at present. However, several units notably darken westward across this meridian; we postulate that this may be somehow linked to early activity along the Clarendon-Linden Fault Zone which runs north-south along the valley of Little Tonowanda Creek in this area.

Linden Bed and coeval diastemic contact

Extending from Little Tonowanda Creek north of Linden (Loc. 22) westward to Murder Creek southeast of Darien (Loc. 17) is a conspicuous band of very irregular, often turnip-shaped concretions that contain complex internal laminae of concentrated *Styliolina* shells and abundant phragmocones of the goniatite *Koenenites styliophilus* (Figs. 5-7). This unit termed the Linden

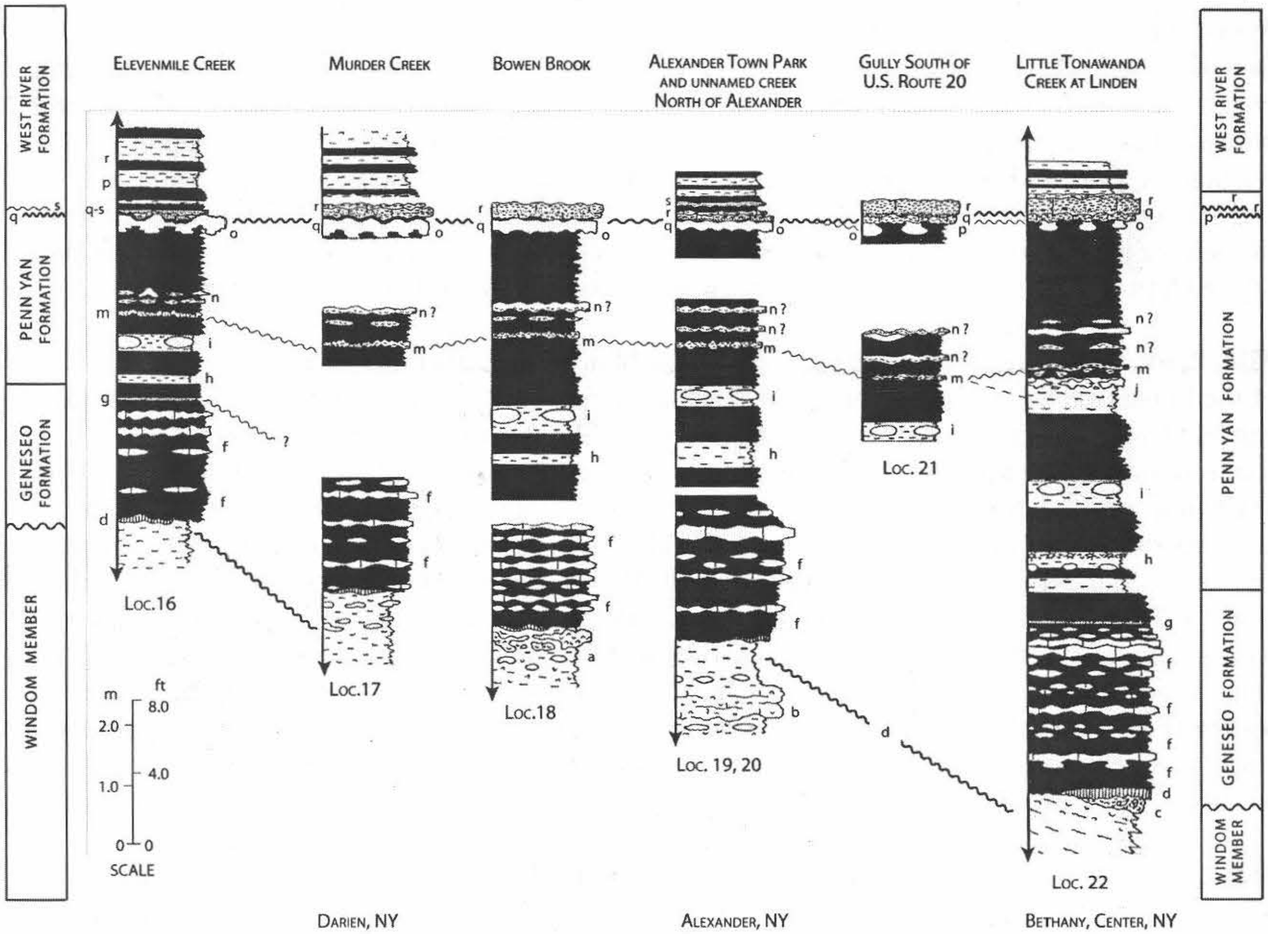

Figure 5. Lower and medial divisions of the Genesee Group; transect from the Darien Center meridian to Linden, New York. Note continued pattern of eastward expansion (splaying) of units. Numbered sections are listed in Locality Register (see text). Lettered units include: a, Bayview shell-coral Bed of Windom Member; b, Smoke Creek Bed of Windom Member; c, Fall Brook Coral Bed of Windom Member; d, Leicester Pyrite; e, Hard, black, well jointed basal shale division of Geneseo Formation; f, calcareous, resistant, black-brown shale division of Geneseo Formation; g, discontinuity horizon yielding conodonts and fish bones that possibly correlates eastward to the top-Fir Tree Limestone discontinuity (see text); h, interval eastwardly correlative to the Lodi Limestone (see text); i, band of large discoidal concretions and overlying "*Schumacher bed*" horizon of styliolinid debris; j, band of irregular concretions in gray shale not yet correlateable to sections west of Little Tonowanda Creek; m, Linden Bed of irregular *Koenenites* – bearing concretions and coeval discontinuity surface (see text); n, eastwardly splaying bundle of styliolinid and micritic limestone beds herein designated *"Elevenmile Creek beds"*; o, zone of irregular, sculptured concretions herein designated the "*grotesque concretion layer*" ; p, lower division of Genundewa Formation; q, North Evans lag horizon; r, upper division, styliolinid limestone ledge of upper division of the Genundewa Formation; s, thin, conodont- and glauconite-rich layer probably corresponding to "Huddle Bed" in Genesee Valley sections (see text).

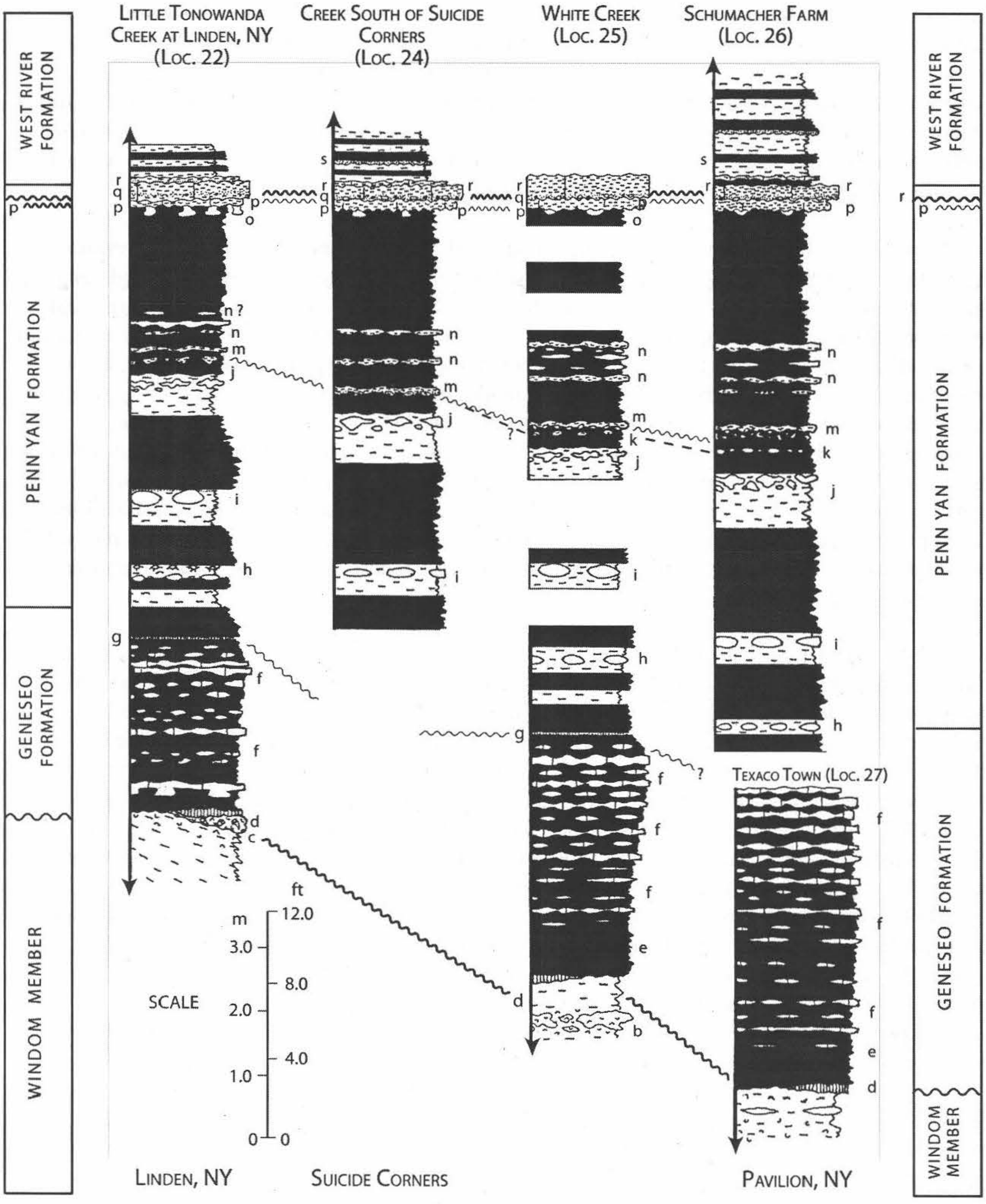

Figure 6. Lower and medial divisions of the Genesee Group along transect from Linden eastward to the vicinity of Pavilion. Note continued eastward thickening of units (see Figure 5 for key to lettered units).

Goniatite Horizon (Kirchgasser and House, 1981) and Linden Horizon (Kirchgasser, 1985) is a thin example of Cephalopodenkalk or "cephalopod limestone" long recognized by European workers (Tucker, 1974; Tucker and Kendall, 1973). It is a precursor to the Genundewa Limestone, a major styliolinid carbonate layer which is a key marker unit in the Genesee Group of western New York. Conodonts from the Linden Bed include the late form of *Ancyrodella rotundiloba,* indicating MN Zone 2 (Kirchgasser et al., 1989; Kirchgasser, 1994). New work on this unit by Baird shows that it correlates westward to a discontinuity surface (Figs. 4-7).

At Murder Creek the Linden Bed layer of cephalopod-rich limestone marks the westernmost limit of the Linden Bed (Figs. 5, 7D). At this section laterally disconnected lentils and irregular limestone masses, profusely packed with *Koenenites* phragmocones, hashy fossil debris, and secondary pyrite, show abrupt lateral margins coeval to a black shale roofed and floored diastem contact (Fig. 7D). Between the limestone masses the discontinuity lag consists of flattened *Styliolina* in association with fish bones and conodonts. The top, side, and basal margins of the limestone masses are irregular and pitted. Moreover, molds in the shape of *Koenenites* phragmocones within the limestone masses that are connected to the exterior are filled with black shale (Fig. 7D). This indicates that the Linden Bed at this locality is a residual corrosional remnant of a more nearly continuous layer. Dissolution of this limestone associated with the development of a submarine erosion surface within the basin has produced limestone remnants or "corrodules". The insoluble product of this process is represented by the bone-conodont lag concentrate (Fig. 7D).

West of Murder Creek, the cephalopod – bearing corrodules are entirely absent at this horizon, and a black shale-on-black shale erosional contact is developed west to Little Buffalo Creek (Loc. 13) northwest of Marilla (Figs. 4, 7). A thin 1-6 mm-thick lag blanket composed of flattened *Styliolina* in association of bone-conodont debris marks this disconformity at Elevenmile Creek (Loc. 16) and Durkee Creek (Loc. 15; see Fig.4). Reworked concretions typically displaying pyrite suffused corroded exteriors also occur at this level in these same sections (Fig. 7C). At Cayuga Creek (Loc. 14a), the tributary to Cayuga Creek (Loc. 14b; STOP 3), and Little Buffalo Creek (Loc. 13), reworked concretions are largely absent and the contact is marked by a subtle bedding plane reentrant within a fissile black shale interval (Figs. 4, 7B). At Buffalo Creek (Loc. 12) minor erosion beneath a higher styliolinid limestone layer has resulted in the juxtaposition of the limestone bed onto the Linden Bed discontinuity (Figs. 4, 7A). The base of this 2-3 cm-thick limestone layer is marked by abundant fish bones, fish scales, and conodonts (Fig. 7A). West of Buffalo Creek the Linden Bed discontinuity horizon is overstepped by the younger top-Penn Yan disconformity and it is absent at Cazanovia Creek (Loc. 11), the next section to the west of Buffalo Creek (Fig. 4).

East of Linden the correlations of this unit are more uncertain. Near Suicide Corners (Loc. 24), the cephalodod-rich, turnip-shaped Linden Bed nodules seen at Linden appear to be missing, though they may be reworked into an overlying *Styliolina* – rich limestone bed (Fig. 6). At White Creek (Loc. 25), the next section to the east, abundant concretions can be seen to be reworked into the aforementioned overlying styliolinid-rich layer, but smooth, *in situ* concretions rich in *Koenenites* can be seen immediately below this contact (Figs. 6, 7F). At a creek on the Schumacher Farm southwest of Pavilion (Loc. 26), the Linden Bed seems to have begun its eastward expansion with increased vertical differentiation of component beds and nodule layers (Fig. 6).

Black shale unit above Linden Bed

Between the Linden Bed and an overlying bundle of styliolinid-rich and micritic limestone layers, herein informally designated the "*Elevenmile Creek Beds*" division, is a fissile black shale interval marked by abundant *Pterochaenia* and flattened *Styliolina* at its lower contact with the Linden Bed or its coeval diastem. This division thins westward from 60 cm at Schumacher and White creeks, 23 cm at Alexander, 11 cm at Cayuga Creek tributary (STOP 3), and 2 cm at Little Buffalo Creek (Figs. 4-7). At Buffalo Creek (Loc. 12), the next section west of Little Buffalo Creek (Loc. 13), this black shale unit is missing due to overstep by a condensed limestone unit representing the local expression of the overlying Elevenmile Creek Beds division (Figs. 4, 7A,B).

"Elevenmile Creek Beds"

From Buffalo Creek east into the Canandaigua Valley is an eastwardly thickening bundle of micritic and styliolinid-rich limestone beds and lentils, herein informally designated *"Elevenmile Creek Beds,"* which markedly splay within the upper Penn Yan clastic wedge of the western Finger Lakes region. Our concern here is with correlations within this interval from the Wyoming Valley westward. Sections through this interval in the Genesee Valley and further east are thick and complex necessitating future correlation work in that area. At Buffalo Creek (Loc. 12), this division is represented by a condensed 2-3 cm-thick styliolinid-rich limestone bed that is erosionally juxtaposed onto the Linden Bed discontinuity (Fig. 7A). Proceeding east from there this bed differentiates into a complex of closely-spaced beds and lentils in the Alden-Darien area (Figs. 4, 5). Continuing east, these layers splay into more widely-spaced tabular ledges of variably stylioline-rich micritic limestone in the Little Tonowanda and Wyoming valleys (Figs. 5, 6). As such, this division thickens eastward from 2-3 cm at Buffalo Creek, 12 cm at Elevenmile creek (Loc. 16), 30 cm at Alexander (Loc. 19), and 78 cm at Schumacher Creek (Loc. 26) in the Wyoming Valley (Figs. 4-6). The interval appears to encompass several meters thickness in the Genesee Valley but this has not yet been confirmed. In the section at Linden these styliolinid bands yield shells of the small discoidal goniatite *Acanthoclymenia,* the lowest occurrence of this genus in the region.

Upper Penn Yan black shale division

Above the Elevenmile Creek Beds division is an interval of very black, fissile shale that extends nearly up to the top-Penn Yan disconformity marked by the North Evans Limestone or the lower division of the Genundewa Limestone (Figs. 4-6). Included within the upper part of this black shale unit is a continuous to discontinuous band of large concretions herein informally termed the *"Grotesque Concretion Layer."* Because these concretions appear to be linked to the development of the top-Penn Yan disconformity, they are here treated as a separate division. The black shale division, inclusive of the Grotesque Concretion Layer, thickens eastward from 66 cm at Buffalo Creek, 103 cm at Elevenmile Creek, to 2.0-2.2 meters in the Linden-Pavilion area (Figs. 4-6). Near the top of this interval, within the zone of the Grotesque Concretions in Erie County sections, is a bed of concentrated *Styliolina* that appears to have had some control on the pattern of concretion growth.

Top-Penn Yan layer of Grotesque Concretions

Just beneath the North Evans lag deposit and coeval Lower Genundewa Formation division is a band of large bulbous to complexly tiered, continuous to discontinuous concretions that occupy

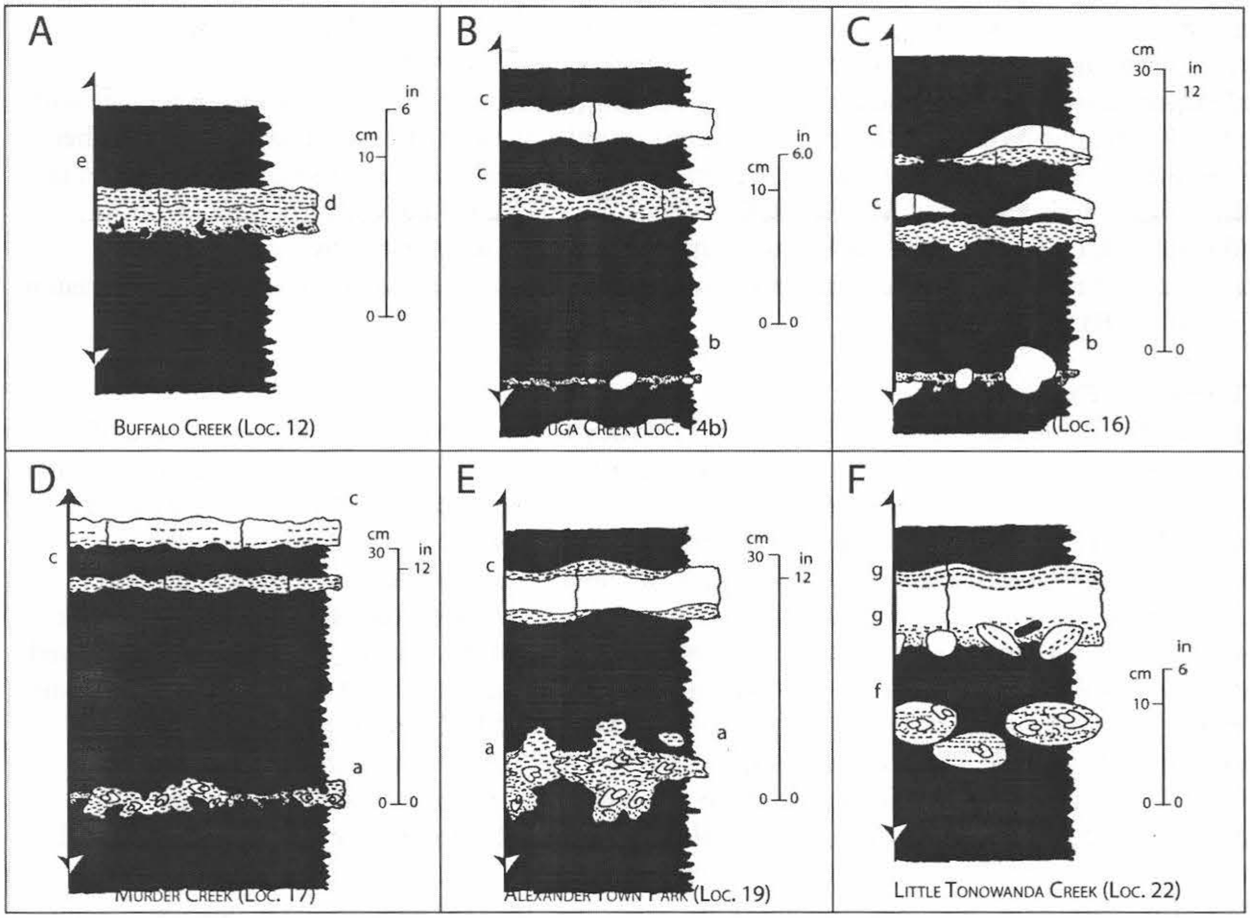

Figure 7. Pattern of westward condensation and corrosional destruction of the Linden Bed of irregular, *Styliolina* – and *Koenenites* – rich limestone nodules to form a black shale-on-black shale discontinuity. Sections D-E show westward condensation and corrosion of the Linden Bed; sections A-C show the discontinuity coeval to the Linden Bed; section F shows a problematic section east of the Linden Bed type section (see discussion in text). Numbered localities are listed in Locality Register (see text). Lettered units include: a, Linden Bed of irregular cephalopod-bearing nodules. This unit has undergone severe corrosion at Murder Creek (section D) to produce laterally separated, etched "corrodules"on a discontinuity surface; b, discontinuity coeval to Linden Bed. This is locally characterized by abundant reworked concretions; c, Elevenmile Creek Bed (bundle of styliolinid and micritic limestone beds); d, conodont-bone rich lag on compound discontinuity where overlying Elevenmile Creek Bed is juxtaposed onto the Linden Bed-equivalent discontinuity surface; e, ultracondensed Elevenmile Creek Bed which is erosionally juxtaposed onto discontinuity coeval to Linden Bed; f, *Koenenites* –bearing concretions possibly equivalent to the Linden Bed; g, Styliolinid limestone unit with lag deposit at base which also may be equivalent to the Linden Bed (see text).

the topmost part of the upper Penn Yan Shale division (Figs. 4-6). These black concretions are locally in contact with the post-Penn Yan carbonate units, but often are separated from the North Evans and Genundewa by a thin parting of black shale. The concretions are most prominently displayed and continuous in Erie County sections; they become thinner and more discontinuous eastward across Genesee County (Figs. 4-6). Multiple stages of concretion growth over an extended time period are indicated by growth partings that are sometimes layered or tiered. The complex growth architecture of the concretions gives rise to the informal name *"Grotesque Concretion Layer"* given to this unit by the present authors. One concretion showed discrete interior growth bands when fragmented. Patterns of complex growth and overgrowth within concretions were influenced locally by the presence of one or more thin styliolinid limestone

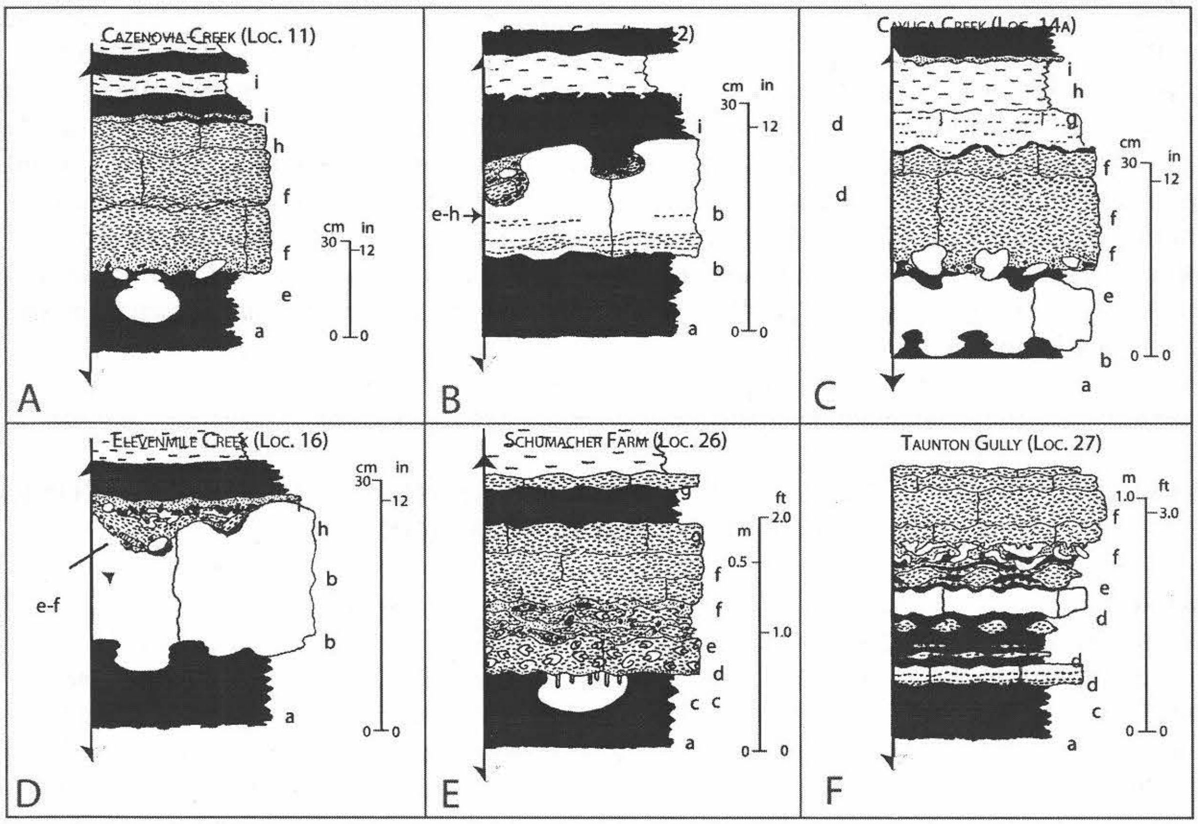

Figure 8. Genundewa Formation and associated units at several key sections (see discussion in text). Note extremely thin, lag-dominated North Evans/Genundewa deposits at Buffalo and Elevenmile creeks. Numbered localities are listed in Locality Register. Lettered units include: a, black shale division of topmost Penn Yan Formation; b, grotesque concretion layer; c, discontinuity at base of lower limestone division of Genundewa Formation; d, lower limestone division of Genundewa Formation. Note abundant goniatites in this unit at Schumacher Farm; e, North Evans lag deposit; f, upper limestone division of Genundewa Formation; g, unnamed basal division of the West River Formation; h, thin styliolinid carbonate layer in the basal West River interval which is rich in conodonts and glauconitic grains. This may correlate to the "Huddle Bed" in Genesee Valley sections (see text); i, succession of alternating black and gray shale beds within the West River Formation.

beds within the topmost part of the black shale interval in Erie County. At Buffalo Creek (Loc. 12), there is evidence that the concretion band was distinctly eroded or corroded prior to final accumulation of the North Evans lag deposit; the top of the concretion band displays pits filled with lag debris (Fig. 8B). Small clay concretions having a top-, or mason jar-shape, are abundantly reworked into the North Evans from Cazenovia Creek (Loc. 11) eastward to Elevenmile Creek (Loc. 16), suggesting that concretions were forming continuously within the topmost Penn Yan black shale interval in response to the downward migration of the disconformity surface; once formed the nodules would have been eroded out to form part of the North Evans placer. From Murder Creek (Loc. 17) eastward to the Genesee Valley few concretions are seen to be reworked. However, some erosion and exposure of preformed concretions in this zone is evidenced by the presence of open borings of the ichnogenus *Trypanites* into the top surfaces of *in situ* concretions which are in direct contact with the overlying Genundewa. The grotesque concretion horizon is, thus, viewed as having an *underbed* origin; its presence is integrally related to the development of the top-Penn Yan disconformity. Underbed limestones and concretion layers are believed to form through prolonged sulfate reduction and/or methanogenic activity along a stable redox boundary. Shifting of the North Evans debris blanket combined with prolonged exposure and erosion of variably dewatered Penn Yan organic – rich muds, would have been an ideal setting for such diagenesis. The irregular, undulating, base-North Evans erosion surface is a notable feature of Genundewa Formation sections in westernmost Genesee and eastern Erie counties. This "hillyness," which controls the thickness of overlying units (see Fig. 8 B,D), is believed to be due to differential cementation of the underlying Penn Yan top black shale division followed by differential compaction (dewatering) of uncemented mud both during and following the episode of grotesque concretion formation.

STRATIGRAPHY OF MEDIAL GENESEE DEPOSITS: NORTH EVANS LIMESTONE AND GENUNDEWA FORMATION

Overview

The North Evans/Genundewa succession was examined in detail from Pike Creek (Loc. 1) at Lake Erie eastward to the Wyoming Valley (Figs. 4-6, **8-9**) with the addition of a few selected sections from the Genesee Valley. The North Evans Limestone is a bone-conodont-rich lag that largely underlies styliolinid grainstone deposits of the upper Genundewa division in Erie and westernmost Genesee county sections, but which grades eastward, in part, into styliolinid – rich facies of the lower Genundewa division across Genesee and western Livingston counties. It marks the position of an internal discontinuity within the extremely condensed deposits of the Genundewa which overstep the lower Genundewa succession towards Buffalo. This discontinuity grades eastward into a condensed zone across Genesee County and closes to apparent continuity at the meridian of Geneseo. The North Evans rests on the top-Penn Yan regional disconformity and it conspicuously oversteps both the Geneseo and Penn Yan formations westward across Erie County (Figs. 4-6, 8).

The Genundewa Formation is composed of massive to nodular styliolinid limestone with distinctive fossils and sedimentary structures. The Genundewa contact with the overlying West River Formation is gradational where the section is expanded. However, an overlying subsidiary lag unit and associated diastem within the basal West River shale appears to overstep the North

Evans/Genundewa succession locally where local structure or bathymetric paleorelief produced extreme condensation and erosive telescoping of medial Genesee deposits (Figs. 4, 8, 9-14).

Lower Genundewa Formation division

In the Genesee Valley-Linden area, the overall Genundewa interval is thicker and more differentiated than it is further west. In the Wyoming and Little Tonowanda Creek valleys, the North Evans lag of conodonts and glauconite is not at the base of the Genundewa, but centered at the top of the lower third of the Genundewa succession (Fig 8E). The base of the Genundewa is probably disconformable to some extent with the underlying Penn Yan Shale in this area, as exemplified by a sharp basal contact and by *Trypanites* that extend downward into subjacent Penn Yan concretions (Fig 8E), but no major ("North Evans")-type, crinoidal-glauconitic, bone-

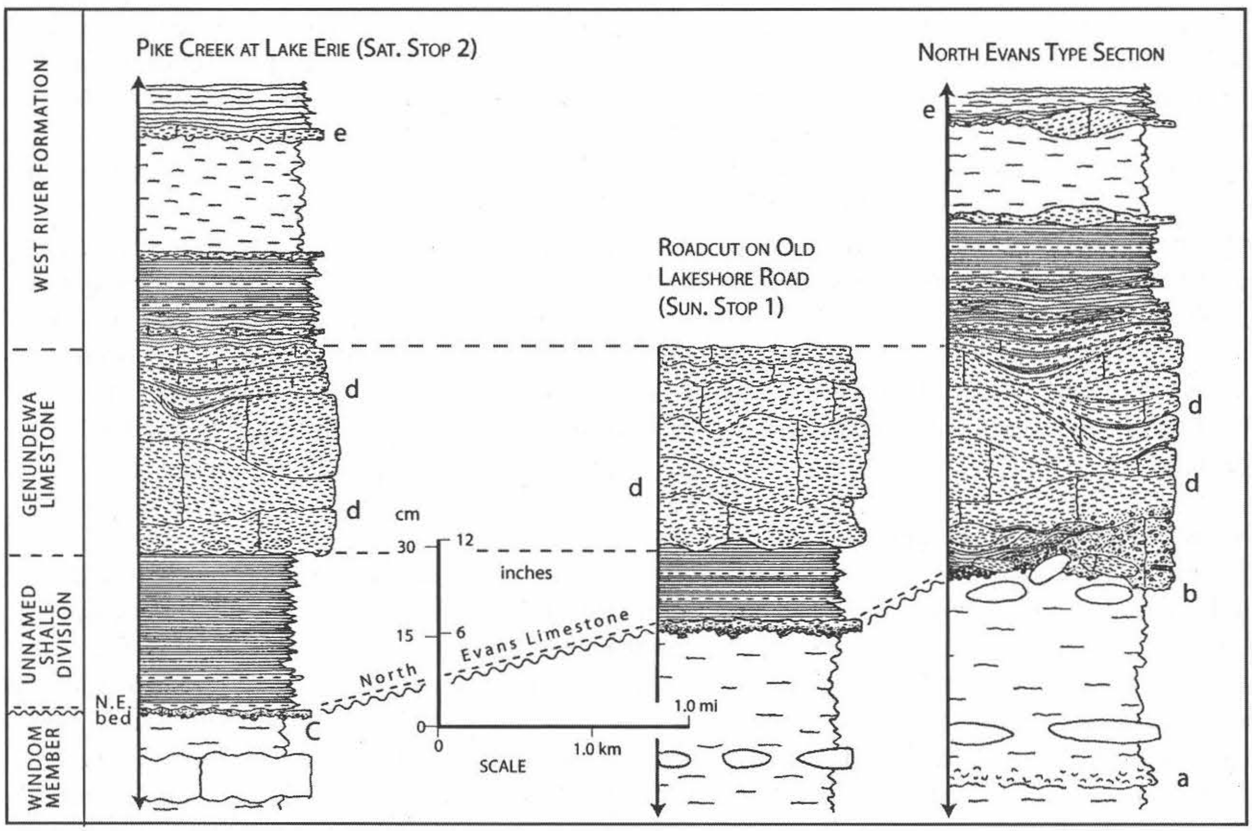

Figure 9. Genundewa Formation and associated units in sections both at and near Lake Erie in southwestern Erie County. Note westward thinning of the North Evans lag deposit from its type section on Eighteenmile Creek (Loc. 3) in association with the westward appearance of-, and thickening of, an overlying unnamed black shale unit. Note also the conspicuous bedding undulation and local channeling within the styliolinid grainstone facies of the Genundewa upper division limestone (see discussion in text). Lettered units include: a, Amsdell Bed of Windom Member; b, thick crinoidal subfacies of North Evans Limestone observed by the Amtrack railroad overpass on Eighteenmile Creek (North Evans type section); c, thin, bone, detrital pyrite – rich subfacies of North Evans Limestone seen below dark shale unit; d, massive ledge of Genundewa Formation upper division limestone; e, thin, lenticular, styliolinid limestone unit rich in conodonts and glauconitic grains. This bed may correlate eastward to the "Huddle Bed" (see discussion in text).

conodont-rich lag zone is seen at its base. Typically, the basal, 15-23 cm-thick bed of the Genundewa near Linden (Loc. 22), Bethany (Loc. 23; see STOP 4: Figs. 8E, 14), and Pavilion (Loc. 26) is a styliolinid packstone-grainstone unit containing abundant goniatites including *Koenenites styliophilus* (to be described as a new subspecies), *Acanthoclymenia* genundewa and *Tornoceras uniangulare compressum* (House and Kirchgasser, 1993; in press). Along with the Linden Bed of the Linden-Murder Creek region, this bed is a prime example of *cephalopodenkalk* of European workers. We will see a classic exposure of this Genundewa subfacies at Bethany Center (Loc. 23; see STOP 4; Fig. 14). The basal Genundewa cephalopod-rich layer grades upward into 5-15 cm-thick nodular zone rich in styliolinid packstone-grainstone lentils that are surrounded by hashy, debris-rich shale partings and burrow fillings (Figs. 8E, 10, 14). This nodular interval, rich in glauconite, crinoidal debris, conodonts, and occaisional fish fragments, is interpreted to be the North Evans lag zone in this area. The North Evans discontinuity regionally oversteps the lower Genundewa division west of the Geneseo meridian such that only the North Evans lag deposit and the overlying ledge of the Upper Genundewa Division are typically seen from Elevenmile Creek westward (Figs. 8, 9, 13). In actuality the westward terminus of the lower Genundewa division is, also in part, a gradual westward transition of Genundewa facies into North Evans facies through a complex, time-rich process of extreme sedimentary condensation (telescoping) of styliolinid beds interspersed with multiple erosion events (Figs. 10, 11). In the Elevenmile Creek-Cazenovia Creek region, reworked and quasi-reworked nodules and masses of styliolinid carbonate, complexly interspersed throughout the North Evans, serve as evidence for this. At Spezzano Ravine (Loc. 28) and Taunton Gully (Loc. 29) on the west side of the Genesee Valley (Fig. 8F) the North Evans zone of reworking has largely closed with only minor evidence of reworking evident. At Fall Brook (Loc. 30), south of Geneseo, no confirmed evidence of erosion and associated North Evans facies has been observed.

North Evans Limestone and Lag Deposit

The thin North Evans blanket of lag debris was once thought to intergrade laterally eastward to the Leicester Pyrite or overlying Penn Yan Formation (Rickard, 1975). Subsequent work by Baird and Brett, 1982 showed that the North Evans could be followed along the base of the Genundewa as far east as Elevenmile Creek (Loc. 16). North Evans debris is traceable *within* the Genundewa eastward through the Wyoming Valley and tentatively to Taunton Gully (Loc. 29) near Leicester (Figs. 6, 8E, F, 10, 11, 14). As such, the North Evans marks a regional discontinuity traceable for more than 100 kilometers. Moreover, the North Evans is now seen to overstep the Leicester, Geneseo, Penn Yan, and lower Genundewa divisions westward to produce a temporally huge hiatus (Figs. 1, 10).

As with the Leicester, the North Evans is very coarse; it contains reworked glauconite coated concretions, teeth, spines, and scales of sharks and bony fish, abundant pelmatozoan debris, and a rich and famous conodont component. In Erie and westernmost Genesee counties the North Evans contains concretions derived from underlying units; from North Evans northeast to the north branch of Smoke Creek (Locs. 3-10), concretions are reworked directly from the Windom Shale (Figs. 9, 12); from Cazanovia Creek (Loc. 11) eastward to Elevenmile Creek (Loc. 16), they are derived from lower Genesee Group strata (Figs. 4, 5, 8, 10, 12, 13). Reworked concretions are often coated by a thin sheen of glauconite imparting a green color to nodule surfaces, and many show development of diagenetic pyritic halos under their surfaces.

Concretions reworked from upper Penn Yan strata (see STOP 3), often display peripheral, medial reentrants around their exteriors much like the indentation around the mouth of a mason jar or the indentation along the periphery of a yo yo. No encrusters have been seen on concretions, but some concretions show borings of the ichnogenus *Trypanites*. The North Evans is famous for its fish fauna (see Hussakoff and Bryant, 1918; Turner, 1998), but the fauna as a whole has yet to see modern taxonomic study. Ptychtodont tritors (lozenge-shaped crushing teeth) and cuspate cladodont shark teeth are common and large armored fish plates can also be found. The North Evans Limestone ("conodont bed" of Hinde, 1879) contains conodonts representing possibly seven conodont zones, an interval of possibly over two million years (scale of Kaufmann, 2006) admixed into a complex lag blanket (Bryant, 1921; Huddle, 1981; Kirchgasser and Kozlowski, 1996; Kirchgasser and Vargo, 1998; Kirchgasser, 1994, 1996, 1998,

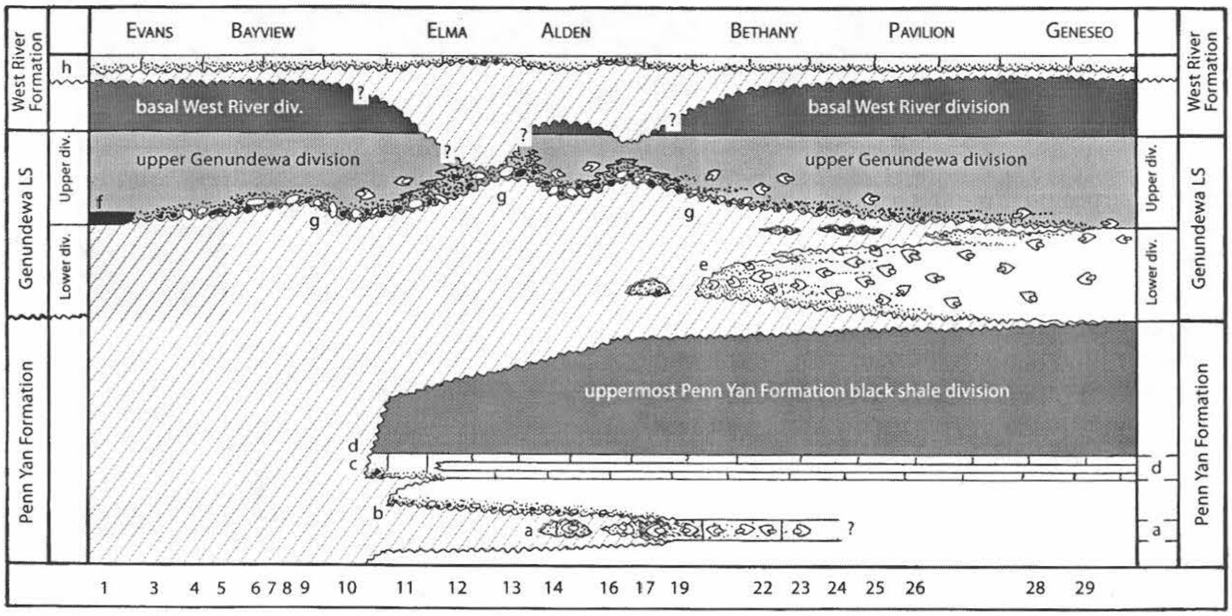

Figure 10. Inferred chronostratigraphy of medial Genesee Formation interval in western New York. Note the discovery of a discontinuity coeval to the Linden Bed, westward erosional overstep of the lower Genundewa Formation division west of the Genesee Valley, local maximal condensation of the upper Genundewa Formation division in a region of inferred basin upwarp or basin margin slope, and inferred partial-to-nearly complete truncation of the Genundewa Formation by a younger discontinuity associated with the Huddle Bed horizon in the basal West River Formation (see discussion in text). Lettered units include: a, Linden Bed; b, discontinuity coeval to Linden Bed; c, minor local diastem below westernmost condensed unit of Elevenmile Creek beds interval; d, Elevenmile Creek beds interval; e, thick North Evans lag deposit at North Evans type section that may be, in part, coeval to a localized Genundewa Formation dark shale unit ; f, local dark shale unit within upper division of Genundewa Formation near Lake Erie; g, North Evans lag deposit below Genundewa Formation upper division limestone unit; h, conodont- and glauconite-rich lag deposit in basal West River Shale that may be equivalent to the "Huddle Bed" in Genesee Valley sections. Numbers at base correspond to numbered localities listed in Locality Register.

2001, 2002, 2004). The taphonomic (final burial) age of the North Evans Limestone correlates to the upper part of early Frasnian MN Zone 2, indicated by *Ancyrodella recta,* the youngest conodont in the mix. The North Evans, as with most lag beds, poses a depositional paradox; even though the lag content records an enormous span of time, the actual final depositional event producing the bed may have been geologically instantaneous.

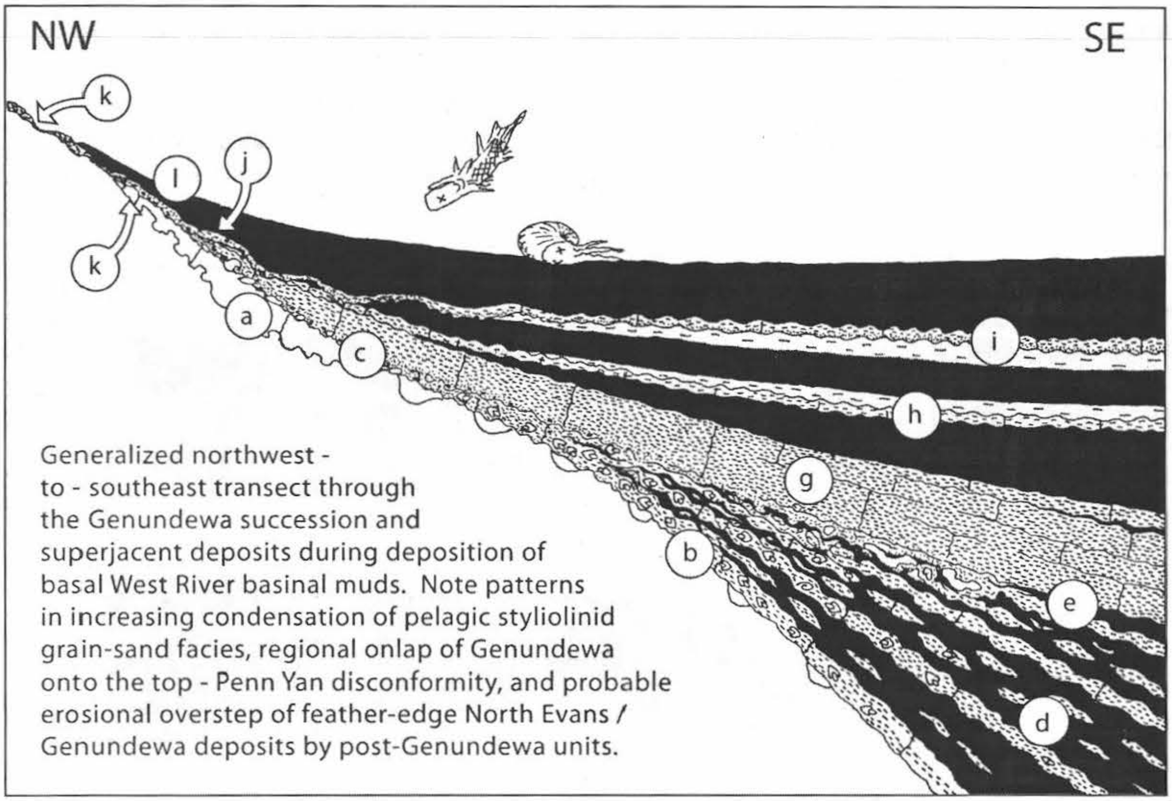

Figure 11. Generalized inferred transect of Genundewa Formation divisions normal to depositional strike, based on regional stratigraphic trends (see text). Paleoenvironment is that of basinal, near anoxia timed with accumulation of West River black muds following Huddle Bed lag accumulation. Lettered units include: a, grotesque concretion layer; b, discontinuity at base of Genundewa Formation lower limestone division; c, North Evans lag deposit; d, Genundewa Formation strata temporally equivalent to North Evans lag facies in Buffalo area sections; e, inferred basinward (downslope) limit of North Evans erosion and absence of North Evans lag debris; inferred upslope erosive overstep of grotesque concretion layer; g, *Styliolina* grainstone blanket corresponding to the Genundewa Formation upper division limestone; h, post-Genundewa styliolinid limestone bed which merges downward onto the Genundewa towards the inferred basin margin; i, Huddle Bed lag layer which converges downward into condensed Genundewa Formation facies and eventually oversteps it (see j) north of the present-day Genundewa outcrop margin; k, "Leicester –type detrital pyrite lag accumulating on the lower West River-age basin margin erosional slope surface (north of the present-day Genesee Group outcrop margin); l, inferred northwestward onlap of West River basinal mud over the erosion surface and burial of pyritic lag material (see discussion in text).

The important difference between the North Evans and Leicester is the dominantly carbonate nature of the former and the overwhelmingly insoluble character of the latter. We believe that the North Evans lag accumulated under conditions that were less dysoxic and, by implications, shallower than those applying to the Leicester. In essence, the North Evans lag is what the Leicester may have looked like at an upslope position on the Taghanic sediment-starved, erosional ramp prior to its subsequent dissolution at greater depth (Brett and Baird, 1982; Baird and Brett, 1986). At STOP 1, we will see a variant of the North Evans that more closely resembles the Leicester owing to the presence of a localized black shale unit that closely overlies it at the southwesternmost limit of its outcrop by Lake Erie (Fig. 9). In this area, North Evans lag debris was exposed to more severe dysoxic conditions than elsewhere; it is distinctly more pyrite-rich at its base and is greatly reduced in volume.

Upper (post-North Evans) division of Genundewa Formation

The Genundewa Formation upper division is usually expressed as a 18-40 cm-thick, waterfalls-capping ledge of dark gray, styliolinid grainstone-packstone carbonate. It is regionally widespread (Figs. 1, 4-6, 9-10, 12-14), but is nearly absent or greatly thinned at four localities in eastern Erie County and westernmost Genesee County (see Figs. 4, 5, 8, 10). The Genundewa upper division ledge roughly corresponds to the original *"Pteropod Limestone"* of Grabau (1898-1899) and other early workers who studied this unit in Erie County and who ascribed its component enigmatic microfossil to a group of planktonic marine mollusks known as pteropods. *Styliolina fissurella* is a problematic, 1-2 mm-long, calcareous conical shelled organism of uncertain affinities. It was originally described erroneously from flattened material, hence the specific name "*fissurella*" (Hall, 1843, 1879). Subsequent workers placed these organisms in a variety of groups; initially pteropod mollusks, later tentaculitids, and most recently protista (see Lindemann and Yochelson, 1994; Lindemann, 2002). The astonishing abundance of this taxon in the Genundewa Formation overall constitutes a major regional bioevent, or *epibole*; this organism appears to have been a form of extinct plankton that must have undergone periodic "blooms" in the epicontinental sea. In the Genundewa, these shells are uncompressed and are sometimes replaced or casted by pyrite. Although this unit is volumetrically almost entirely composed of *Styliolina,* other fossils, including the diminutive bivalve *Pterochaenia* and rare goniatites, including *Acanthoclymenia* and *Manticoceras,* first occurrence of this key genus in New York, can be found at its top. Crinoid ossicles and wood debris are also locally common in this unit. The conodont fauna of the upper Genundewa division includes *Ancyrodella rugosa* indicating MN Zone 3 (Kralick, 1994). The biota is of low diversity and suggests a dysoxic stressed environment, particularly, when compared to the rich, high diversity benthic fauna of carbonate units in the underlying Hamilton Group. Devonian styliolinid limestone facies is also known from European and North African sections where it is understood to represent condensed pelagic facies which accumulated in sediment-starved settings on the order of tens to hundreds of meters of water depth (see Tucker and Kendall, 1973; Tucker, 1974; Bandel, 1974). The Genundewa Formation overall compares most closely to the "cephalopodenkalk" (cephalopod limestone) facies of the German Rhenohercynian region; this carbonate accumulated on structural "highs" (schwellen) where styliolines, goniatites, diminutive bivalves, and ostracodes accumulated in a sediment-starved regime (Tucker, 1974). Basins between these swells received contemporaneous accumulation of thick shale units and turbiditic facies that yield mainly ostracodes and little else. Compared to descriptions of the Rhenohercynian cephalopodenkalk, the Genundewa notably lacks micrite and is much more nearly a styliolinid grainstone.

However, it is locally packed with goniatite phragmocones in a manner typical of many cephalopodenkalk units (see STOP 4; Fig. 14).

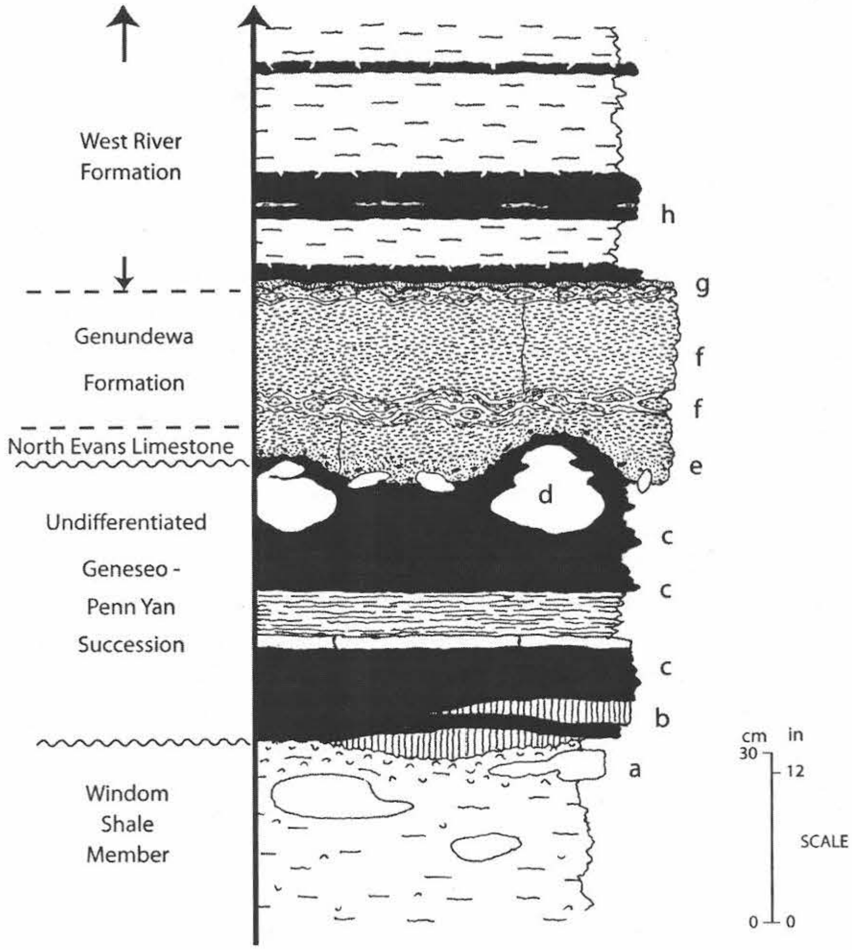

Figure 12. Genundewa Formation and associated units at Cazenovia Creek (Loc. 11; see STOP 2). Lettered units include: a, gray Windom Shale (within the Amsdell Bed interval) yielding abundant specimens of the small brachiopod *Emanuella praeumbonata* ; b, Leicester Pyrite; c, a severely truncated lower Genesee Group shale section composed of undifferentiated dark gray and black shale facies; d, large multilayered concretions corresponding to the "grotesque concretion layer"; e, pelmatozoan-rich, conodont, bone, reworked concretion and glauconite-bearing lag deposit corresponding to the North Evans lag deposit; f, upper Genundewa Formation styliolinid limestone division ledge; g, 1.0 cm-thick lag bed rich in conodonts and glauconitic grains which may correlate to the Huddle Bed in Genesee Valley sections; h, black and gray shale bed succession of the West River Formation.

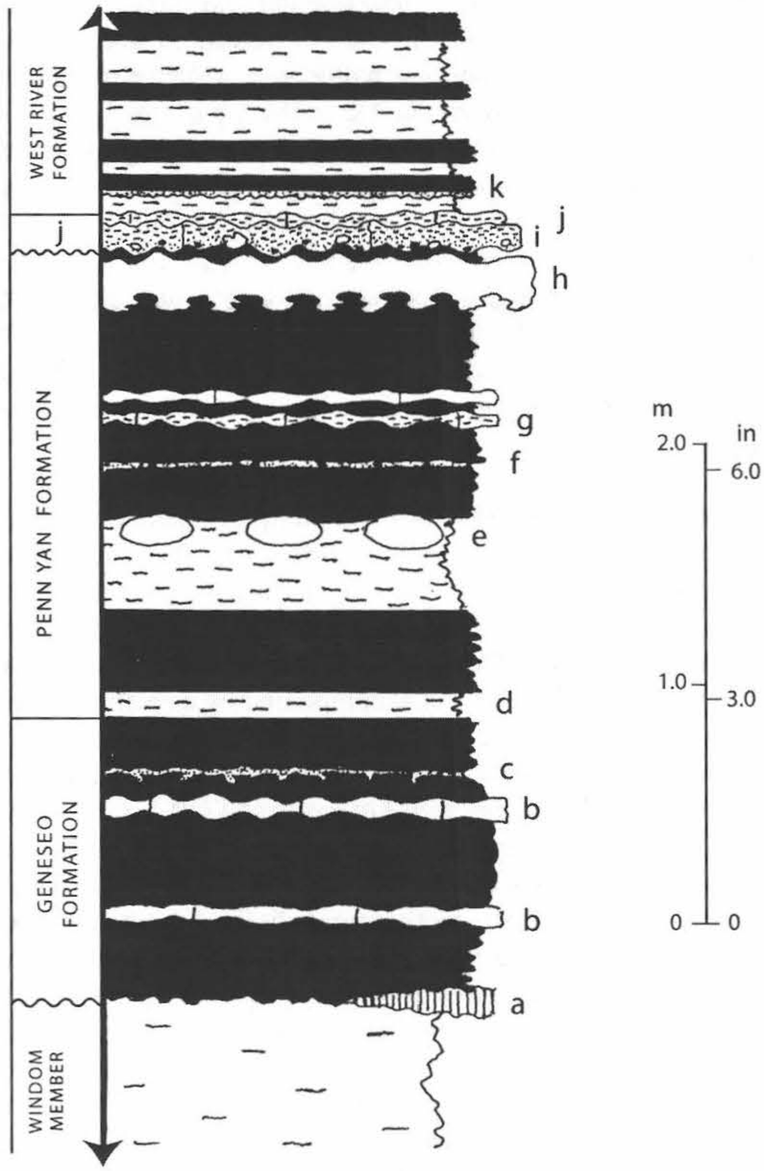

Figure 13. Lower and medial Genesee Group succession exposed on west flowing tributary of Cayuga Creek (Loc. 14b; see STOP 3). Lettered units include: a, Leicester Pyrite; b, calcareous, resistant black and dark gray shale interval of medial Geneseo Formation; c, thin styliolinid-, conodont-, and bone-rich debris layer probably correlative with the top-Fir Tree discontinuity (see text); d, dark gray shale unit probably equivalent to the Lodi Limestone interval (see text); e, layer of large discoidal concretions within dark gray shale; f, styliolinid-, conodont- and bone-rich debris layer which correlates to the Linden Bed; g, Elevenmile Creek Beds interval; h, grotesque concretion layer; i, North Evans lag deposit; j, resistant ledge of Genundewa Formation upper division styliolinid limestone bed; k, very thin styliolinid hashy limestone horizon which may correspond to Huddle Bed horizon in Genesee Valley sections (see text).

The upper Genundewa interval in Erie and Genesee counties is typically massive, but when weathered, the limestone typically splits apart into nodular and flaggy beds. Nodules occur as laterally linked to separate zones of sparry styliolinid limestone surrounded by muddy styliolinid partings. Bedding in this unit is usually laminar with some evidence of bioturbation. Goniatites are rather rare in the upper Genundewa division, but at the top of the unit at Eighteenmile Creek, small discoidal shells of *Acanthoclymenia* occur as replacements in pink barite. Preparation for both the Saturday and Sunday field trips led to discovery of distinctive cross stratification within the upper Genundewa division along the Lake Erie shore bluffs southwest of Pike Creek (Loc. 1; see STOP 2 of Saturday fieldtrip A4, this volume) and nearby on Eighteenmile Creek (Loc. 3; Fig. 9c). Several stacked sets of low angle cross stratified styliolinid grainstone can be seen with distinct thickening and thinning of beds in the cleaner, longer sections. Locally, beds are distinctly cut out where channelization has occurred. This pattern resembles small-scale hummocky cross-stratification suggesting the influence of deep-storm wave impingement at the substrate (Fig. 9a, c).

Along Eighteenmile Creek (Loc. 3) and in exposures along the Lake Erie shore (Loc. 1; see STOP 1 this fieldtrip, and STOP 2 of the Saturday A4 fieldtrip, this volume), a very localized shale unit is observed to separate the North Evans layer from the overlying ledge of the Genundewa upper division (Figs. 9, 10). This unnamed, 25 cm-thick, brownish black, flaggy shale unit contains flattened *Styliolina*, *Pterochaenia*, and wood debris. The westward appearance and thickening of this shale unit is accompanied by a reciprocal westward-southwestward thinning of the underlying North Evans Limestone (Figs. 9, 10). At the North Evans type section on Eighteenmile Creek (Loc. 3), adjacent to-, but immediately downstream from, the Amtrack railroad overpass, about four km east of STOP 1, the North Evans is a 6-12 cm thick lag accumulation dominantly composed of crinoidal debris, but also characterized by abundant conodonts, glauconitic grains, fish debris, and reworked concretions (Fig. 9c). At the overpass bridge, the North Evans grades directly into styliolinid grainstone facies of the overlying Genundewa, and no dark shale is present. However, further downstream from that bridge to the northwest, a succession of creek bank sections shows dramatic thinning of the North Evans with the appearance of the feather edge of the dark shale as a nodular parting above the lag unit. Moreover, the North Evans undergoes a lateral, spectral change from a thick crinoidal unit to a thin layer of mixed carbonate grains and detrital pyrite that is distinctly more "Leicester"- like (Fig. 9c). In the next key section (Loc. 2; Fig. 9b) along Lake Shore Road (STOP 1), the intervening shale is 10-11 cm-thick and the North Evans is only about 2-3 cm-thick. This transition from dominantly calcareous to largely insoluble grains, accords closely to the appearance of the overlying dark shale, and it suggests that carbonate dissolution of the exposed lag was more intense in the more basinal subenvironment of STOP 1 than at localities further to the northeast.

Given that this localized shale unit has a conformable (non erosional?) upper contact with the upper Genundewa division ledge with no evidence of an erosional lag deposit, we herein believe that this intervening shale unit is a local basinal subfacies of the Genundewa upper division (Fig. 10). If a significant bone-conodont lag unit were present at this horizon, we could argue for, (and test for), the idea that the local black-brown shale unit was part of the Lower Genundewa division or even the Penn Yan Formation. Evidence for an abrupt, but conformable, shale-limestone contact at STOP 1 and along the lakeshore argues against these other possibilities.

However, observation of reworked nodules of styliolinid grainstone and black shale clasts within the North Evans lag near the Amtrack railroad overpass on Eighteenmile Creek (Fig. 9c), offer the possibility that this material may be derived from the lower Genundewa division and/or underlying Penn Yan strata which are now absent due to erosional overstep.

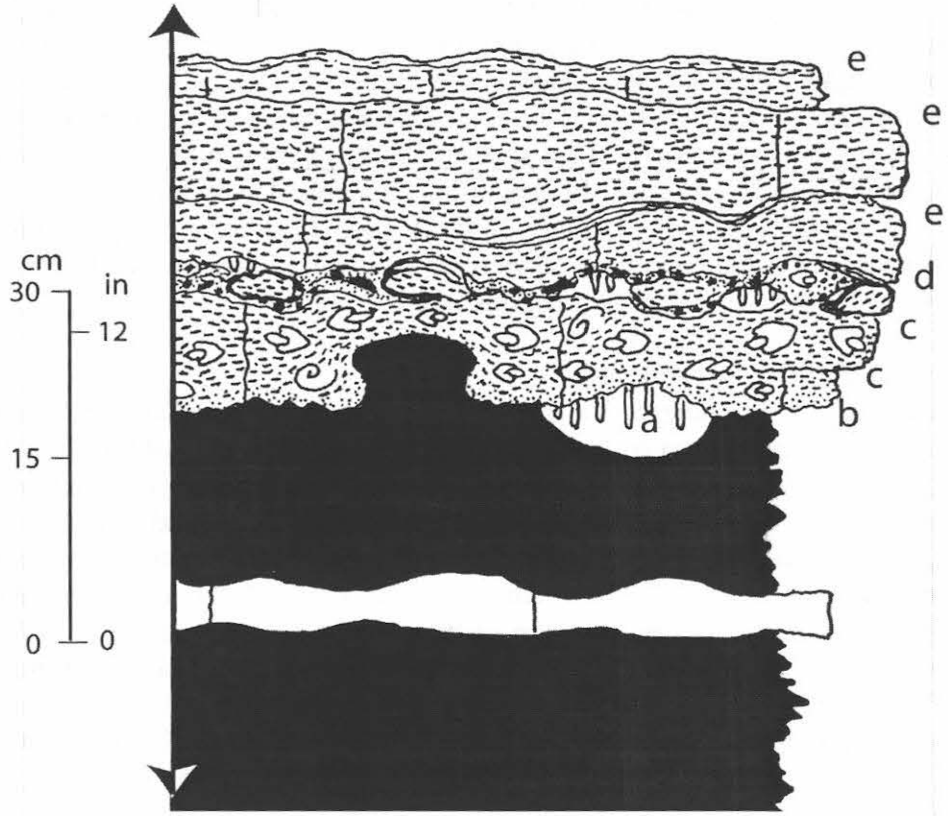

Figure 14. Genundewa Formation section in roadcut section along US 20 at Bethany, New York (Loc. 23; see STOP 4). The Genundewa is exposed below the Bethany Center Road overpass on both sides of US 20. Lettered units include: a, underbed concretion within uppermost Penn Yan Formation that corresponds to grotesque concretion layer further to the west; b, discontinuity flooring Genundewa Formation lower limestone division. Note *Trypanites* (borings) that extend downward into top-Penn Yan concretions from this contact; c, lower Genundewa limestone division showing development of classic cephalopod limestone (*cephalopodenkalk*) facies composed of styliolinid limestone with abundant phragmocones of the goniatite *Koenenites*. These can be collected easily from loose blocks on the US 20 highway berm east of this outcrop; d, North Evans interval of nodular styliolinid limestone, debris filled burrows, and thin hashy shale partings rich in conodonts and occasional fish fragments; e, resistant ledge of styliolinid grainstone – packstone facies corresponding to Genundewa Formation upper division limestone.

At two localities (Buffalo and Little Buffalo creeks) in eastern Erie County and two localities (Durkee and Elevenmile creeks) in western Genesee County (Figs. 4, 5, 8B, D), the Genundewa Limestone is very thin to nearly absent. Moreover, in these sections the Genundewa interval yields abundant pelmatozoan ossicles, fish debris, conodonts, reworked pyrite, reworked concretions, and it more closely resembles the North Evans lag facies at these places (Figs. 4, 8B, D). All four sections show a highly irregular top-Penn Yan disconformity surface directly capping, or nearly capping, the grotesque concretion layer. These depressions are filled with styliolinid carbonate complexly admixed with North Evans debris and reworked concretions whereas inter-depression "highs" are overlain by only a thin veneer of lag (Fig. 8B,D). At Buffalo Creek (Loc. 12), the most extreme condition is seen, in which North Evans-type debris-fills are observed in erosional pits excavated into the grotesque concretion bed; these pits are surrounded by inter-pit highs where the Penn Yan grotesque concretion band is directly overlain by West River black shale deposits (Fig. 8B). This is all the more notable, in that the next section to the west, Cazenovia Creek (Loc. 11), displays a thick upper Genundewa limestone succession (Figs. 4, 8A, 12). A clue potentially explaining the anomalously thin North Evans/Genundewa sections at Cazanovia , Buffalo, Little Buffalo, Durkee, and Elevenmile Creeks, is the development of a thin conodont-, glauconite-, and detrital pyrite-rich bed at the Genundewa-West River contact in all of these sections (see below).

Basal deposits of the West River Formation

From Alexander eastward into the Genesee Valley, and also in southern Erie County, the top contact of the Genundewa is gradational with the shale-dominated succession of the West River Formation (Figs. 5, 6, 8, 9-10). This is normally expressed as an upward change from massive styliolinid grainstone-packstone carbonate across one or two thinner, more mud-rich, top-Genundewa ledges, into a succession of alternating black and gray shale units that is typical of the West River Shale (Figs. 4-6, 8, 9). Generally, the West River is distinctly characterized by rhythmic black –gray shale alternations that thicken eastward and basinward as the West River thickens from 2.5-3 meters in thickness near Buffalo to approximately 22-25 meters in the Genesee Valley. This succession is marked by a few mappable layers ("Huddle Bed" conodont-glauconitic-rich styliolinid limestone layer near the base (Huddle, 1981); Bluff Point siltstone bed and associated goniatite layer in the upper middle part (Kirchgasser and House, 1981; House and Kirchgasser, 1993 and in press) ; and the Williamsburgh Bed of reworked, detrital pyrite and conodonts near the top. Herein, we will focus only on the "Huddle Bed" and a lower styliolinid limestone layer near the base of the West River succession.

At Fall Brook (Loc. 30) near Geneseo, the "Huddle Bed" is a 5-6 cm-thick bed of styliolinid limestone with a thin zone of conodont- and glauconite-rich lag debris at its base (Huddle, 1981; Kirchgasser et al., 1994, fig. 10).This unit (USGS Sample/horizon SD 8122) occurs 2.2 meters above the top of the Genundewa, and is 1.3 meters above another, lower bed of styliolinid limestone in that same section. Recent mapping by Baird shows that these beds thin and converge downward and westward towards the Genundewa as the West River succession thins due to condensation. At Spezzano Ravine (Loc. 28) on the west side of the Genesee Valley, the lower and upper styliolinid – rich limestone beds are respectively 0.6 and 1.7 meters above the Genundewa; at Schumacher Ravine (Loc. 26) near Pavilion, they are 9.0 cm and 73 cm above the Genundewa, and at a creek south of Suicide Corners (Loc. 24), the lower limestone bed has merged onto the top of the Genundewa, and the inferred Huddle Bed is only 30 cm above the Genundewa (Figs. 6, 8E). At Alexander (Loc. 19) the Huddle Bed appears to be only 8 cm above the base of the West River Shale. West of the Alexander meridian the Huddle Bed can no longer be seen as a discrete unit within the West River. In the Schumacher Creek-Alexander region the Huddle Bed is much thinner than it is in the Genesee Valley, so it could be reasonably inferred to be lost as a mappable layer west of Alexander. However, discovery of the ultra-thin North Evans/Genundewa sections at Elevenmile, Durkee, Little Buffalo and Buffalo creeks, which are all marked at their tops by a thin, conodont-, glauconite-, and detrital pyrite-rich bed, offers the possibility that the Huddle Bed is erosionally juxtaposed onto the North Evans/Genundewa deposit at these localities (Figs. 8B, D, 10, 11). The extraordinary conodont richness of this capping lag layer at these sections can be explained as due to local cannibalization of the North Evans conodont fraction by the younger sub-Huddle bed discontinuity (see Figs. 8B, D, 10, 11). A key test of these field observations will be to identify conodonts so far sampled from the putative Huddle Bed at these localities. At Fall Brook the Huddle Bed yields the conodonts *Ancyrodella recta* and *A. rugosa* indicating MN Zone 3, the presumed age of the upper division of the Genundewa Limestone in that section.

Another potential test of this idea can be accomplished by the reconstruction of this stratigraphic pattern along a different, independent transect. At Eighteenmile Creek (Loc. 3) and on the Lake Erie shore at Pike Creek (Loc.1) two outlier styliolinid beds occur, approximately 20 cm and 35 cm above the Genundewa Limestone within the basal West River Shale (Fig. 9a, c). The higher of these two beds is conspicuously marked by abundant conodonts and glauconite grains along its base, suggesting the horizon of the Huddle Bed. Proceeding northeast across western Erie County, the two beds appear to descend towards the underlying Genundewa, such that these beds are respectively 1.5 cm and 18 cm above that unit at "KB" Creek in Bayview (Loc. 6). At Cazenovia Creek (Loc. 11; see STOP 2) both beds are essentially in contact with the top of the Genundewa and would be semantically (and mistakenly) be called "Genundewa" (Figs. 4, 8A, 12). This pattern would similarly explain the drastic eastward thinning of the Gennundewa towards Buffalo Creek as being due, in part, to erosional overstep by the sub-Huddle Bed contact (Figs. 10, 11). Hence, this pattern is a mirror image of westward overstep by the sub-Huddle Bed contact to the west of Alexander (Figs. 4, 5, 10).

MEDIAL GENESEE GROUP DEPOSITIONAL TRENDS

Recent mapping in preparation of this fieldtrip paper has identified regional patterns of deposition and erosion within the Genundewa Formation and superjacent basal West River Formation succession (Figs. 10, 11). In particular, the mirror image opposing patterns of erosional overstep onto North Evans/Genundewa deposits serve to approximate depositional trends in the study area. An east-northeast trending, projected, erosional limit of the Genundewa deposit is suggested by the relationship of the broadly curving Genundewa outcrop belt east of Buffalo and patterns of condensation and erosional overstep within the overall North Evans-Huddle Bed interval (Figs. 10, 11). The trend of most abrupt lateral changes within this interval should project along a north-northwest-south-southeast transect nearly normal to the outcrop belt east of Buffalo; Figure 11 is an idealized transect across inferred paleoenvironmental strike for the eastern Erie County region. Two major processes account for the northwestward loss of the medial Genesee condensed carbonate units. The first is the overstep of the North Evans/Genundewa interval by the Huddle Bed discontinuity. The second is inferred onlap of the upper Genundewa division onto the top-Penn Yan disconformity. Just as the Geneseo Formation onlaps to near extinction onto the Taghanic disconformity (Fig. 1) the Genundewa should similarly terminate onto the foreland basin erosional slope (Fig. 11). Some of the drastic thinning of the North Evans/Genundewa interval appears to reflect this as exemplified by its localized thinning and thickening around minor irregularities on the top-Penn Yan disconformity surface that may have been produced by differential mud compaction around the underlying grotesque concretions (Figs. 8B, D). This implies that styliolinid limestone passes laterally (upslope) into North Evans facies as the erosional limit is approached (Fig. 11). It also suggests that within Genesee Formation deposits north of the present-day outcrop limit, that are now eroded away, the West River Shale would have overlain the Penn Yan Formation directly with probable development of a "Leicester-type" detrital pyrite lag deposit on the intervening disconformity (Fig. 11).

Finally, it is possible to rationally place the medial Genesee Formation succession into a sequence stratigraphic framework. The top-Penn Yan contact below the lower Genundewa division marks a lowstand event as indicated by the sub-Genundewa erosional contact and the marked facies discontinuity between the top-Penn Yan black shale facies interval below and the *Koenenites*-rich styliolinid granstone beds above. Some of the falling stage systems tract may be evidenced by a styliolinid limestone bed seen in the grotesque concretion interval of the topmost Penn Yan at some localities. The intra-Genundewa North Evans lag, of greater implied temporal significance is believed to be a hybrid sequence boundary – basin flexure generated feature; it underlies a generally transgressive interval of condensed pelagic carbonate overlain by basinal shale facies which would be consistent with a general eustatic transgressive systems tract succession. However, this spectacular lag unit appears to grade laterally to continuity within a facies of styliolinid carbonate in the Genesee Valley; this suggests only a modest lowstand event in that area, only 80 kilometers east of Buffalo. The pronounced development of the North Evans mainly in western Genesee and Erie counties suggests that localized basin margin upwarp may have been as strong an influence as eustasy in producing the North Evans lag bed. The upper Genundewa – lower West River succession is a classic transgressive interval with condensed styliolinid limestone, recording dysoxic conditions and terrigenous sediment-starvation, that is succeeded by dark West River shale that records severe dysoxia to near-anoxia. The Huddle Bed fits well into this picture as a maximum flooding surface unit recording the end of transgression and sediment-starvation and the beginning of early highstand conditions in the basin.

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LOCALITY REGISTER

Localities cited in text and illustrated in figures are listed below. Although a long succession of rock divisions may be present in many of these sections, only Genesee Group units, relevant to this study are listed in the descriptions below.

Lake Erie shore area localities (see Fig. 9)

Locality 1: Lake Erie shore bluff Genundewa Formation section south of Pike Creek (Saturday A4 fieldtrip STOP 2).

Locality 2: Genundewa Formation outcrop on Lakeshore Road opposite fishermen's parking area immediately south of Eighteenmile Creek (see STOP 1).

Locality 3: North Evans type section and overlying Genundewa and West River Formation successions below Amtrack railroad overpass along Eighteenmile Creek.

South Buffalo area localities where North Evans lag deposit is merged into overlying Genundewa Limestone as a composite limestone unit, and the North Evans directly overlies the Windom Member (not shown as figured graphic sections)

Locality 4: North Evans and overlying Genundewa limestone formation above Windom Member in creek ditch below Amsdell Road in town of Wanakah.

Locality 5: North Evans and overlying Genundewa limestone formation in abandoned quarry bordering Amtrack railroad tracks southwest of Cloverbank Road in town of Wanakah.

Locality 6: North Evans/Genundewa section on unnamed northwest-flowing stream ("KB creek") adjacent to WKBW radio mast south of Big Tree Road in the town of Bayview.

Locality 7: North Evans/Genundewa section at south edge of abandoned Penn Dixie quarry in the town of Bayview. This is now managed as a fossil park and nature center by the Hamburg Natural History Society Inc. (see Sunday fieldtrip B5, this volume).

Locality 8: North Evans/Genundewa section along New York State Thruway (roadcut along northbound lane of I-90) immediately north of Big Tree Road overpass.

Locality 9: North Evans/Genundewa section along the east fork of the south branch of Smoke Creek. Section is downstream from (west of) the California Road overpass in the town of Windom.

Locality 10: North Evans/Genundewa section on the north branch of Smoke Creek downstream from the Lake Road overpass and west of Boldt Road. .

North Evans/Genundewa sections underlain by lower Genesee Formation strata (see Figs. 4-6, 12-14)

Locality 11: Leicester Pyrite-West River Shale succession on large, east-facing cutbank of Cazenovia Creek, east of Northrup Road and 0.8 miles southwest of Springbrook (see STOP 2).

Locality 12: Leicester Pyrite-West River Shale succession on east-facing cutbank along Buffalo Creek, 0.65 mile south of Bullis Road overpass and immediately east of Iroquois High School.

Locality 13: Leicester Pyrite-West River Shale succession on Little Buffalo Creek 0.5-0.65 mile northwest of (downstream from) Bullis Road-Two Rod Road intersection in Marilla.

Locality 14a: Good Leicester Pyrite-West River Shale succession on Cayuga Creek immediately upstream from (south of) the Clinton Street overpass, 3.0 miles south-southwest of Alden.

Locality 14b: Leicester Pyrite-West River Shale succession along west-flowing tributary of Cayuga Creek, 0.2-0.35 mile west of (downstream from) Exchange Street overpass, 2.6 miles south of Alden (see STOP 3).

Locality 15: Leicester Pyrite-West River Shale succession on Durkee Creek, a tributary of Spring Creek. Section is 0.1-0.25 mile southeast of (upstream from) County Line Road-Seven Day Road junction, 2.0 miles southeast of Alden.

Locality 16: Leicester Pyrite-West River Shale succession along Elevenmile Creek, 0.1- 0.3 mile downstream from (west of) Warner Road overpass and 0.5-0.6 mile south of US 20.

Locality 17: Discontinuous Leicester Pyrite-basal West River succession on Murder Creek immediately east of hamlet of Griswold, between 0.6 and 1.0 mile southeast of Darien.

Locality 18: Discontinuous Leicester Pyrite-Genundewa Limestone succession on Bowen Brook, 0.4-0.65 mile north of US 20, 1.8 miles west-northwest of Alexander.

Locality 19: Leicester Pyrite-basal West River Shale succession on unnamed, east-flowing creek at Alexander Town Park, 0.5 mile southwest of Alexander .

Locality 20: Poor, discontinuous Geneseo Formation-Penn Yan Formation succession on unnamed southeast-flowing gully 0.6 mile north of Alexander.

Locality 21: Small northwest-flowing gully exposing upper part of Penn Yan Formation and Genundewa Limestone, 0.15 mile south of (upstream from) US 20 overpass, 2.5 miles east of Alexander.

Locality 22: Waterfall section on east-flowing tributary of Little Tonowanda creek essentially at junction of the side stream with the larger creek. Waterfall section is 0.25 mile north of a large, Genundewa Limestone-capped waterfall on Little Tonowanda Creek at Linden.

Locality 23: Genundewa Formation section in roadcut along US 20 at Bethany, NY. Outcrop is developed on both sides of US 20 beneath the Bethany Center Road overpass (see STOP 4)

Locality 24: Upper Penn Yan Formation-through-West River Shale Formation on unnamed, west-flowing tributary of Black Creek, 0.75 mile west of East Road and 1.1 miles southwest of Suicide Corners on US 20.

Locality 25: Discontinuous Leicester Pyrite-Genundewa Limestone succession on White Creek which crosses US 20 0.75 mile east of the intersection called Suicide Corners. The Geneseo and Penn Yan Formations are exposed downstream from (north of) the US 20 overpass and the Genundewa Limestone is exposed upstream from (south of) that overpass.

Locality 26: Penn Yan Formation-West River Shale Formation succession exposed on unnamed, southeast-flowing creek west of Roanoke Road and immediately upstream from the Hudson Road overpass, 1.3 miles northwest of the hamlet of Pearl Creek. Locality is on property of the large Schumacher dairy farm, hence the informal name *"Schumacher Creek"* for this locality.

Locality 27: Geneseo Formation succession exposed on small, unnamed, northeast-flowing gully, 1.25 miles southeast of the junction of routes US 20 and NY 63 at Texaco Town. Section is immediately downstream from (adjacent to) the NY 63 overpass, 0.25 mile southeast of the junction of NY 63 and Roanoke Road.

Genesee Valley outcrops cited in text but not shown as graphic sections

Locality 28: Leicester Pyrite-West River Shale Formation succession exposed along east-flowing Spezzano Ravine, 0.5 mile west of (upstream from) the NY 36 overpass, 0.75 mile southwest of the hamlet of Wadsworth.

Locality 29: Leicester Pyrite-West River Shale Formation succession along east-flowing Taunton Gully, 1.8 miles northwest of Leicester. Section is 0.0 to 0.5 mile east of (downstream from) Botsford Road overpass.

Locality 30: Classic Leicester Pyrite-West River Shale Formation succession along west-flowing Fall Brook, 1.3 miles south-southwest of Geneseo. Steep, Genundewa Limestone-capped waterfall and gorge immediately west of access car pull off just to the north of the NY 63 overpass over creek.

ROADLOG AND STOP DESCRIPTIONS

Leave the Adam's Mark Hotel in downtown Buffalo and head south to the junction with I-90. Enter I-90 eastbound towards to junction of I-90 and the New York State Thruway (I-90). The road log starts at the junction of I-190 and the Thruway where we will enter the Thruway (I-90) proceeding in the southbound direction.

Accumulated Miles	Incremental Miles	Road log description
0.0	0.0	Enter New York State Thruway (I-90) from I-190; proceed south on I-90 towards Erie, Pa.
2.0	2.0	NY 400 Expressway (overpass). Continue straight (south) on I-90.
2.45	0.45	Cross Cazenovia Creek.
3.4	1.95	Lackawanna Toll Plaza of New York State Thruway; continue straight on Thruway.
4.4	1.0	Cross south branch of Smoke Creek.
5.15	0.75	Exit to the right for NY 179 (Mile Strip Road) at the Blasdell exit of the Thruway. Pass through exit toll gate.
5.45	0.3	Junction of Thruway exit feeder with NY 179 (Mile Strip Road); turn right and proceed west on Mile Strip Road.
6.85	1.4	Junction of Mile Strip Road with NY 5. Enter (merge right) into junction rotary feeder which turns left (to south).
7.05	0.2	Exit junction rotary (to right) for entrance feeder to NY 5 southbound.
7.15	0.1	Merge into southbound NY 5. Ford plant on the left.
7.95	0.8	Fork to left for Hamburg. Bear right (straight) on NY 5.
8.65	0.7	View of Lake Erie to the right. Bluff exposure is Ledyard Member of Ludlowville Formation (Middle Devonian).
8.75	0.1	Entering town of Athol Springs. We will continue through communities of Athol Springs, Mount Vernon, and Wanakah.
12.35	3.6	Junction of NY 5 with Lake Shore Road. Bear right onto Lake Shore Road.

15.55 3.2 Bridge over Eighteenmile creek. Upstream cutbank to the left shows the Middle Devonian Hamilton Group divisions unconformably overlain by Late Devonian strata of the Genesee Group.

15.65 0.1 Turn left immediately south of Eighteenmile Creek bridge into parking lot maintained for fishermen.

STOP 1. Taghanic disconformity exposed in small cut (Loc. 2) along Lake Shore Road opposite parking area (see Fig. 9b). Exit vehicles and proceed across Lakeshore Road to roadcut slightly south of the parking lot entrance. People must watch for traffic coming down the hill around the blind corner.

This roadcut displays, in upward succession, the upper medial part of the Windom Member of the Moscow Formation (Fig. 9a), the Taghanic disconformity, a thin, local expression of the North Evans lag deposit below the Genundewa Formation (Fig. 9c), a thin unnamed and very localized dark gray to black shale unit within the Genundewa Formation, and the resistant ledge of styliolinid limestone representing the upper division of the Genundewa Formation (Fig. 9d). The disconformity below the North Evans spans possibly 7 conodont chronozones representing approximately 2 million years of geologic time (see text); units observed in the Finger Lakes region and in eastern New York (topmost parts of Moscow Formation, Tully Formation, and lower half of Genesee Group succession) are missing here such that Late Devonian strata belong to the upper part of early Frasnian MN Zone 2 are juxtaposed onto Middle Devonian beds belonging to the Middle *varcus* chronozone (see text).

The North Evans is here represented by a thin, 1-2 cm-thick lag unit rich in pelmatozoan debris, glauconite grains, a conspicuous mixed conodont fauna, and fish teeth, incuding numerous small cladodont shark teeth and rare ptychtodont tritors (Fig. 9c). Secondarily oxidized (limonitized) detrital pyrite grains are common in this mix, especially at the base. The tubular clasts represent pyritized burrows that were exhumed from the underlying Windom Shale (see text). Conodonts at the bedding-plane contact with the Windom include *Ancyrodella recta* that indicates the upper part of Frasnian conodont MN Zone 2. Above the North Evans is a 28 cm-thick fissile dark shale unit that can be accessed below the Genundewa Limestone overhang. As noted in the text, this shale is extremely local in extent, being found only in the vicinity of the Lake Erie shore (Fig. 9). Because there is no lag deposit and implied discontinuity separating the shale from the overlying Genundewa upper division limestone bed at this section and at Pipe Creek, we believe that the shale unit is a local basinal subfacies of the upper Genundewa division or a post-North Evans Genundewa division not normally expressed elsewhere (see text). The presence of the dark shale unit roofing the North Evans in this section is believed to explain, in part, the thin, pyrite-rich character of the North Evans lag deposit in sections where it is present (Fig. 9). More intense dysoxia associated with the basinal subenvironment represented by onset of black shale deposition could explain local dissolution of the carbonate fraction normally seen in the North Evans (see text). The overlying ledge of the upper Genundewa division (Fig. 9d) displays the typical outcrop character of this unit. This bed is composed almost exclusively of uncompressed *Styliolina fissurella* with accessory occurrences of the small bivalve *Pterochaenia*, wood debris, and rare goniatites, and it probably represents an offshore dysoxic regime characterized by prolonged sediment starvation and surface water productivity supporting *Styliolina* as plankton.

		Return to vehicles. Retrace route back to the junction of Mile Strip Road and the I-90 exit in Blasdell, but continue east from there on Mile Strip Road.
25.85	10.2	Junction of Mile Strip Road with the New York State Thruway exit in Blasdell; continue straight (east) on Mile Strip Road.
26.05	0.2	Mile Strip Road crosses over New York State Thruway. Continue straight (east) on Mile Strip Road.
28.95	2.9	Junction of Mile Strip Road with US 20. Turn left and proceed to the northeast on US 20.
32.75	3.8	Junction of Transit Road with US 20. Turn right (south) onto Transit Road.
33.25	0.5	Junction of Transit Road with Northrup Road. Turn left onto Northrup Road.
33.5	0.25	Pull off onto shoulder of Northrup Road just west of Cazenovia Creek.

STOP 2. Large cutbank on west side of Cazenovia Creek (Loc. 11) displaying Genundewa Formation and associated units (see Fig. 12). Exit vehicles and proceed across private driveway and yard. Continue along graded trail down to floodplain of Cazenovia Creek. Turn left (north) and follow improvise trail through tall brush to outcrop approximately 300 meters north of the base of the graded trail. The improvised trail may be wet and muddy and Cazenovia Creek may be high making examination of the outcrop difficult.

This large, clean exposure shows several units to advantage (Fig.12); these are, in ascending order, the Amsdell Bed (*Emanuella praeumbona* –rich interval) of the Windom Member (Fig. 12a), the Leicester Pyrite of the Geneseo Formation (Fig. 12b), an undifferentiated interval of black and dark gray shale corresponding to the lower part of the Genesee Group (Fig.12c), a band of laterally separated large concretions corresponding to the "grotesque concretion" level (Fig. 12d), the North Evans lag deposit (Fig. 12e), a thin zone of nodular styliolinid limestone above the North Evans comprising the lower part of the Genundewa upper division (Fig. 12f), the massive, main part of the upper division of the Genundewa Formation (Fig.12f), a thin conodont- and glauconite-bearing lag layer (Fig.12g), and the West River Formation (Fig. 12h).

This section offers an excellent opportunity to view Leicester pyrite lenses along the Taghanic disconformity if the water is not too high. These lenses (Fig.12b), laterally separated by bank intervals lacking lag pyrite, may be profiles of channel fill deposits that accumulated on the Taghanic dysoxic erosional substrate (see text). They are composed of gravel-sized grains of detrital pyrite that correspond to reworked diagenetic pyrite derived mainly from the Windom Shale. Pyritic burrow tubes make up most of the Leicester clasts, but pyritic steinkerns of orthoconic and goniatitic cephalopods, as well as fish bones and teeth, can be seen.

The black-dark gray shale succession between the Leicester and North Evans Limestone beds (Fig.12c) represents the feather edge of a much thicker Geneseo Formation succession to the east (Figs. 4-6). At the north branch of Smoke Creek (Loc. 10), four miles west of here, the sub-North Evans disconformity is juxtaposed directly onto the Windom Member.

The North Evans lag (Fig.12e) contains a mixture of pelmatozoan debris, conodonts, glauconitic grains, reworked concretions, and fish bone/tooth debris. It grades upward into styliolinid grainstone facies of the overlying Genundewa Limestone. The Genundewa is quite thick at this section; it thins dramatically eastward from here, being essentially absent at Buffalo Creek (Loc. 12), the next section to the east (see Fig. 8 A,B). The top of the Genundewa at this section is marked by a 1.0 cm-thick lag unit (Fig.12g) rich in conodonts and glauconitic grains that may link to the "Huddle Bed" diastemic horizon recognized by others (see text). This discontinuity may be, in part, responsible for erosional overstep of portions of the Genundewa Formation seen in sections east of here (see text).

Return to vehicles. Continue north on Northrup Road.

34.5	1.0	Northrup Road crosses Cazenovia Creek. This classic fossil locality can be seen to the right; the Tichenor Limestone, basal division of the Moscow Formation, holds up the low waterfall and overlies fossiliferous shale of the Wanakah Member of the Ludlowville Formation. In the distance, the high bank displays the Windom Member and overlying Genesee Group strata.
34.8	0.3	Junction of Northrup Road with NY 16/78 at Springbrook. Turn left onto NY 16/78.
35.6	0.8	Junction of NY 16/78 with US 20 in Elma; turn right (north) onto US 20.
36.4	0.45	NY 400 Expressway overpass. Continue straight on US 20.
36.75	0.35	Junction of US 20 with Bullis Road; turn right (east) onto Bullis Road and proceed toward Marilla.
41.75	5.0	Bullis Road crosses Buffalo Creek. Basal limestone divisions (Tichenor Member, Menteth Member) of the Moscow Formation visible below bridge in floor of creek.
44.05	2.3	Junction of Bullis Road with Two Rod Road at intersection in Marilla; turn left (north) onto Two Rod Road.
45.2	1.15	Junction of Two Rod Road with Clinton Street (NY 354); turn right (east) on Clinton Street.

47.85 2.65 Clinton Street bridge over Cayuga Creek. Excellent exposure of Windom Member-West River succession along creek upstream from bridge. This section (Loc. 14a) has been used for field stops in the past, but access is a problem presently. We will, instead, employ a nearby section for STOP 3.

48.05 0.2 Junction of Clinton Street with Exchange Street; turn left (north) onto Exchange Street.

48.55 0.5 Exchange Street bridge over west-flowing tributary of Cayuga Creek; pull off into driveway or road shoulder adjacent to bridge.

STOP 3. Upper Windom Member-into-West River Formation succession exposed along west-flowing tributary of Cayuga Creek (Loc. 14b; see Fig. 13). Exit vehicles and proceed west across private yard and sloped wooded area behind yard to creek bottom. Proceed left (west) down the creek to a low waterfall where the Moscow Formation-Genesee Group contact can be seen.

At the waterfall black shale deposits of the Geneseo Formation rest abruptly on the Taghanic Disconformity which caps the sloped exposure of gray Windom Shale in the falls face. Lenses of Leicester Pyrite (Fig.13a) can be spotted owing to the trail of rust seepage developed below them on the outcrop face. The Geneseo Formation is marked mainly by a calcareous black shale interval (Fig.13b) that is capped by a discontinuity (Fig.13c) characterized by comminuted fossil debris and a few fish bones and conodonts. It may correlate to the top-Fir Tree discontinuity in the Finger Lakes region, but this idea has yet to be tested (see text). Strata, probably equivalent to the Lodi Limestone interval in the Finger Lakes area (Fig.13d), are cryptic in this area; we will skip over this part of the section. However, above a band of prominent, brown weathering concretions (Fig.13e), is a black shale-on-black shale discontinuity (Fig.13f) that is marked by a pavement of flattened *Styliolina* in association with conodonts and cuspate cladodont shark teeth. This contact can best be accessed at the minor falls just above the concretion band where it occurs 12 cm below a 3 cm-thick styliolinid bed (Fig.13g) which is the local condensed expression of the Elevenmile Creek bed.This conodont-bone-rich horizon (unit 13f) is found to correlate eastward to the Linden Bed (see text).

At a third low falls lip the North Evans-Genundewa succession (Fig.13i,j) is observed to overlie a conspicuous band of irregular "sculptured" concretions herein referred to as the *grotesque concretion layer* (Fig 13h). These concretions are interpreted as underbed concretions which formed in a zone of fluctuating redox conditions below the downward migrating top-Penn Yan erosion surface; concretions of this type that formed earlier can be seen as reworked clasts within the overlying North Evans lag deposit (Fig.13i). The North Evans is well developed here and can be easily sampled. The massive, nodular, falls-capping, upper Genundewa division ledge is also well developed at this locality (Fig.13j). A thin conodont- and glauconite-rich layer, herein interpreted to be the equivalent of the Huddle Bed, is present 10 cm above the Genundewa on nearby Cayuga Creek (Loc. 14a). However, on this creek that bed is very thin and cryptic; it probably is expressed as a thin styliolinid-rich layer (Fig.13k) 12 cm above the top of the Genundewa. The West River Formation is best viewed in the north-facing cutbank section immediately upstream from the Genundewa falls. Here it consists of thin alternations of gray and black shale with a few thin siltstone beds in the upper part (Fig.13).

Return to vehicles. Continue north on Exchange Street.

49.55 1.0 Junction of Exchange Street with Henskee Road; turn right (east) on Henskee Road.

50.45 0.9 Junction of Henskee Road with Sullivan Road; continue east on Henskee Road.

51.15 0.7 Junction of Henskee Road with County Line Road; turn right (north) onto County Line Road.

51.75 0.6 County Line Road crosses Durkee Creek, a tributary of Spring Creek. This shows a complete Windom Member-basal West River Formation succession (Loc. 15). Just beyond view is the thickest Leicester pyrite lens known with a maximum thickness of about 28 cm.

52.65 0.9 Junction of County Line Road with US 20; turn right (east) onto US 20 and continue towards Bethany Center.

54.25 1.6 Junction of Harlow Road and US 20. Darien Lakes State Park visible to north. Continue straight (east) on US 20.

54.75 0.5 Bridge over Elevenmile Creek. Long Devonian section on this creek extending from the lower part of the Ludlowville Formation up through the study interval, into the Middlesex Formation. When foliage is absent, the Tichenor and Menteth members of the Moscow Formation can be seen to cap a waterfall visible from the road.

56.45 1.7 Junction of NY 77 with US 20 in Darien Center; continue east on US 20.

58.25 1.8 Cross Murder Creek in Darien. Intermittent Moscow, Geneseo, Penn Yan, and Genundewa formation sections occur along this Creek (Loc. 17).

63.65 5.4 Junction of US 20 with NY 98 near Alexander; continue east on US 20.

66.35 2.7 Bridge over Little Tonowanda Creek. The Kashong Member-Windom Member boundary is visible in the creek floor to the north of the bridge. Good Genesee Group sections are developed much further upstream near Linden. Structural dip observed in the vicinity of this creek is the local expression of the Clarenden-Linden fault zone.

69.65 3.3 Exposure of Genundewa Formation exposed below the Bethany Center Road Bridge on both sides of US 20. We will examine this section momentarily as part of STOP 3.

69.85 0.2 Turn left (north) onto Old Telephone Road from US 20. Park vehicles on right hand shoulder of Old Telephone Road immediately north of the turn off of US 20.

STOP 4. Genundewa Formation exposure (Loc. 23) on both sides of US 20 below Bethany Center Road and loose Genundewa blocks in roadfill along US 20 immediately east of the Old Telephone Road junction with US 20 at Bethany, New York (Fig. 14). Exit vehicles and proceed west uphill along US 20 to the Bethany Center Road overpass. The Genundewa Limestone is displayed in the protected area directly under the bridge.

This section illustrates the twofold internal division of the Genundewa Limestone, characteristic of this area, with the lower and upper Genundewa styliolinid units (Fig.14, units c and e) separated by a complex, nodular, internal zone of conodont-rich lag debris corresponding to the North Evans Limestone (Fig.14, unit d). Of particular note is the presence of *Trypanites* (borings) into attached Penn Yan concretions (Fig.14a); this indicates the presence of a sub-Genundewa discontinuity (Fig.14b) in this area that is distinctly older than the intra-Genundewa North Evans contact (see text). Also important is the excellent development of cephalopod limestone facies within the lower Genundewa division (Fig.14c). We will see this facies to advantage in loose blocks in the US 20 road berm downhill past the Old Telephone Road exit where the vehicles are parked. This is one of the best examples of *cephalopodenkalk* facies to be seen in North America. Such facies is interpreted to represent deeper water dysoxic conditions where remains of pelagic (planktonic and nektonic) organisms accumulated in a sediment-starved setting (see text). The goniatite fauna includes *Koenenites styliophilus, Acanthoclymenia genundewa,* and *Tornoceras uniangulare compressum* (see Kirchgasser et al., 1994, for illustrations of Genesee Group goniatites). Pyritic replacements of *Acanthoclymenia* occur here at the contact with the overlying West River Formation.

Proceed east, downhill along US 20 to road fill berm past the Old Telephone Road pull off. Sample blocks exposed on the partly overgrown berm slope. Return to vehicles.

End of field trip.

THE PENN DIXIE SITE: A CLASSIC AND UNIQUE PALEONTOLOGICAL & OUTDOOR EDUCATION CENTER

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The Hamburg Natural History Society, Inc. (HNHS) is a nonprofit educational corporation that owns and operates The Penn Dixie Paleontological and Outdoor Education Center in Hamburg, New York (Fig. 1). The HNHS was founded in 1993 to promote the study of the natural sciences, with a particular emphasis on field activities associated with the geological and biological sciences. The HNHS offers a wide variety of hands-on educational programming to students of all ages, both at the Penn Dixie Site and off-site at local schools, libraries, and civic group meetings. Since its inception, the HNHS has expanded its educational curriculum to include public educational programming in astronomy and ornithology to complement its core study in geology and fossil collecting and identification. Unlike conventional museums or research facilities, the Penn Dixie Site is a hands-on outdoor educational facility—one at which visitors of all ages are encouraged to actually collect and keep 380-million-year-old fossils – "*Where Science Comes Alive*".

The site of a former quarry operation that was the source of calcareous shale excavated and used for cement aggregate by the Penn Dixie Cement Company. A majority of the 57-acre site was quarried until the late 1960s, during which time 9 to 10 feet of shale was removed from the surface. A gray, somewhat flat "desert-like" or "lunar landscape-appearing" surface now occupies a majority of the site. After quarry operations ceased, weathering forces began to expose 380 million-year-old Devonian fossils preserved within the Windom Shale. This highly fossiliferous unit underlies the entire site and provides an inexhaustible supply of fossils. In addition to the Windom Shale, several limestone units (the Genundewa, North Evans, and Tichenor outcrop on the surface. The Wanakah Shale is also exposed, underlying the Tichenor Limestone, in a tributary that flows into Rush Creek and in cliffs along Rush Creek on the northern section of the site. All of these units contain a variety of fossils.

PRESERVATION OF THE PENN DIXIE SITE

The HNHS administers and maintains the Penn Dixie Site, a 32.5-acre former shale quarry that was purchased by the Town of Hamburg in 1995 and deeded to the HNHS in 1996. The HNHS then took immediate steps to clean up the site and establish plans for its transformation into a

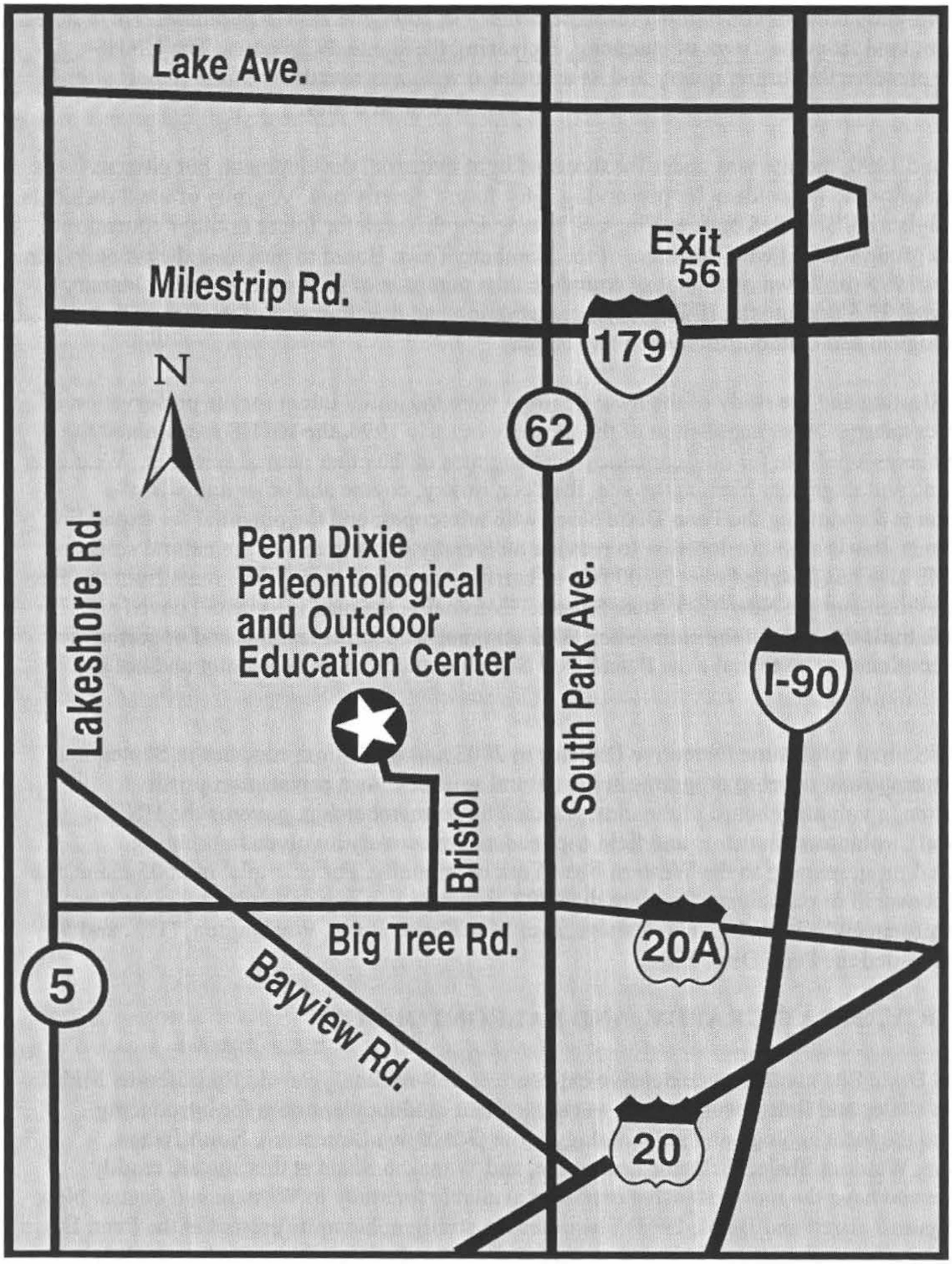


Figure 1. Location of Penn Dixie Paleontological and Outdoor Education Center in Hamburg, New York.

truly world-class outdoor educational resource center. In 2004, the HNHS purchased 16.75 acres of adjacent land from the Town of Hamburg, increasing the site to 49.25 acres. The HNHS' efforts to preserve the former quarry and its associated wetlands saved one of the richest sites of 380-million-year-old Devonian Era fossils in the eastern United States.

In 1989 and 1990, the site was under the threat of light industrial development, but citizens from the community had other ideas for preserving it for future generations. A group of local residents and geologists collaborated on acquiring and preserving this area for future outdoor educational use. This group worked with members of the Hamburg Town Board to purchase the property. In December 1995 the Town of Hamburg completed the purchase of the property and in January 1996 deeded 32.5 acres to the HNHS. The preservation and development of the Penn Dixie Paleontological and Outdoor Education had begun.

Fossil collecting and the study of the local geology were the initial intent for the preservation of this former quarry. After acquisition of the property in early 1996, the HNHS reexamined the other resources available for outdoor education programs in the other natural sciences. With over 143 nesting and migratory birds at the site; the deer, turkey, coyote and other animals; the spacious area for viewing the Penn Dixie Skies with telescopes; and the potential for expanding the wetlands, this is a unique location to provide a diversity of programs in the natural sciences. The HNHS also has installed over 2,100 feet of barrier-free paved trails with grants from the East Hill Foundation and the New York State Senate. Eventually, the plan is to install paved and boardwalk trails throughout the entire site. With all these wonderful features and opportunities, the goal continues to be to make the Penn Dixie Site an outdoor education center and not a museum.

The HNHS hired a full-time Executive Director in 2003 and a full-time educator in September 2004 to manage and develop programs in the natural sciences. As a private non-profit organization, a volunteer board of directors, elected by its membership, governs the HNHS. HNHS staff, volunteer educators and field trip leaders are actively involved in bringing educational programming to the Western New York community. For example, in 2005 alone, the HNHS sponsored or participated in more than 392 programs that were attended by more than 77,081 children and adults. In 2005, visitors from 34 different states, Washington, D.C., and 5 countries visited the Penn Dixie Site.

GEOLOGY, STRATIGRAPHY, AND PALEONTOLOGY

The Penn Dixie Site contains an extensive exposure of 380-million-year-old fossiliferous Middle Devonian shales and limestones, serving as an excellent outdoor classroom for introducing students to the local geology and paleontology. The Genudewa Limestone, North Evans Limestone, Windom Shale, Tichenor Limestone, and Wanakah Shale at this site are readily accessible and have the most extensive exposure available for study in Western and central New York. Figure 2 (Brett and Baird, 1982) illustrates the stratigraphic units present at the Penn Dixie Site. Prime exposures of these units are present (except for the West River Shale, which is mostly covered by overburden at the south end of the site). Brett (1974) and Baird and Brett (1982), along with Beuhler and Tesmer (1963), provide a detailed discussion of the stratigraphy and paleontology of these units. The warm tropical seas that covered this region of Western New York 380 million-years ago, when the region was 20 to 30 degrees south of the equator, provided an environment conducive to a variety of invertebrate and vertebrate animals. The shales and limestones that formed during this time period preserved the remains of the diverse and abundant

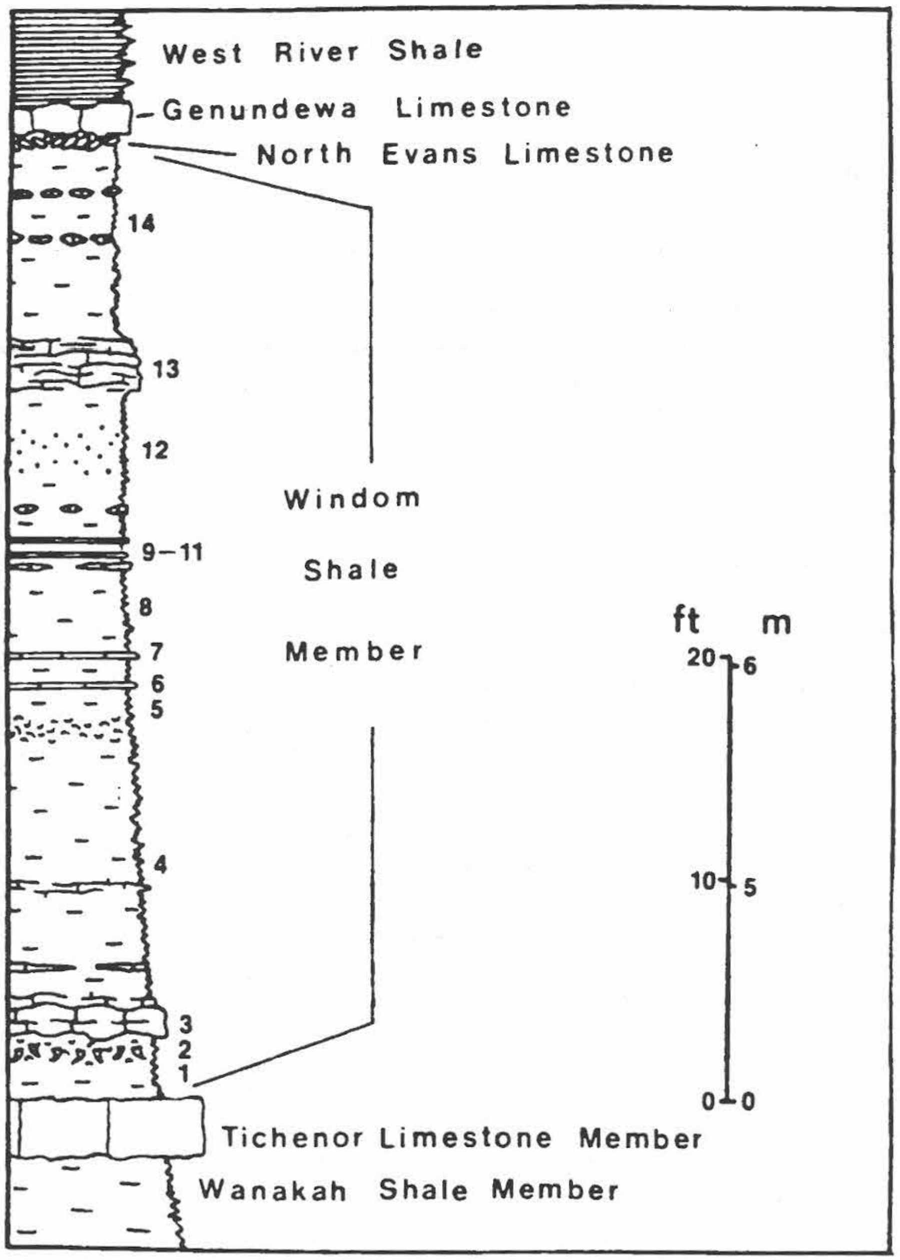


Figure 2. Stratigraphic units present at the Penn Dixie Site. Stratigraphic subdivisions of the Windom Shale Member; standard section at Penn Dixie and unnamed creek near Big Tree; Units include: 1) *Ambocoelia umbonata* beds; 2) Bay View coral bed; 3) Smoke Creek bed; 4) barren shale interval; 5) Big Tree bed; 6, 7) A-B limestones; 8) Buffalo pyritic beds; 13) Amsdell bed; 14) upper *Ambocoelia? praeumbona*-bearing Shales. (Modified from Brett and Baird, 1982).

fauna that occupied these seas. The following brief discussion of the units present on the site begins with the lower Wanakah Shale at the north end through the West River Shale to the south.

Wanakah Shale

The Wanakah Shale is a medium-gray to light-blue gray calcareous shale that weathers to a sticky clay. The Wanakah is exposed in the northeast section of the site in a tributary to Rush Creek and in the high banks on the south side of Rush Creek. The tributary is a popular area for fossil collecting, viewing the large calcareous concretions, and some pyritized burrows, rather than the steeper cliffs along Rush Creek. Brachiopods, bryozoans, trilobites, gastropods, pelecypods, echinoderms, corals, sponges, ostracodes, and some pyritized fossils may be found. Limited area in the tributary does not provide access for large groups.

Tichenor Limestone

The Tichenor Limestone overlies the Wanakah Shale and outcrops at the northern end of the site. Pyrite coating the surface of the Tichenor has weathered, exhibiting a reddish-rusty color that stands out from the surrounding overlying gray Windom Shale. At the northeast section of the site, an unexplained domal feature of the Tichenor, with several feet of relief, is present. This feature is not believed to be a result of the quarrying operation, but possibly from glacial rebound. A large exposure of the eroded limestone surface is adjacent to this feature and extends north to one of the on-site ponds. This area is often referred to as "crinoid heaven" due to the countless number of pelmatozoan columnals that are found lying on the surface. The Tichenor Limestone contains corals, brachiopods, pelecypods, trilobites, bryozoans, and echinoderms, all of which are difficult to remove from the hard limestone. The Tichenor Limestone is approximately 1.5 to 2 feet thick and underlies most of the site, dipping to the south-southwest along with the other units on site.

Windom Shale

The Windom Shale is a medium to dark gray, variably calcareous mudstone with several thin argillaceous limestones, concretionary beds, and pyretic horizons (Beuhler and Tesmer, 1963). In addition, at the southwest portion of the site there is an excellent exposure of phosphate nodules covering the surface. The Windom also weathers to a sticky clay. The Penn Dixie site has the most complete and best exposure of Windom Shale in New York State, approximately 42 feet thick. Brett and Baird (1982) described 14 subdivisions within the Windom that could be recognized at this location (Fig. 2). Fossil assemblage zones were described in Brett (1974) and Brett and Baird (1982). A disconformable basal contact with the Tichenor Limestone is exposed in the domal outcrop in the northeast section of the site. The upper Windom beds have been scoured, and shale clasts can be observed in the overlying North Evans Limestone. The Windom contains a variety of corals, brachiopods, pelmatozoan columnals, bryozoans, trilobites, gastropods, pelecypods, cephalopods, and more rarely fish remains, plant material, and blastoid and crinoid calices. The upper Windom has a variety of pyritized fossils, burrows, and most-likely fecal remains weathering out on the surface. Some of the pyritized fossils include brachiopods, pelecypods, cephalopods, trilobites, and blastoids (Fig. 3). The weathering shale exposes thousands of specimens lying on the surface, waiting to be found after 380 million years.

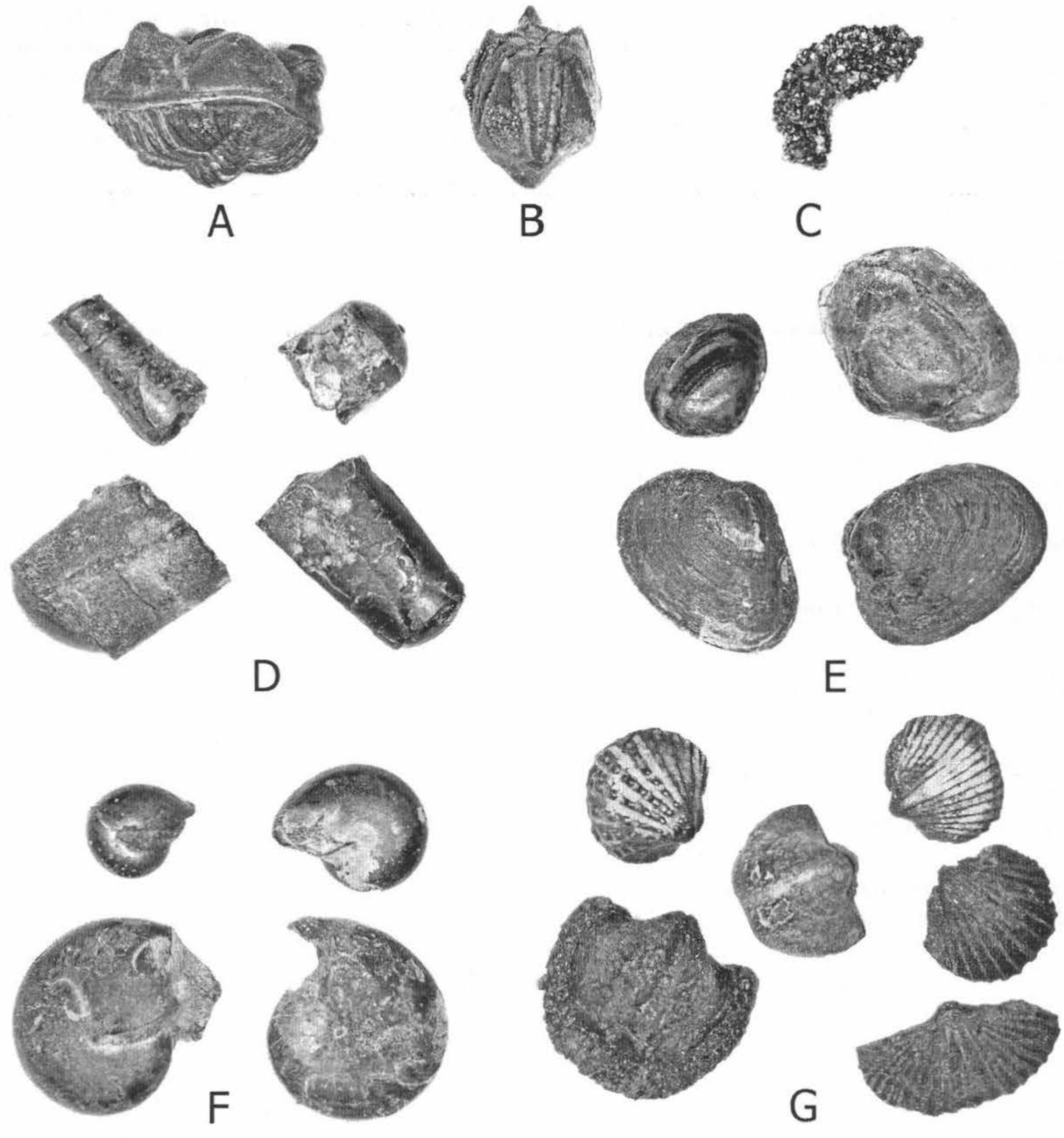


Figure 3. Pyritized fossils that weathered out of the Penn Dixie pyrite beds at the southern end of the site from the Upper Windom Shale. A-*Greenops* sp., enrolled, 0.7 cm wide; B- Blastoid calyx, 0.5 cm wide; C-pyrite, 1 cm long – burrows, nodules, or other shapes may weather out; D-Cephalopods, *Michlinoceras* sp. , largest 1 cm across; E-Pelecypods, largest 1 cm; F-Cephalopods, *Tornoceras* sp. , smallest is 0.4 cm across; G-Brachiopods, largest is 1 cm across; Blastoid was collected by Amanda Czechowski, pyrite specimens collected by Richard Spencer, and all other specimens collected by the author.

Enrolled trilobites can be commonly found washed out of the shale after a good rainstorm, along with horn corals, brachiopods, and pelmatozoan columnals. Multiple complete trilobites on a slab have been collected from the Lower Windom and complete specimens of *Phacops rana,* like the specimen in Figure 4, keep collectors returning for their perfect specimen. Sections of the Windom are not as fossiliferous as others, but careful study of the stratigraphic subdivisions identified by Brett and Baird (1982) will yield some interesting discoveries. In addition, Penn Dixie staff and volunteer guides will direct visitors to the better collecting areas on the site.

North Evans Limestone

The North Evans Limestone is a buff-colored, weathered dark-gray crinoidal limestone that is 1.5 to 4 inches thick and contains angular clasts derived form the underlying Windom Shale. Erosional lag concentrations of hiatus concretions, pelmatozoan fragments, conodonts, fish plates, teeth, and mandibles, along with some brachiopod valves, are present (Brett and Baird, 1982). Carbonized plant remains are also found in this unit. Although a variety of fish remains have been found at the Penn Dixie Site (Fig. 4), they are difficult to find even with the good exposure of North Evans present. The buff-colored weathered surface of the North Evans and bone material make this unit easily recognizable.

Genundewa Limestone

The Genundewa Limestone is a nodular, medium dark-gray, poorly bedded limestone that weathers to a light gray, which has been referred to as the "Styliolina Limestone" directly overlying the North Evans Limestone (Buehler and Tesmer, 1963). Carbonized wood can be frequently found, but other examples of the fauna are more difficult to obtain.

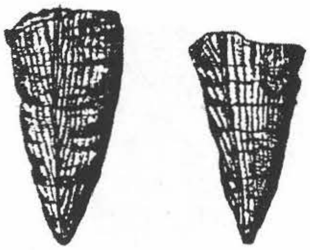
West River Shale

The West River Shale is dark gray to black in color and overlies the Genundewa Limestone. Most of this unit is covered by overburden at Penn Dixie and Eighteen Mile Creek provides a better opportunity to view this unit. Conodonts, cephalopods, pelecypods, and fish remains have been reported from the West River Shale at other localities in Western New York (Buehler and Tesmer, 1963).

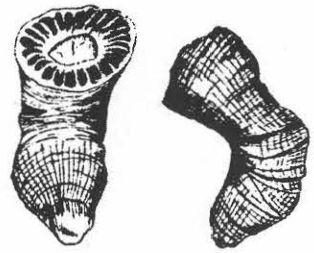
The preservation, diversity, abundance of fossils, and the extensive bedrock exposures at the Penn Dixie Site makes this an excellent outdoor classroom for students as well as amateur and professional paleontologists to be introduced to Western New York geology and paleontology. In addition, students and possible future scientists from pre-school through college are being introduced to the rich geologic history of Western New York by the thousands each year. Plates 1 through 4 illustrate some of the more common fossils that can be found at the Penn Dixie Site. Weathering of the Windom Shale results in many corals, brachiopods, pelmatozoan columnals, and trilobites being continually exposed on the surface. Those who extend the effort to dig into shale are rewarded with an extensive introduction to the variety of fossils preserved within the Windom. The northern section of the site provides an excellent outdoor classroom for students and visitors to be introduced to fossils and the local geology. Many specimens found at Penn Dixie can be viewed on the web site at www.penndixie.org.

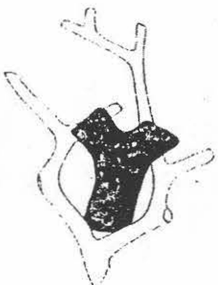
Plate 1
Fossils of the Penn Dixie Site

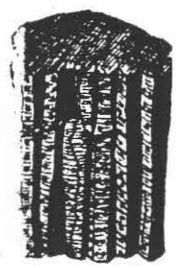
CORALS

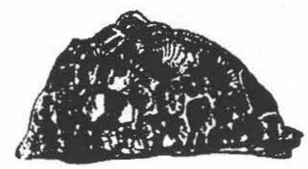

Sterolasma rectum


Cystophyllum americanum

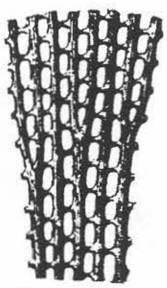

Amplexiphylum hamiltoniae


Trachypora sp.

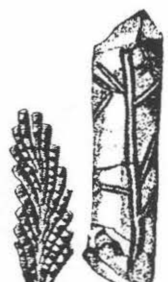

Favosites hamiltomiae


Pleurodictyum americanum

BRYOZOANS


Fenestella sp.


Hederella sp.

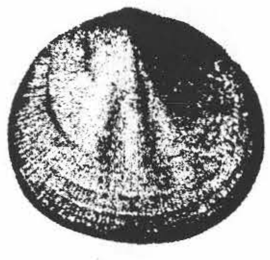

Reptaria stolonifera

Drawings from "Geology and Palaeontology of Eighteen Mile Creek" by Amadeus Grabau
Plates compiled by Scott Clark

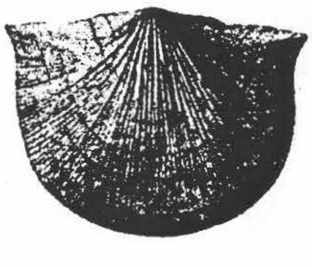
Plate 2
Fossils of the Penn Dixie Site

BRACHIOPODS

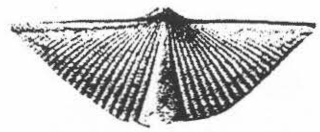
Orbiculiodea sp.



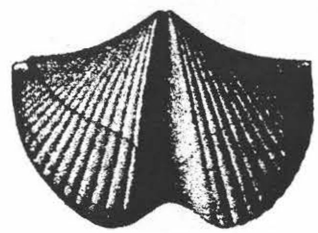
Rhipidomella sp.



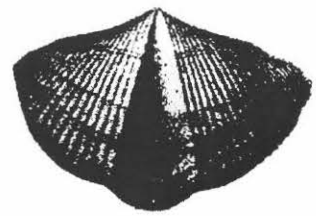
Stropheodonta demissa



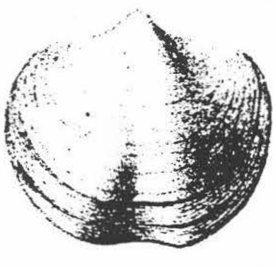
Mucrospirifer mucronatus



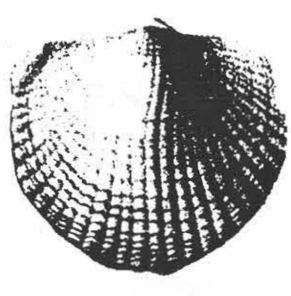
Spinocyrtia granulosa



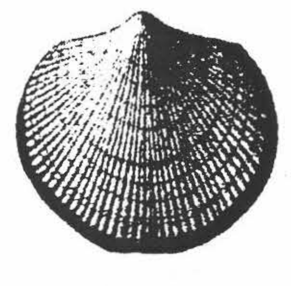
Mediospirider auduculus



Athyris spiriferoides



Spinatrypa spinosa

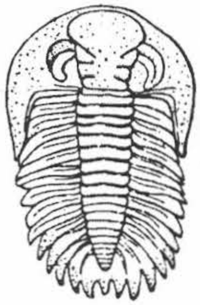


Pseudoatrypa devonica

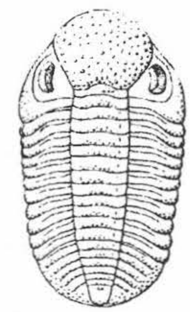
Drawings from "Geology and Palaeontology of Eighteen Mile Creek" by Amadeus Grabau
Plates compiled by Scott Clark

Plate 3
Fossils of the Penn Dixie Site

TRILOBITES



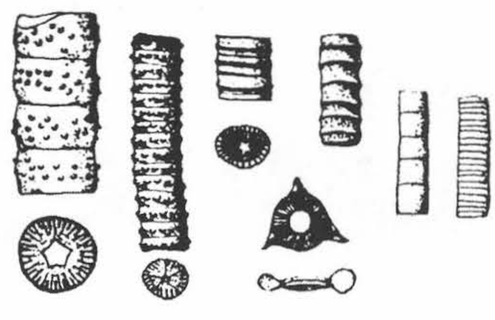
Greenops boothi



Phacops rana

CRINOIDS

Ancyrocrinus bulbosus

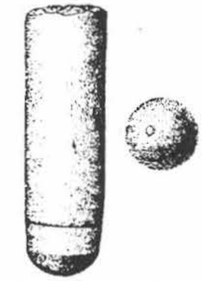


Various Crinoid segments

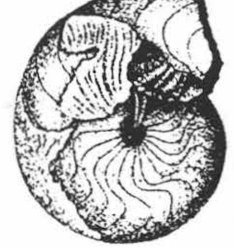
Drawings from "Geology and Palaeontology of Eighteen Mile Creek" by Amadeus Grabau
Plates compiled by Scott Clark

Plate 4
Fossils of the Penn Dixie Site

CEPHALOPODS



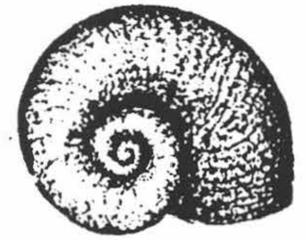
Michlenoceras sp.



Tornoceras uniangulare

Spyroceras sp.

GASTROPODS

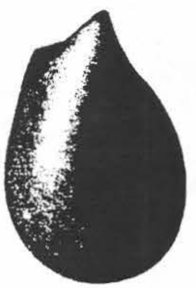


Naticonema lineata

PELECYPODS

Pterinopecten sp.

Palaeoneilo sp.



Plethomytilus sp.

Drawings from "Geology and Palaeontology of Eighteen Mile Creek" by Amadeus Grabau
Plates compiled by Scott Clark

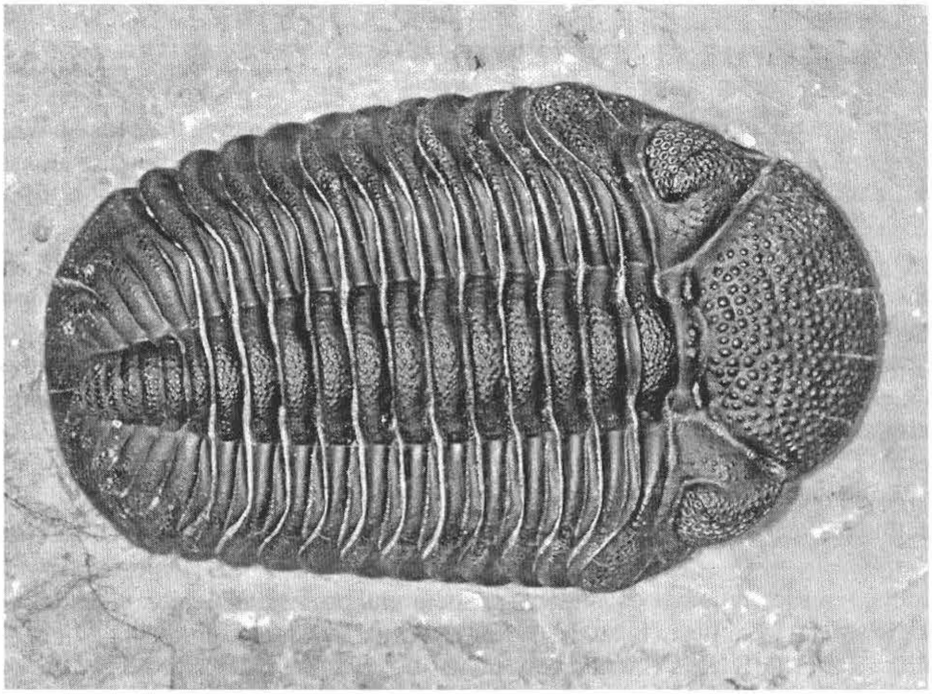


Figure 4. ***Phacops rana*** **collected by Jon Luellen and prepared by Gerry Kloc. Collected during the Dig with the Experts in May 2005 from the Lower Windom Shale at Penn Dixie. Trilobite is 2 inches long.**

Figure 5. Fossil fish bone material, 5.6 cm in the longest dimension, collected from weathered North Evans Limestone during August 2006 by the author.

CURRENT PROGRAMS AT PENN DIXIE

The opportunity to actually find and collect ancient creatures that roamed the seas of Western New York 380 million-years ago fascinates children and adults alike. Children are amazed that these fossils are older than the dinosaurs and their parents and that they can keep what they find! The Penn Dixie Site provides an opportunity to open a whole new world of geology and paleontology, along with the other natural sciences, to students, scouts, senior citizens, and the general public. It has provided an outdoor educational experience for many hearing impaired, visually impaired, totally blind, and physically challenged individuals. The preservation and continued development of this site is extremely important. Commercial and residential development, along with landowners restricting property access, have made many fossil collecting sites extinct or no longer accessible. In addition, Penn Dixie can accommodate large groups, whereas many streambeds, road and railroad cuts, and shoreline exposures cannot or are impractical. Attempts to preserve collecting sites, such as Penn Dixie, must be made, or many classic collecting and geologic locations will be lost for future generations to visit and study.

Students, scouts, families, summer day camps, amateur and professional geologists find this classic geologic site an ideal place to study geology, collect fossils, observe over 143 nesting and migratory birds, view the WNY skies and explore nature. Guided tours, astronomy programs, birthday parties, Boy and Girl Scout activities, corporate and civic picnics, and family outings are available by reservation.

The HNHS has scheduled a variety of programs for the 2006 season. The Penn Dixie Site is open to the public every Saturday, from May through October 9 AM to 4 PM; Monday through Saturday, 9 AM to 4 PM, mid-June through August to collect fossils. Group, birthday party, corporate, visits, or other events may be scheduled by calling (716) 627-4560. Events are held rain or shine. Special Events, evening astronomy programs, bird walks, summer day camps, group and family tours are held throughout the year. Some of the Special Events and activities scheduled for 2006 include:

- "Dig with the Experts" on May 20th
- 12th Annual Children's Day on June 4th
- 7th Annual Miss Buffalo Nature Cruise and Buffalo Lighthouse Tour on June 11th
- "Big Toys Event" on July 16th
- Mid-Summer's Night Adventure on August 12th
- Special Event on September 11th
- 9th Annual WNY Earth Science Day Celebration on September 30
- 4th Annual Scare-assic Park Halloween Event – LOST TREASURE OF THE GHOSTLY PIRATES on October 7.

Evening astronomy programs are held at Penn Dixie one Saturday night a month April through November. Visits may be scheduled at other times by calling Penn Dixie at (716) 627-4560. Additional on-site and off-site events are open to the public, which are listed on the Penn Dixie website www.penndixie.org.

A GROWING MEMBERSHIP ORGANIZATION

The HNHS has experienced phenomenal growth since its inception only thirteen years ago. While most of the HNHS members come from Western New York, the society

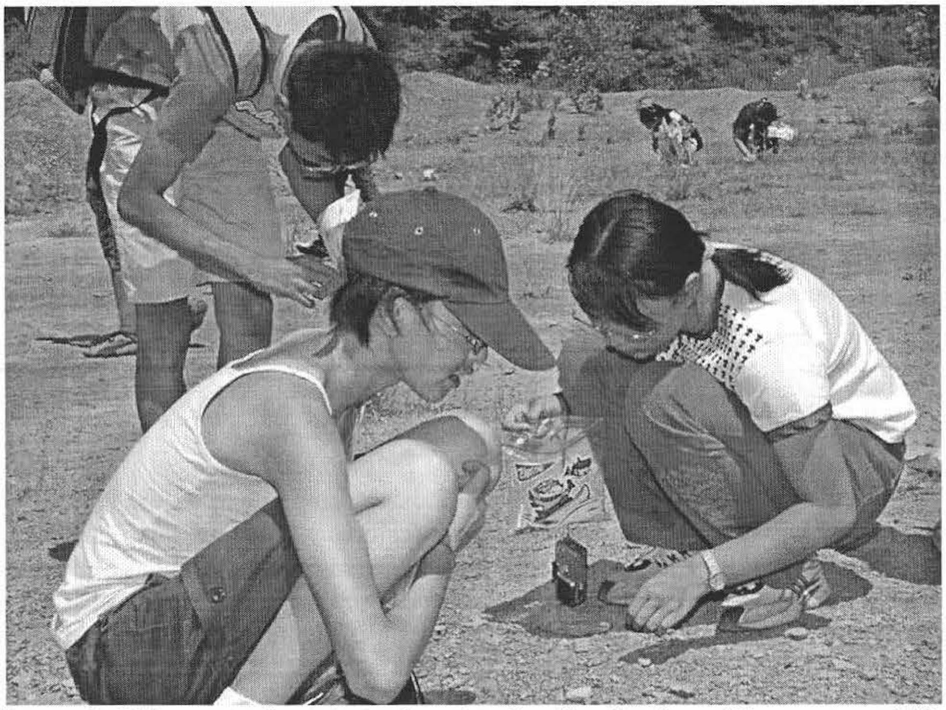


Figure 6. Chinese students introduced to fossil collecting with a group of 40 Chinese and Algerian students that visited Penn Dixie in 2005.

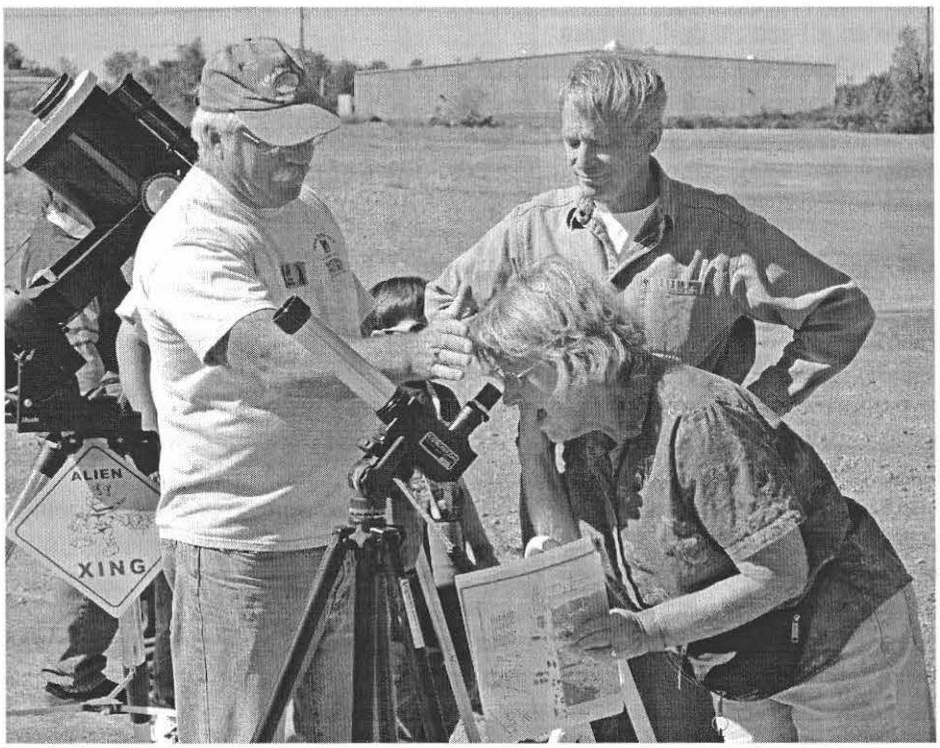


Figure 7. Visitors viewing solar flares and sunspots through filtered telescopes during the 8th Annual WNY Earth Science Day celebration at Penn Dixie in 2005.

counts among its membership residents from over 20 states, and Canada. Over 800 memberships, at a variety of levels, contribute to the daily operations of the HNHS and the Penn Dixie Site, along with increasing the HNHS endowment fund. Visitors from all over the U.S. and from Algeria, Australia, Canada, China, England, France, Germany, Israel, Italy, Japan, Lebanon, Mexico, New Zealand, Pakistan, Scotland, Spain, Sweden, and Switzerland have found the Penn Dixie Site a tremendous educational resource. **Penn Dixie was ranked No. 20 by attendance of the Top 25 Tourist Attractions in WNY in 2004 & 2005** by Business First of Buffalo, NY. The HNHS has a membership and corporate drive underway in 2006 to help attain a self-sustaining level in the near future.

OUR PLANS FOR GROWTH

The HNHS has established an ambitious capital fundraising campaign to ensure that the HNHS can continue to share with the public the many unique features of the Penn Dixie Site and meet the ever-increasing demand for the outdoor educational programming that it has been providing to the public over the past eleven years. The primary focus of the HNHS growth plan is the complete transformation of the Penn Dixie Site into a state-of-the-art educational facility in the natural sciences. The HNHS has taken a phased approach toward this goal and has:

- Removed the accumulated trash and debris from years of neglect and illegal dumping
- Completed a new entrance roadway and welcome sign to the Penn Dixie Site
- Drafted architectural drawings created for an outdoor education center on site
- Constructed an all-season parking area
- Installed over 2,100 feet of paved barrier-free trails
- Constructed registration and educational shelters
- Installed six picnic tables and two bike racks
- Cleared more than a mile of nature trail through a wooded portion of the site surrounding the eastern and northern sections of the former quarry

Future plans for the site include:

- Additional paved and boardwalk trails throughout the site so that it is entirely wheelchair and handicapped accessible
- Build an outdoor education center building for year-round programming
- Construct an astronomy pad with facilities for evening and day programs
- Enlarge and enhance the wetland areas
- Install additional information panels on fossil identification, site stratigraphy, geology, and biology for self-guided tours
- Pave the entrance roadway and parking areas
- Construct an amphitheater for outdoor programming
- Formalize classes and programming in the natural sciences for year-round on-site visitors
- Develop adult education and children's programming
- Make the education center available for community meetings.

The current plans for the outdoor education center include a community room for large lectures and meetings, a room for exhibits, a media center that will house a library and other teaching materials, a gift shop, classrooms, a research lab, and an observation deck to view the site. More than 80% of the square footage of the center will be devoted to education. A seismographic and climatological station are also planned to be included in the center.

HOW CAN YOU HELP

The HNHS has some ambitious plans to further develop this site into a world class outdoor educational, recreational, and tourist attraction for the Niagara Region. In completing the first phase, the HNHS has effectively preserved a unique educational and green space resource for future generations. With the completion of the next phases of development, the HNHS will maximize the educational opportunities afforded by the Penn Dixie Site for all of Western New York and the region. The Penn Dixie Site is already proving a powerful draw for visitors from all across North America and, indeed, the world. Completion of the site's educational facilities will only enhance this draw and bring increased numbers of visitors to the site. Programs and events were made possible by the support of over 320 volunteers in 2005, a 5% increase in the number of volunteers over 2004.

The difficult economic conditions in Western New York are now impacting the HNHS and the Penn Dixie Site. Decreases in support from Erie County, New York State, the Federal Government, and foundations have cut funding needed to implement our programs. The HNHS (a non-profit organization) needs to secure additional funding to keep our current level of public programs at Penn Dixie. The economic climate affecting non-profit organizations in our region has placed some unparalleled financial challenges on them. Many groups are requesting limited funds from foundations and government resources. The HNHS is attempting to raise funds by increasing memberships, admissions, programs, donations, grants, and seeking corporate support. The HNHS' goal is to become self-sustaining. Many members and donors, who have not even been to Penn Dixie, are willing to support our cause to preserve and develop this classic site for future generations.

You can help continue the tremendous advances and accomplishments that have been made to date by:

- Sending a donation.
- Taking out a membership.
- Recruiting a new HNHS member.
- Bringing visitors to Penn Dixie.
- Recruiting a Corporate member.
- Enrolling your family in a program or summer day camp.

The HNHS is actively seeking financial support from a variety of sources to attain its goal of transforming the Penn Dixie Site into an educational resource that fully utilizes and shares the unique resources contained within the site. If you are interested in learning more about how you can help support the HNHS and the Penn Dixie Site, please call the HNHS at 716/627-4560. Visit our web site www.penndixie.org for program and membership information. We look forward to having you visit Penn Dixie in 2006.

ACKNOWLEDGEMENTS

I thank my wife, Linda, for her review of this manuscript and Richard Spencer for taking the photographs and helping incorporate them into this document. Thanks also to Scott Clark who compiled the fossil plates of specimens that may be found at Penn Dixie and the Penn Dixie Location map. I also thank the HNHS Board, members, and volunteers who have unselfishly provided their time and talents to the preservation and development of the Penn Dixie Paleontological and Outdoor Education Center.

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ROAD LOG FOR PENN DIXIE SITE VISIT

Total Miles	Miles from last point	Route Description
		Depart the Adams Mark Hotel and turn right
0.0	0.1	Proceed to first stop sign and turn right onto Lower Terrace Street.
0.2	0.1	Proceed to Bingham Street (next right) and turn right
0.4	0.2	Proceed under Rt. 190 bridge to next stop sign, stay in left lane. Make left following Rt. 190 S & thruway Rt. 90 signs.
0.7	0.3	Stay in right hand lane and take Rt. 5 west over the Sky Way Bridge.
0.7	7.6	Continue on Rt. 5 through Lackawanna (note former Bethlehem Steel Plant on right) and continue through Lackawanna to Bay View Road (first left after Ford Plant). Turn left on Bay View Rd crossing over RT 5 East bound traffic.
8.3	1.0	Proceed on Bay View to intersection with Big Tree Rd, first stop sign. Turn left onto Big Tree Rd.
9.3	0.4	Proceed East on Big Tree Rd, at the sixth street turn left onto Bristol Rd.
9.7	0.3	Proceed North on Bristol Rd to North St, turn left.
10.0	0.1	Proceed West on North St to Penn Dixie entrance.

USING MARINE FOSSILS TO UNLOCK THE MIDDLE DEVONIAN PALEOENVIRONMENTS OF WESTERN NEW YORK
(FOR K-12 TEACHERS AND COLLECTORS)

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INTRODUCTION

The physiographic province of the Allegheny Plateau offers a geologically rich visage of the Paleozoic prehistory of western New York State. The region possesses a relatively complete (and highly fossiliferous) section of Middle Devonian rocks. This section of Devonian strata thickens towards the east, directly reflecting the resultant sedimentation from the Acadian orogeny, and inspiring the name of the Catskill clastic wedge to the strata (Brett, 1986). As mountain building progressed to the east and southeast, most of New York State was covered by a shallow sea in what is known as the Appalachian Basin (Isachsen et al., 2000). During this time, the amount of sediments being deposited changed, the type of sediments changed, and local sea levels rose and fell. These paleoenvironmental fluctuations set the stage for major changes in the regional paleontology during the Middle Devonian.

This paper and the accompanying field trip seek to introduce science educators and fossil collectors to some interesting Devonian sites in western New York (Figure 1). The trip explores rocks chiefly from the Middle Devonian Hamilton Group, which represents a 5 to 7 million year time period between about 377 and 384 million years ago (Brett, 1986). The Hamilton Group includes four formations: the Marcellus, Skaneateles, Ludlowville, and Moscow. Formations within the Hamilton group are divided by three separate limestone units: the Stafford, Centerfield, and Menteth/Portland Point (Patchen and Dugolinsky, 1979; Brett, 1986). The Genesee Formation, which in some places overlies the Hamilton Group, is another important unit in understanding the regional geologic setting (Brett and Baird, 1994). The sequence of Devonian depositional events along with data from fossil assemblages helps us to characterize the "big picture" of the local geological history.

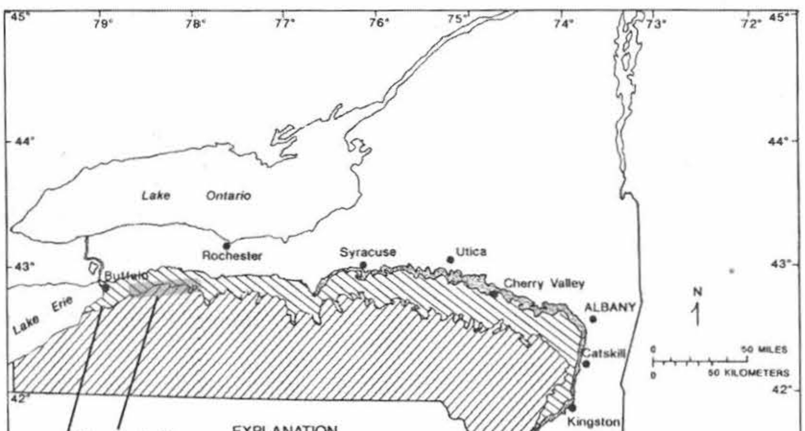


Figure 1. Location of Devonian strata in New York State with field trip area. Adapted from Isachsen, et al., 2000.

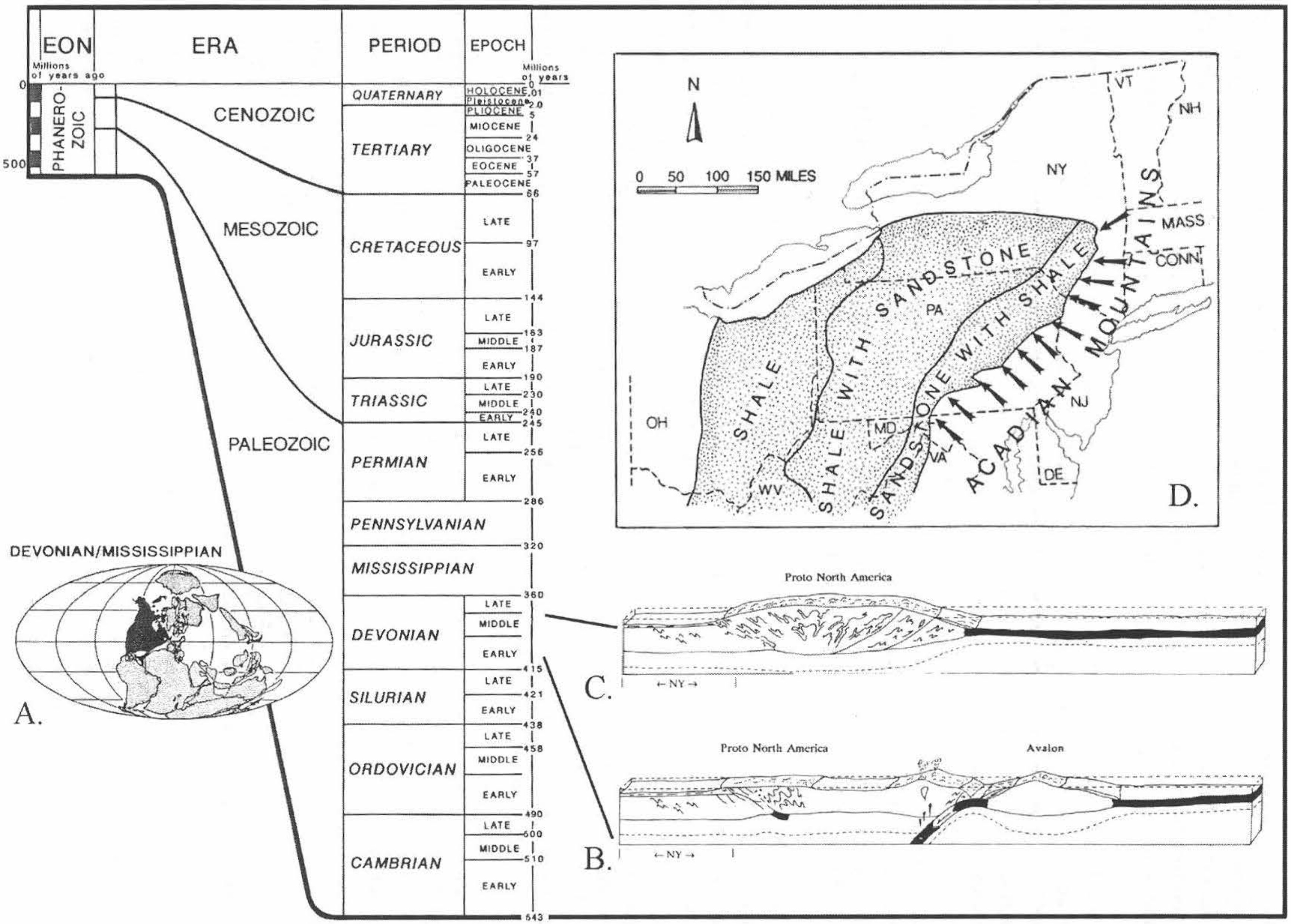


Figure 2. See explanation in section below. Modified from Isachsen, et al., 2000 and NYSE Earth Science References, 1999.

NEW YORK STATE DURING THE DEVONIAN

Early Devonian times saw the eastern shelf of the proto-North American continent covered by the Iapetus Ocean. New York State was also underneath a broad, shallow sea. The presence of thin limestone beds in central and eastern New York indicates that the landscape was relatively flat and free from sediment influx. Corals, bryozoans, crinoids, sponges, brachiopods, trilobites, and other marine inhabitants flourished during this time (Isachsen et al., 2000).

Sea levels dropped near the end of the Early Devonian, causing an erosional unconformity to form across much of New York. By about 390 million years ago, a shallow sea had returned to New York. Coral reefs and primitive fish dominated the paleoecology. The Onondaga Limestone represents a depositional period during the final hurrah of the placid marine environment. During this time several micro-continents were proximal to proto-North America, and closing (Figure 2A). The Early Devonian marked the shrinking of the Iapetus Ocean as its tectonic plate was being subducted beneath northeastern proto-North America (Figure 2B). About 390 million years ago, micro-continents known as the Avalon terranes crashed into proto-North America, resulting in the Acadian Orogeny (mountain building event). The collision lasted through the Middle Devonian (Isachsen et al., 2000).

By 380 million years ago, Avalon had "fused" to the North American craton (Figure 2C). Although parts of New York were slightly deformed by the collision, other parts of proto-North America experienced much greater deformation. The Acadian Orogeny had its greatest effects on present-day New England. Tall and extensive mountains formed to the east and southeast of New York (Figure 2D). The "rugged, lofty" Acadian mountains served as an excelled source of sediment for fluvial erosion (Isachsen et al., 2000). Rivers flowing from the Acadians deposited sands, silts, and clays over the shallow sea covering New York (arrows in Figure 2D). Thicker sections of sediment accumulated as the seafloor sunk (subsided). The "wedge" of sedimentary rock deposited during erosion of the Acadian Mountains is informally known as the "Catskill Delta" (Isachsen et al., 2000).

Fluctuations in local depositional conditions led to changing types of sedimentation throughout the Middle Devonian. The sedimentary record shows that the shorelines of the "Catskill Delta" of New York changed over time (Figure 3). This was due to several circumstances. One contributing factor was that the crust beneath the sea floor continually sunk as more sediment was loaded into the basin. Climate changes (such as increased/decreased precipitation) could have also played a role in the erosion rates of the Acadian Mountains. Also, the amount and types of sediment being eroded would have had a large influence on depositional conditions. All of these factors influenced a dynamic "Catskill Delta" shoreline during the Middle and Late Devonian (Miller, 1986; Isachsen et al., 2000).

Western New York State was relatively far from the "Catskill Delta" sediment source, and many of the Middle Devonian rocks reflect this relationship. The Hamilton Group contains four formations, each separated by a thin limestone bed. The Genesee Formation lies directly above the Hamilton Group and will also be explored in this trip (Figure 4). Rocks from these strata are dominantly shales and siltstones, with some thin limestones and coarse clastic beds (Brett, 1986; Brett et al., 1986; Brett and Baird, 1994; Isachsen et al., 2000).

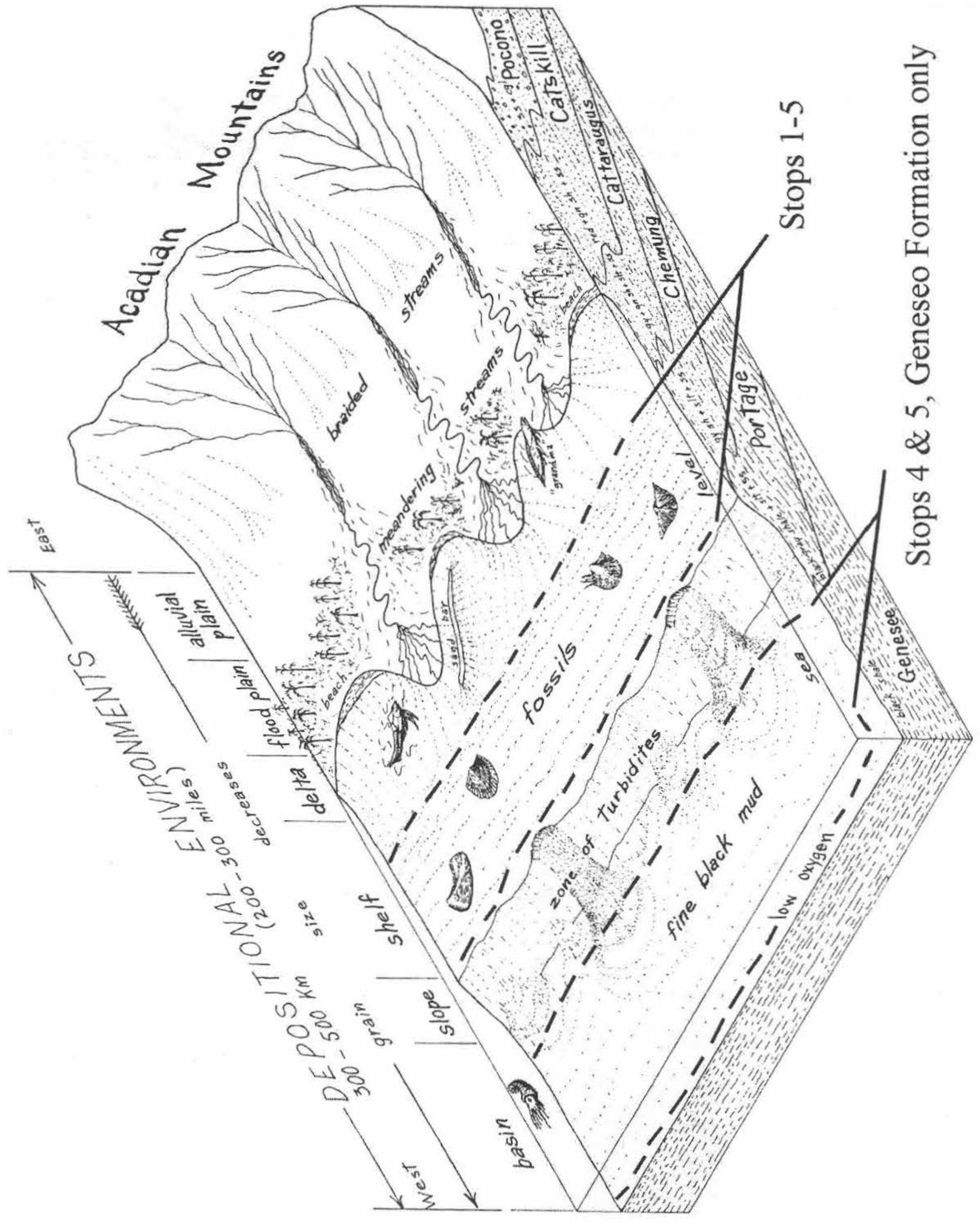


Figure 3. Characterization of "Catskill Delta" depositional settings. As deposition progresses outwards from the source of sediment, the physical energy of transport forces decreases, resulting in smaller clasts being carried. In this model, the relatively high energy streams, beaches, and deltas carry and deposit cobbles, pebbles, and sands; the medium energy offshore currents along the proximal shelf deposit sands; the lower energy currents along the distal shelf and slope deposit silts and clays; and the lowest energy currents in the basin deposit clays with some silts. Field trip sites are marked in their respective paleoenvironmental (but not necessarily geographical) provinces. Adapted from Isachsen et al., 2000.

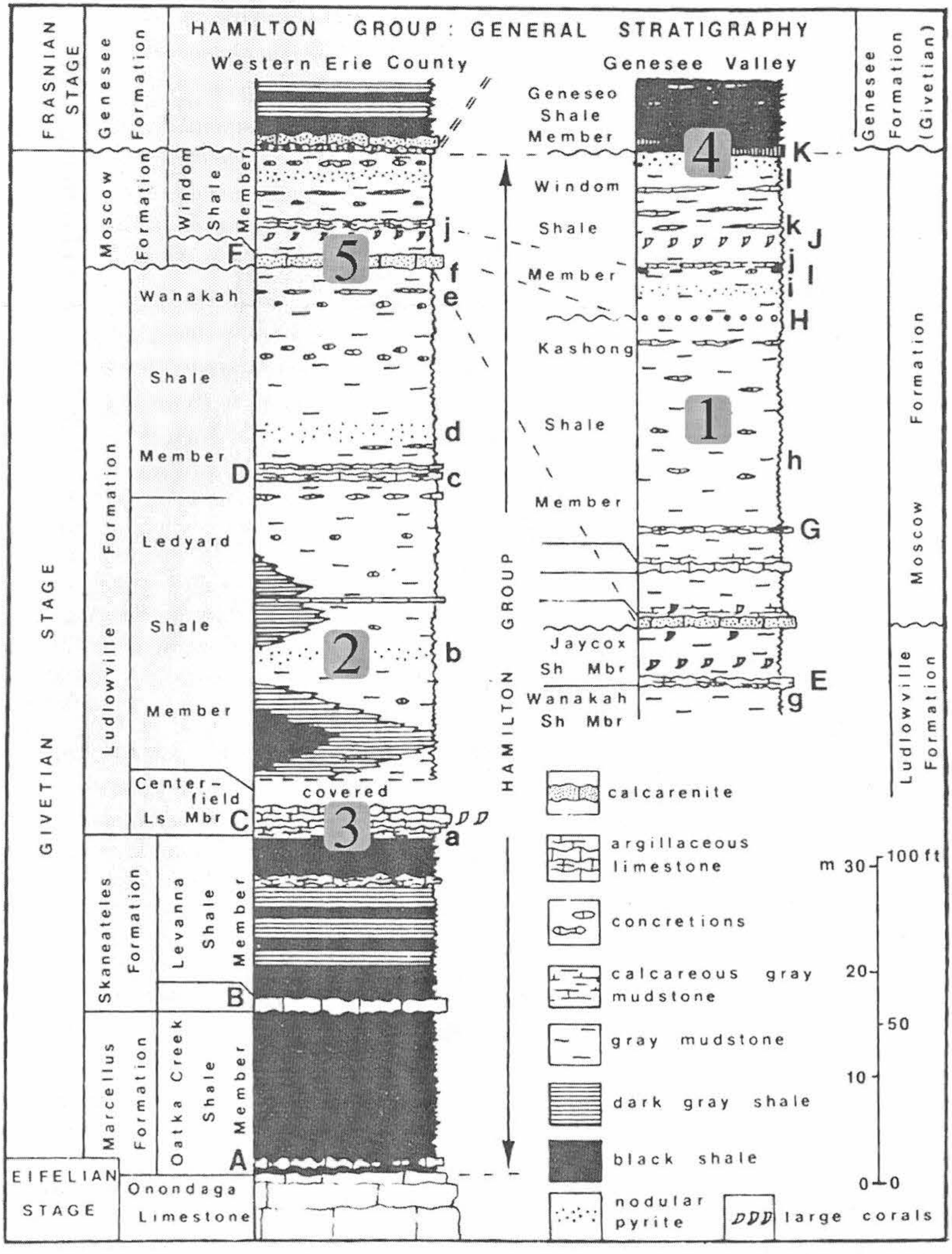


Figure 4. General stratigraphy of the Hamilton Group in western New York. Field trip locations are indicated in relation to their stratigraphic position. Important beds in this field trip include the: F) Tichenor Limestone; G) Menteth Limestone; K) Leicester Pyrite; and b) Alden Pyrite. Modified from Brett et al., 1986.

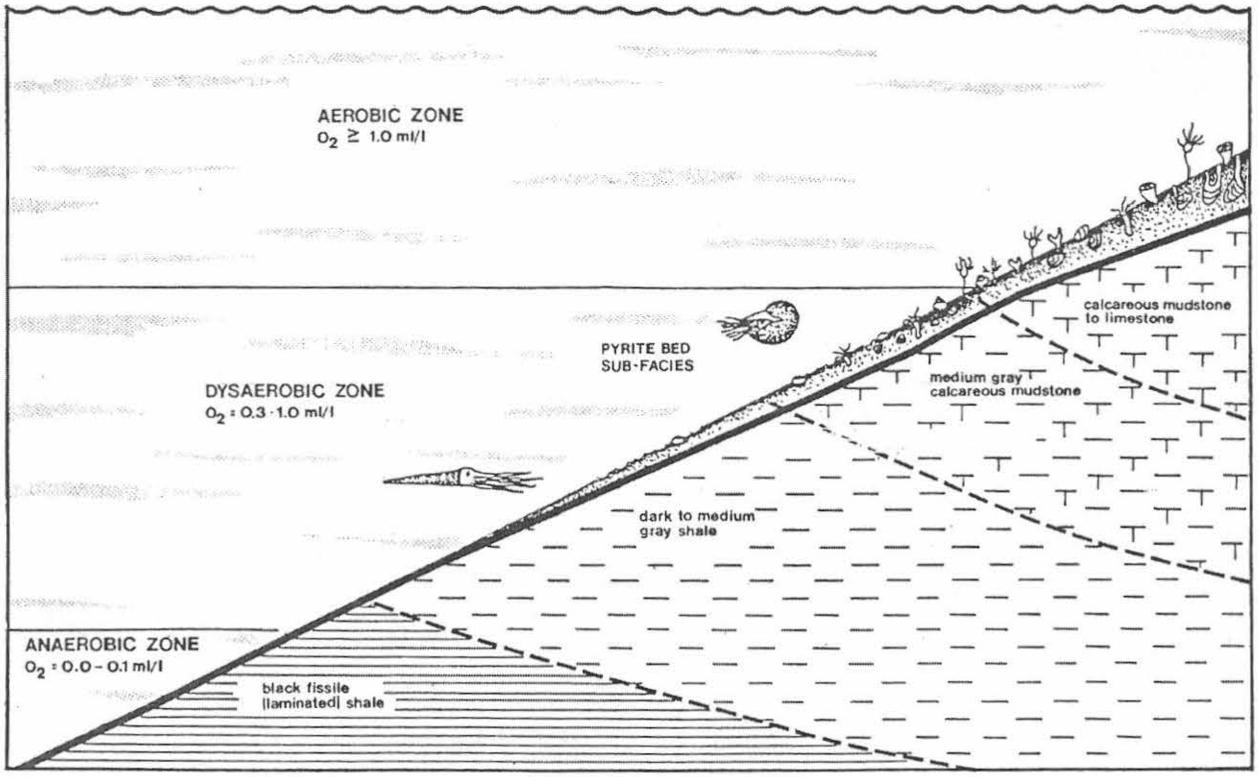


Figure 5. Oxygenation zones in Middle Devonian seas of New York. Generally, for the rocks on this trip, black shales are associated with the anaerobic zone (Facies I), gray shales with the dysaerobic zone (Facies II), and calcareous (limey) mudstones and limestones with the aerobic zone (Facies III/IV). Because of the lack of oxygen in the anaerobic zone, few marine organisms are able to exist. For those who live and die in this zone, the lack of oxygen prevents their immediate decomposition. The remaining organic matter can gradually become trapped hydrocarbons. The dysaerobic zone has a relatively limited supply of dissolved oxygen, and so few crinoids, corals, and bryozoans live in this zone. However, mobile animals such as fish and cephalopods flourish in the water above the dysoxic sea floor, as do some brachiopods and gastropods. Aerobic zones usually have a much greater diversity of life than other marine zones, and as a result the depositional environment tends to include a significant amount of calcium carbonate. For purposes of discussion within this paper, dysaerobic is synonymous with dysoxic, and anaerobic is synonymous with anoxic. Figure modified from Brett et al., 1991.

STOP 1: THE KASHONG SHALE

The Kashong Shale Member of the Moscow Formation is described as a bluish-gray mudstone (Brett and Baird, 1994). Deposition occurred in mostly shallow waters under aerobic conditions (Figure 5). At its thickest, the Kashong Shale is about 25 meters (80 feet) in the Genesee Valley (Figure 6). At Stop 1 (Retsof Locality), only the Middle Kashong is exposed. Common fossils from this unit are *Tropidoleptus carinatus*, *Nucleospira concinna*, *Mucrospirifer*, and *Athyris* (brachiopods); *Pleurodictyum americanum* (the patriotic tabulate coral); *Orthonota undulate* (bivalve); and *Dipleura* (trilobite). Many species of crinoids, blastoids, and bryozoans can also be found at this site. The bryozoans at this locality frequently encrust larger brachiopods and the numerous crinoids. Because bryozoans and crinoids need hardgrounds for larval attachment, this observation indicates that there was relatively little sediment input into the depositional system. Thus, the filter-feeders were forced to grow on other bottom-dwelling organisms (Brett and Baird, 1994).

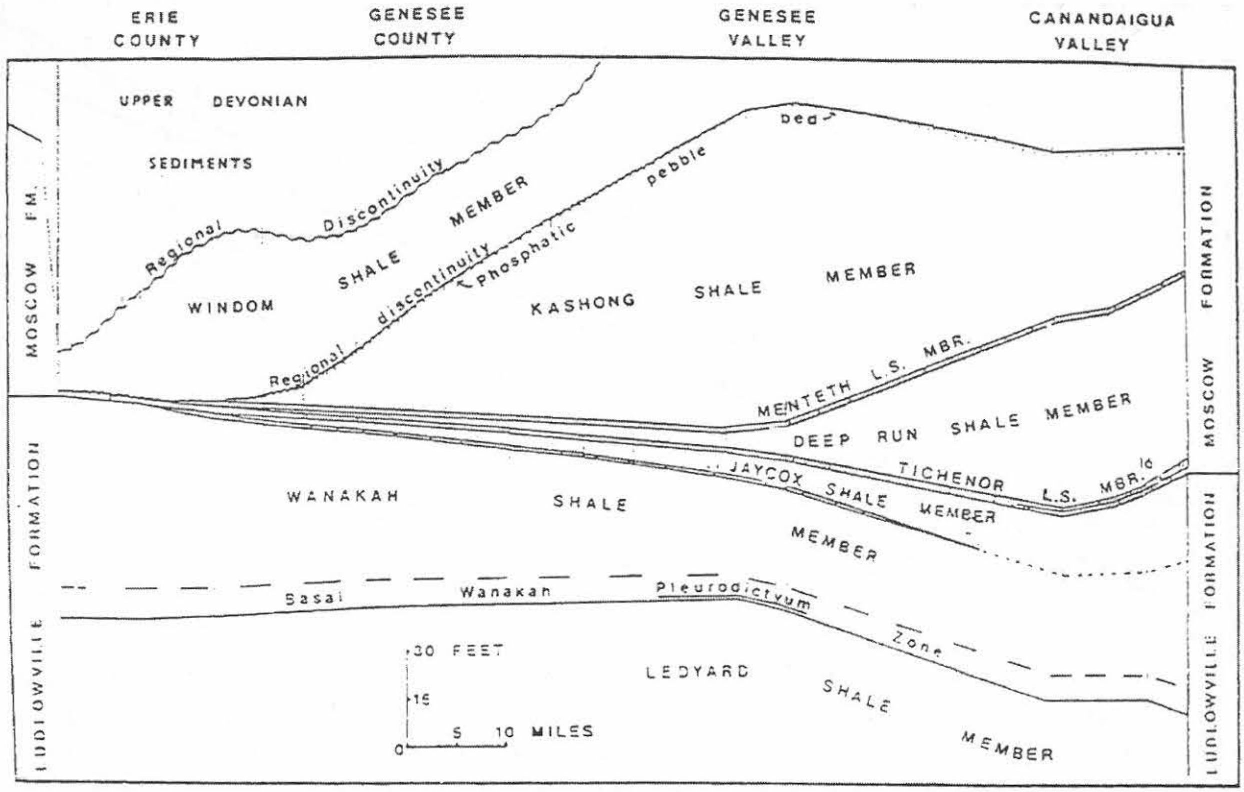


Figure 6. Cross-section of Ludlowville and Moscow Formations in western New York. Modified from Brett and Baird, 1994.

STOP 2: ALDEN PYRITE BED OF THE LEDYARD SHALE

The Ledyard Shale Member of the Ludlowville Formation is described as containing shales and argillaceous (clayey) limestones (Wygant 1986). The Ledyard Shale was deposited in deeper water than the Kashong Shale. Wygant (1986) classifies the Ledyard as a dysoxic sulfidic environment, meaning that there was a relative abundance of iron sulfide in relation to oxygen. These conditions led to the widespread deposition of pyritic beds in the Middle Devonian throughout western New York (Dick and Brett, 1986). Deceased organisms in these zones decomposed under low-oxygen conditions, leading to chemical processes that caused their hard parts to be replaced by iron pyrite. The Alden Pyrite Bed contains an abundance of pyritized trace and body fossils. Among the more common are brachiopods, bryozoans, corals, crinoids, bivalves, and trilobites (Wygant, 1986). Common species from this locality include *Phacops rana* (trilobite); *Tornoceras* (ammonite); and *Lingula*, *Mucrospirifer*, and *Devonochonetes* (brachiopods). Although there are many fossils to be found, the overall biodiversity of the Alden Pyrite is quite limited. Since the pyrite layer is located in a dysoxic zone (Figure 7), the faunal assemblage is not as diverse as it would be in a more oxygenated environment (Kloc, 1983; Dick and Brett, 1986). The local relationship of the Alden Pyrite Bed within the Ledyard Shale is shown in Figure 8.

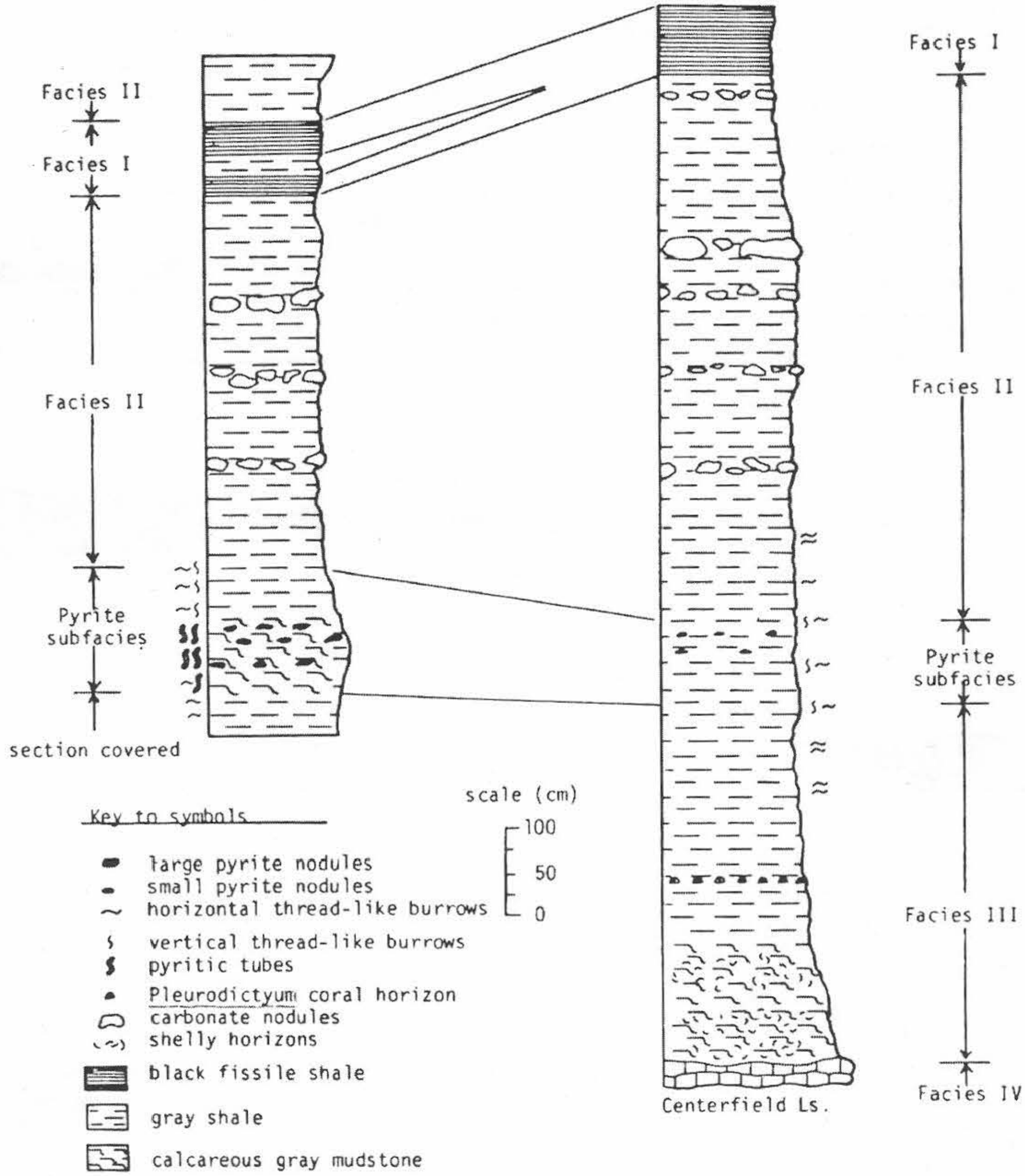


Figure 7. Stratigraphic sections showing two localities of the Alden Pyrite proximal to Stop 2. Adapted from Dick and Brett, 1986.

STOP 3: THE CENTERFIELD LIMESTONE

The Centerfield Member of the Ludlowville Formation is a "shale-limestone-shale" sequence similar to several other units within the Hamilton Group (Savarese et al., 1986). This trip visits the Centerfield limestone, which is sandwiched between upper and lower Centerfield shale units (Figures 8 and 9). Centerfield shales are typically light gray to gray-colored and calcareous, while the Centerfield limestones are approximately 70-80% calcium carbonate with some argillaceous (clay) content (Savarese et al., 1986). Stop 3 on this trip has exposures of weathered outcrop from the middle limestone, with over- and underlying units covered by

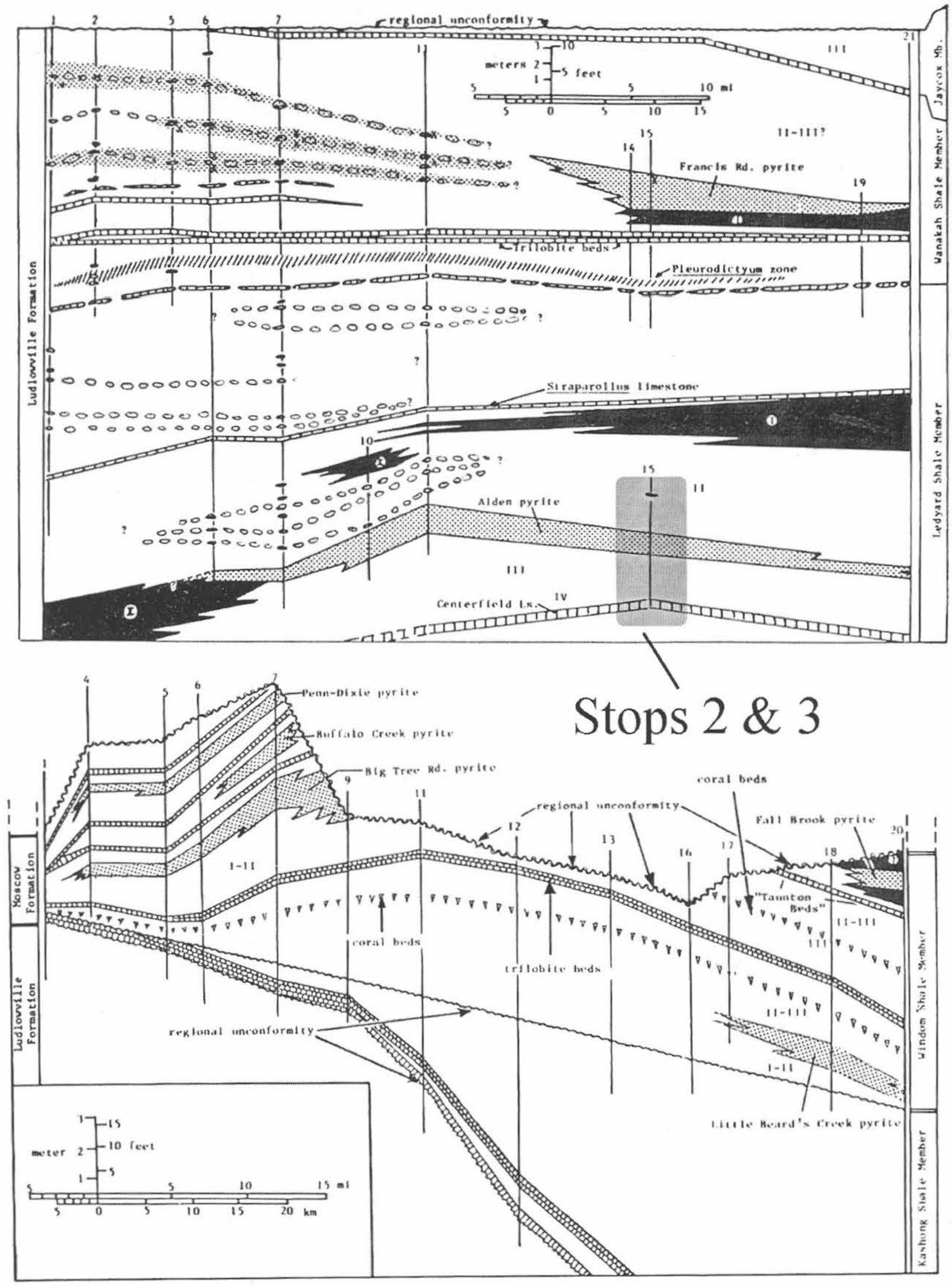


Figure 8. General stratigraphic sections for Stops 2 and 3. Adapted from Dick and Brett, 1986.

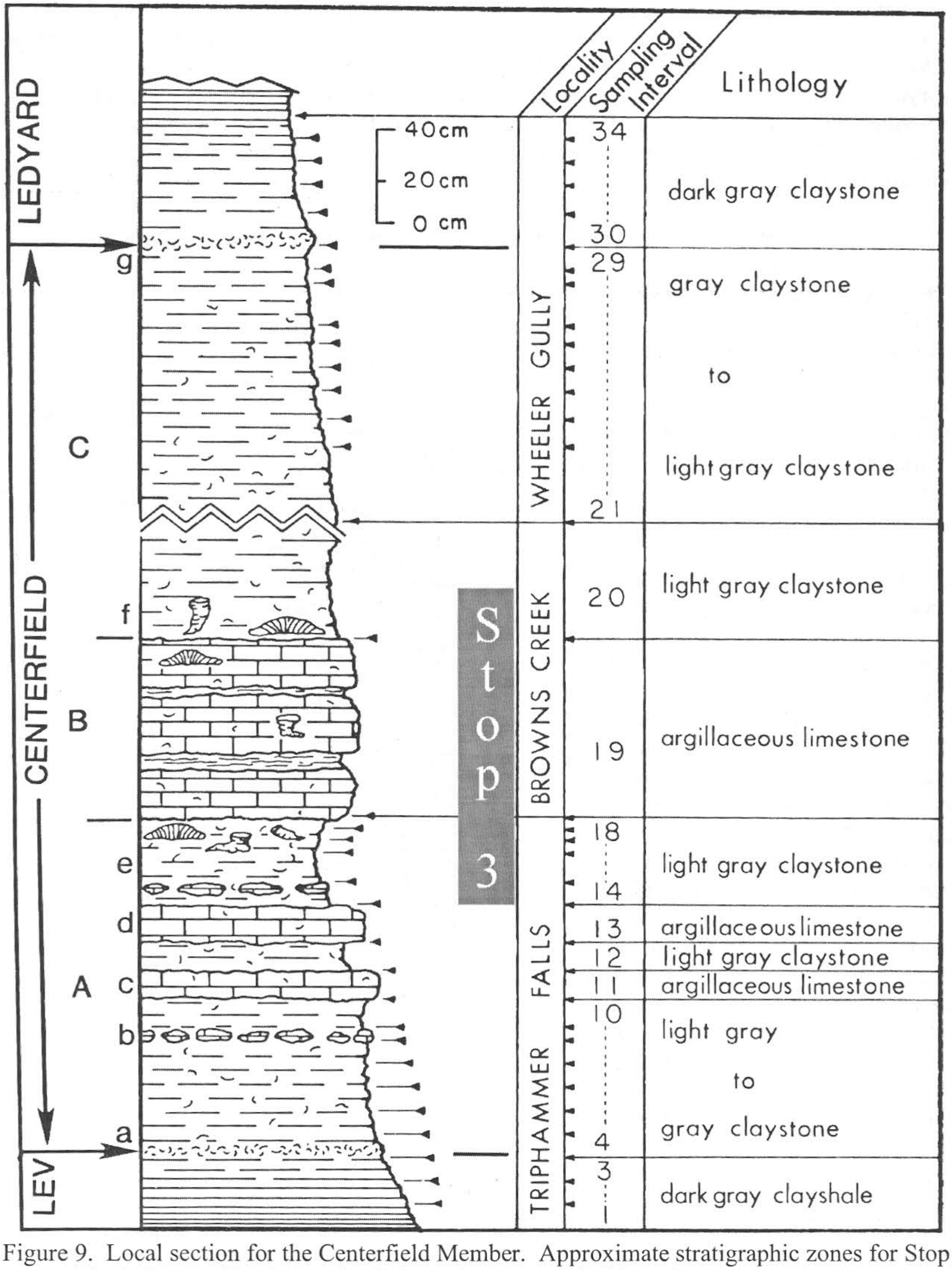


Figure 9. Local section for the Centerfield Member. Approximate stratigraphic zones for Stop 3 include sampling intervals 14 through 20. Modified from Savarese et al., 1986.

vegetation and slope wash. Common fossils at this stop include *Ambocoelia umbonata*, *Longispina mucronatus*, *Athyris*, *Mediospirifer*, and *Nucleospira* (brachiopods); *Heliophyllum*, *Cladopora*, *Botherophyllum*, and *Favosites* (corals); *Fenestella eumaciata* and *Loculipora* (bryozoans); *Actinopteria* and *Cypricardinia* (bivalves); *Platyceras* (gastropod); and the trilobites *Phacops* and *Greenops* (Savarese et al., 1986). Because of the diversity of fossil fauna at this site, the Centerfield limestone is interpreted as an aerobic depositional environment.

STOP 4: THE LOWER BLACK SHALES OF CAYUGA CREEK

The Stop 4 exposure at Cayuga Creek reflects a much different paleoenvironment from the previous sites. Towards the end of the Middle Devonian, the Appalachian Basin began to deepen in western New York. As a result, the type of submarine deposition changed from an aerobic to anoxic environment (Brett and Baird, 1994; Isachsen, 2000). The Windom Shale at this location is very dark gray to black colored and is thinly bedded. Narrow lenses of the Leicester Pyrite occur at the unconformable interface between the Windom Shale and the overlying Geneseo Shale (Figure 4). Like the Alden Pyrite, the Leicester Pyrite bed contains small brachiopods, bivalves, ammonoids, and trace fossils (Dick and Brett, 1986; Brett and Baird, 1994). The Geneseo Shale is the lowermost member of the Genesee Formation and represents an anoxic deepwater environment. A few thin layers of black shale with chonetid brachiopods likely represent brief environmental changes into dysoxic conditions. Also, the rounded concretions in this unit may have assemblages of conodonts and other microfossils, but overall the Genesee Black Shale is nonfossiliferous (Brett and Baird, 1994).

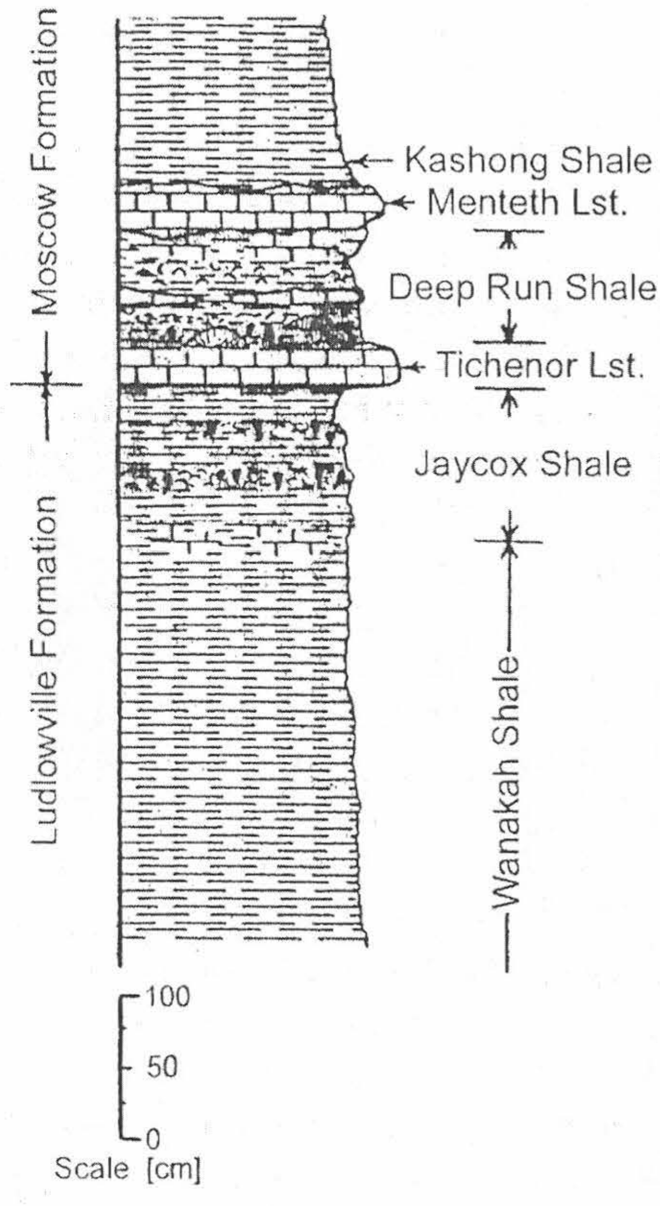


Figure 10. Partial Buffalo Creek stratigraphy. From Maletz, 2006.

STOP 5: THE LOWER WINDOM SHALE AT BUFFALO CREEK

Although deep water environments do not usually have the fossil diversity of shallow water environments, the abundance of fossils isn't necessarily affected. The lowest portion of the Windom Shale Member of the Moscow Formation represents submarine deposition in dysoxic conditions. Unlike the Alden and Leicester Pyrite beds, the lower Windom is not indicative of a highly sulfidic paleoenvironment. The maximum thickness of the Windom Shale is approximately 60 meters (200 feet), but in western New York its thickness is closer to 15 meters (50 feet). Shales and mudstones encompass most of the Windom, with a few thin bands of limestone and concretions scattered throughout (Brett and Baird, 1994). Since it is stratigraphically higher (and younger in age), the bottom of the Windom Shale outcrops upstream of the Tichenor and Menteth Limestones, with Kashong Shale (Figures 10 and 11).

The lower part of the Windom Shale consists of several unique beds. A phosphate bed exists along the bottom Windom contact with the Kashong Shale. This bed consists of a bioturbated gray mudstone with interbedded pebbles (Brett and Baird, 1994). The thin phosphate layer transitions into "soft medium gray" shale. This shale is identified by the presence of the common brachiopod *Ambocoelia umbonata*. Other abundant fossils from this bed include chonetid brachiopods (e.g. *Devonochonetes scitulus*), rugose corals, and the trilobite *Phacops rana*. At Buffalo Creek, the *Ambocoelia* beds can be used to identify the Lower Windom Shale and locate other adjacent sedimentary units (Figure 11; Baird and Brett, 1983; Brett and Baird, 1994).

SUMMARY

Depositional settings in western New York changed many times throughout the Middle Devonian. Sediment runoff from the eroding Acadian Mountains and variations in sea level were directly responsible for many of these changes. The Appalachian Basin deepened and shallowed, sediment supply changed, and the fossil assemblages shifted to accommodate these active processes. Each site on this trip was selected to give the participant multiple glimpses into the variation of the distant past. Through identifying fossils and understanding the conditions by which they were deposited, it should be possible for grade school students to reconstruct the paleoenvironment of the Hamilton Group.

ACKNOWLEDGEMENTS & NOTES

The authors would like to thank Dr. Richard S. Laub from the Buffalo Museum of Science for his assistance in identifying corals and other specimens from several of the sites on this trip. The authors also acknowledge Dr. Gordon C. Baird for his help with recognizing the Retsof locality and its position in the Kashong stratigraphy. Some components of this project were funded by the National Science Foundation through the Buffalo Geosciences Program at the University at Buffalo (www.bgp.buffalo.edu). Special thanks to Kristin A. Sturtevant for review of the text and figures, and to Kathryn M. Whalen for assistance in road logging. The authors credit Dr. Robert D. Jacobi from the University at Buffalo for his trip suggestions and subsequent guidance throughout the process of drafting this guide.

See the Appendix for images of some of the more common fossils on this trip. Figure 14 is a special fossil cartoon activity for children, designed by the Buffalo Geosciences Program.

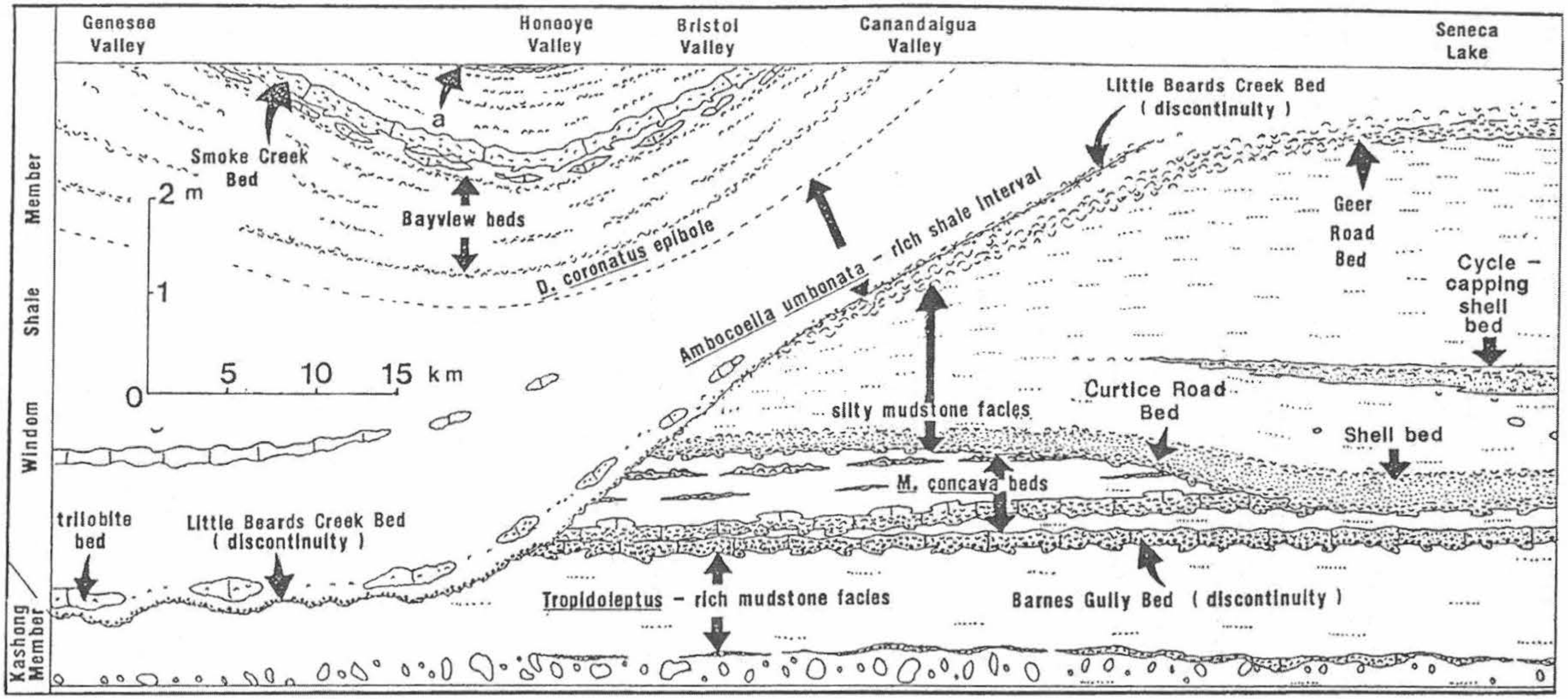


Figure 11. Regional deposits of the Late Middle Devonian. Of particular interest here is the "*Ambocoelia* umbonata - rich shale interval," or simply the "*Ambocoelia* beds" of the Lower Windom Shale. Modified from Brett and Baird, 1994.

APPENDIX

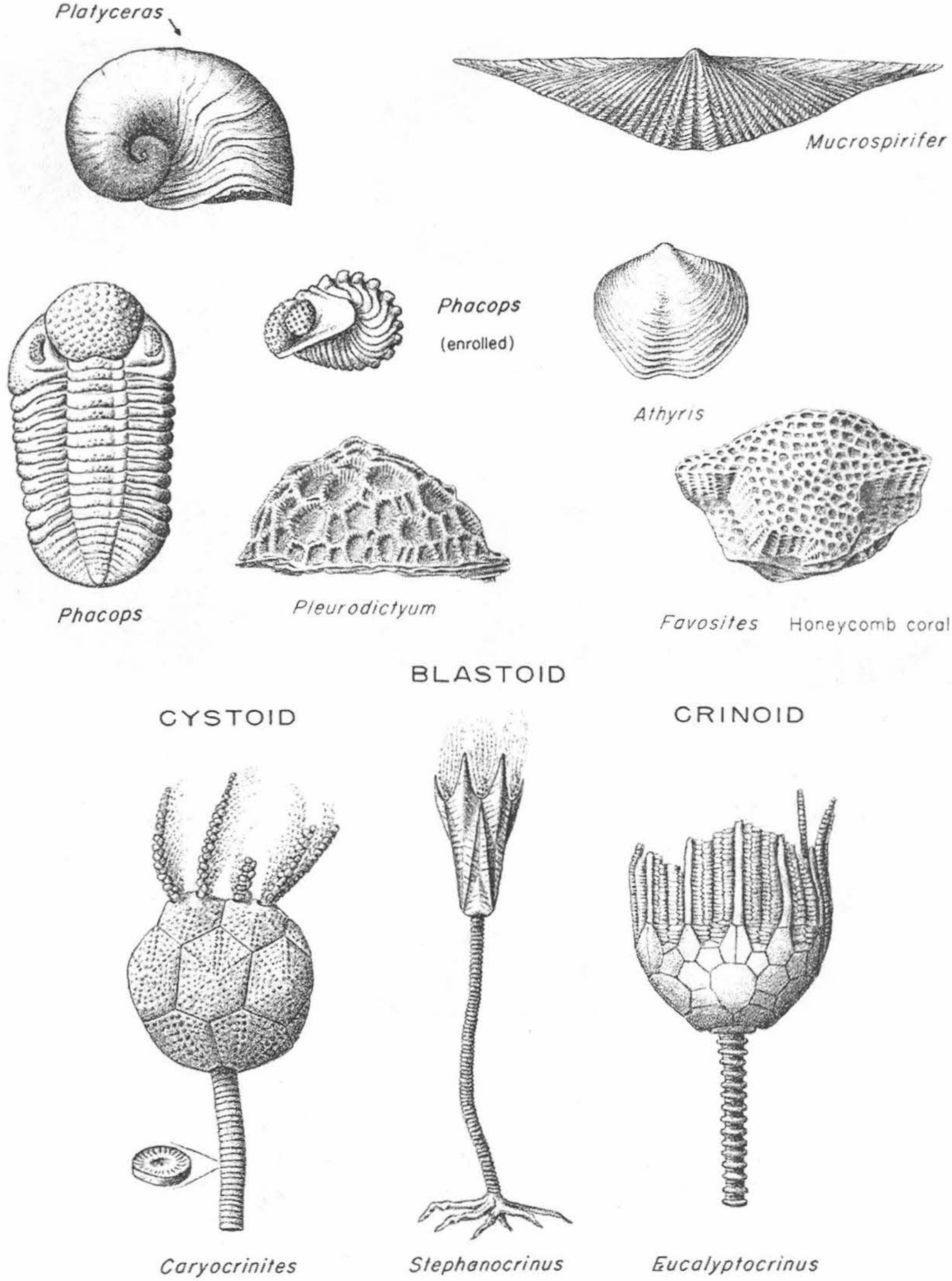
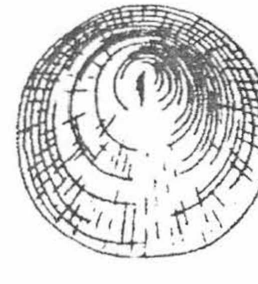
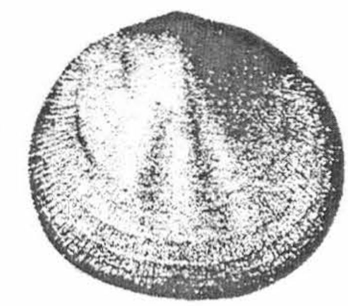


Figure 12. Miscellaneous fossils from trip localities. Modified from Isachsen et al., 2000.

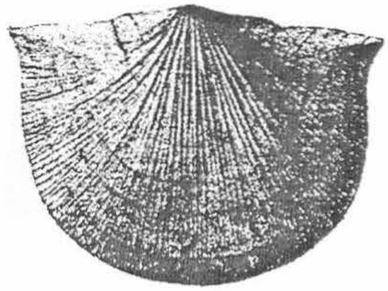
COMMON DEVONIAN BRACHIOPODS



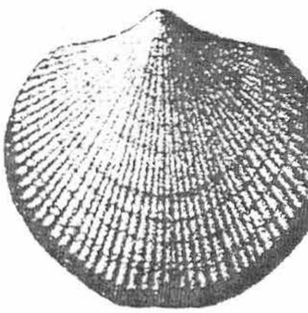
Orbiculiodea sp.



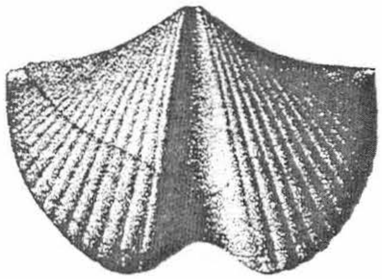
Rhipidomella sp.



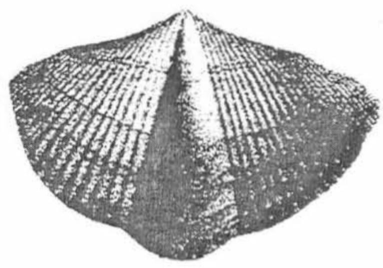
Stropheodonta demissa



Pseudoatrypa devonica



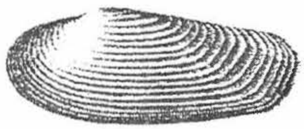
Spinocyrtia granulosa



Mediospirider auduculus

COMMON PELECYPODS (BIVALVES)

Pterinopecten sp.



Palaeoneilo sp.

Plethomytilus sp.

Figure 13. Common brachiopods and pelecypods (bivalves) from the Middle and Upper Devonian rocks found on this field trip. These fossils are particularly abundant in the Wanakah and Windom Shales, and will be typically found at Stops 5 and 6. Based on figures in Bastedo, 1999 and from original drawings by Grabau, 1898-99.

Add your own words to this zany scene from the Paleozoic!

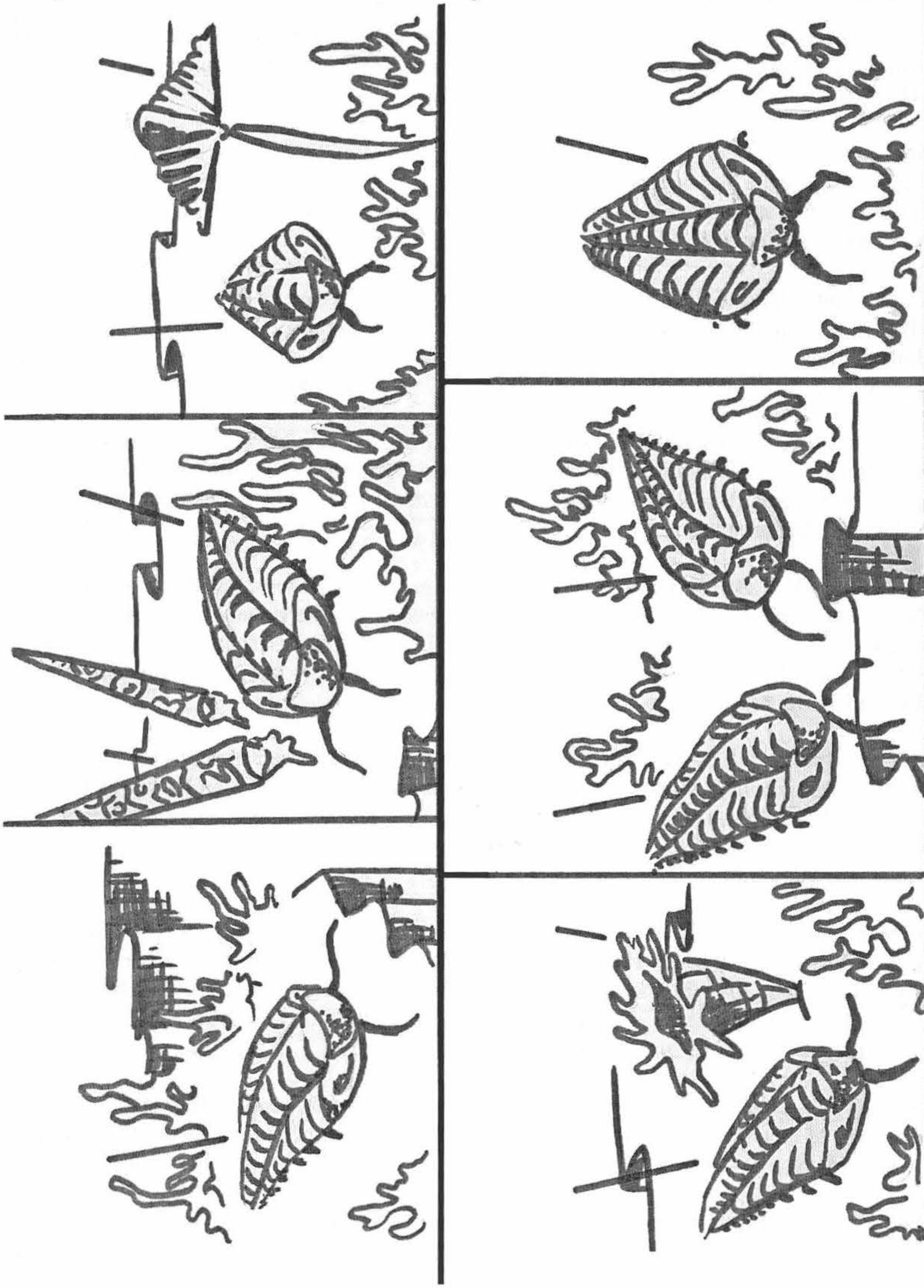


Figure 14. Devonian fossils cartoon drawn by Archana Jayakumar.

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FIELD TRIP ROADLOG

Total	Distance	Directions	Location
0.0	0.0	Head East on Church St. toward Lower Terrace St.	Adam's Mark Hotel, 120 Church St., Buffalo, NY
0.2	0.2	Church St. becomes S. Division St.	S. Division St.
0.4	0.2	Left onto Elm St./NY-33 E	NY-33 E
8.0	7.6	Exit I-290 toward I-90 E/Albany	I-290/I-90 E
39.6	31.6	Take exit 48 (Rt. 98) toward Batavia	Onramp for I-90 E
40.1	0.5	Left onto NY-98 S	Intersection after tollbooth
41.2	1.1	Left onto W Main St./NY-33/ NY-5/NY-63 S	City of Batavia
41.5	0.3	Right onto NY-63 S (Rt. 63)	City of Batavia
53.3	11.8		Town of Pavilion
58.8	5.5		Town of York
62.2	3.4	Left into parking lot near Bank of Castile/Dollar General, **STOP 1**	Off Rt. 63 near intersection of Rt. 63 and Retsof Rd.(Greigsville)

<u>STOP 1.</u> Retsof Locality (of older literature). Cross Rt. 63 at the entrance to the parking lot. Proceed down the gravel trail (westward) towards and just beyond the Rt. 63 Retsof Switch Structure. Head south about 250-300 meters until the flat bedding surface turns into outcrop near old railroad yard. If vegetation is high, it may be difficult to find the trail. The best collecting from this site is found along the bluff wall adjacent to the rail yard, although surface collecting is also possible in some places.

Geologic unit: Kashong Shale Member of the Moscow Formation, Hamilton Group
Geologic age: Middle Devonian
Common fossils: Crinoids, brachiopods, corals, bryozoans

Total	Distance	Directions	Location
62.2	0.0	Right out of parking lot	Rt. 63
71.5	9.3		Drumlin-like landform
79.1	7.6	Left onto Bethany Center Rd.	Intersection of Rt. 63 and Bethany Center Rd., Town of East Bethany
81.2	2.1	Park on gravel shoulder on right	Bethany Center Rd. near shale/mud pit approx. 0.4 mi past Paul Rd.

<u>STOP 2.</u> Alden Pyrite Beds (east-facing shale pit on Bethany Center Road). Exit vehicle and walk approximately 150 meters west into the clayey hills to the right (west) of the road. Pyrite beds outcrop from the weathered bedrock surfaces near recently incised channels throughout the site. Calcareous concretions eroding from the outcrop often contain brachiopods and cephalopods. Pyritized fossils also appear on muddy flats below weathered outcrop.

Geologic unit: Ledyard Shale Member of the Ludlowville Formation, Hamilton Group
Geologic age: Middle Devonian
Common fossils: Pyritized brachiopods, trilobites, rugose corals, cephalopods, gastropods

<u>STOP 3.</u> Delaware, Lackawanna, and Western Railroad Tracks. From Stop 2, head back to the road and walk north about 200 meters, past the guard rails on Bethany Center Rd. and descend left (west) on gravel path into the railroad bed. Abundant fossils can be found both on the surface and in weathered bedrock fragments. Possible lunch stop.

Geologic unit: Centerfield Limestone Member of the Ludlowville Formation, Hamilton Group
Geologic age: Middle Devonian
Common fossils: Rugose and tabulate corals, bryozoans, brachiopods, trilobites, crinoids

Total	Distance	Directions	Location
81.2	0.0	Continue south	Bethany Center Rd.
83.0	1.8	Right onto Rt. 63 S toward	Intersection of Bethany Center Rd. and Rt. 20 W connection road
83.1	0.1	Right onto Rt. 20 W toward Darien Center	Intersection of Rt. 20 connection/ Old Telephone Rd. and Rt. 20
89.4	6.3		Village of Alexander

94.2	4.8		Town of Darien
96.4	2.2	Left onto Rt. 77 S toward Bennington	Intersection of Rt. 20 and Rt. 77 Welcome to Darien Center!
100.9	4.5	Right onto Rt. 354 W	Intersection of Rt. 77 and Rt. 354
101.0	0.1	Right onto Rt. 354 W/Clinton St.	Town of Bennington
104.7	3.7		Cowlesville, NY
105.2	0.5		Town of Marilla
107.0	1.8	Left onto Eastwood Rd, park on right shoulder, **STOP 4**	Intersection of Rt. 354/Clinton St. and Eastwood Rd., Marilla

STOP 4. Cayuga Creek. Cross Eastwood Rd. and descend to the creek via the dirt trail alongside the overpass. Take note of the bedrock on which the overpass trusses are built. The black shales of the lower Geneseo are nearly void of fossils at this location except for sporadic pavements of chonetid brachiopods. Lenses of the Leicester Pyrite contain pyritized burrows and fossil steinkerns.

Geologic units: Windom Shale Member, Moscow Formation, Hamilton Group; Leicester Pyrite (in lenses along unconformity), Genesee Formation; and Geneseo Shale Member, Genesee Formation.
Geologic age: Middle and Upper Devonian
Common fossils: No common fossils, sparse brachiopods

Total	Distance	Directions	Location
107.0	0.0	Turn around and head North	Eastwood Rd. near Cayuga Creek
107.0	0.0	Left onto Rt. 354/Clinton St.	Intersection of Eastwood Rd. and Rt. 354/Clinton St.
109.6	2.6	Left onto Two Rod Rd.	Intersection of Rt. 354/Clinton St. and Two Rod Rd.
110.6	1.0	Right onto Bullis Rd.	Intersection of Two Rod Rd. and Bullis Rd., Town of Marilla
112.8	2.2	Left onto Buffalo Creek Bridge Ramp (Between Stolle Rd. and Girdle Rd.), **STOP 5**	Intersection of Bullis Rd. and Buffalo Creek Bridge ramp

STOP 5. Buffalo Creek. After parking, descend down ramp trail and cross old Buffalo Creek bridge. Proceed down one of several trails to the creek. Approximately 300-400 meters downstream from the bridge, cross the creek in a shallow place to get access to the vast exposures of bedrock. The Lower Windom Shale is particularly accessible on the north bank of the creek upstream of the old Bullis Rd. bridge.

Geologic units: Wanakah Shale and Jaycox Shale Members, Ludlowville Formation, Hamilton Group; Tichenor Limestone, Deep Run Shale Member, Meneth Limestone, Kashong Shale Member, and Windom Shale Member, Moscow Formation, Hamilton Group; and Geneseo Shale, Genesee Formation.
Geologic age: Middle and Upper Devonian
Common fossils: Trilobites, brachiopods, corals, bryozoans

PENN DIXIE (OPTIONAL STOP)

Total	Distance	Directions	Location
112.8	0.0	Turn around, left onto Bullis Rd.	Intersection of Buffalo Creek Bridge ramp and Bullis Rd.
117.8	5.0	Left onto Transit Rd./Rt. 78 S	Intersection of Bullis Rd. and Transit Rd./Rt. 78
117.9	0.1	Right onto Rt. 400 N	Onramp for Rt. 400 N
122.7	4.8	Exit Rt. 400 to I-90 W	Onramp for I-90 W
124.7	2.0	Take exit 56 toward Blasdell	I-90 W
125.2	0.5	Right onto Rt. 179 W/Milestrip Rd.	Intersection after tollbooth
125.4	0.2	Left onto Rt. 62/South Park Ave.	Intersection of Rt. 179/Milestrip Rd. and Rt. 62/South Park Ave.
126.5	1.1	Right (west) onto Big Tree Rd.	Intersection of Big Tree Rd. and Rt. 62/South Park Ave.
126.8	0.3	Right onto Bristol Rd.	Intersection of Bristol Rd. and South Park Ave., Hamburg
127.0	0.2	Left onto North St.	Intersection of North St. and Bristol Rd.
127.2	0.2	Park on entrance road, **STOP 6**	Penn Dixie Paleontological and Outdoor Education Center

STOP 6. **Penn Dixie Fossil Quarry.** The ridge to the south of the site is composed of the North Evans and Genundewa Limestones. Windom Shale comprises the bulk of the mud flats. Tichenor Limestone caps the small drainage to the northeast of the site, and Wanakah Shale lines the banks of the stream and the adjacent Fall Brook. For more information on the localities, see the Penn Dixie field trip in this guide.

Geologic units: Wanakah Shale Member, Ludlowville Formation, Hamilton Group; Tichenor Limestone and Windom Shale Member, Moscow Formation, Hamilton Group; North Evans and Genundewa Limestones, Genesee Formation.
Geologic age: Middle and Upper Devonian
Common fossils: Brachiopods, trilobites, corals, crinoids, bryozoans, cephalopods, fish bones

RETURN TO BUFFALO (FROM STOP 5)

Total	Distance	Directions	Location
112.8	0.0	Turn around, left onto Bullis Rd.	Intersection of Buffalo Creek Bridge ramp and Bullis Rd.
117.8	5.0	Left onto Transit Rd./Rt. 78 S	Intersection of Bullis Rd. and Transit Rd./Rt. 78
117.9	0.1	Right onto Rt. 400 N	Onramp for Rt. 400 N
122.7	4.8	Exit Rt. 400 to I-90 E	Onramp for I-90 E

Continue to 33 West (Downtown) or I-290

QUATERNARY GEOLOGY AND LANDFORMS BETWEEN BUFFALO AND THE GENESEE VALLEY

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INTRODUCTION

The Quaternary geology and landforms in western New York State north of the southern Finger Lakes (Valley Heads moraine) formed during the final recession of the Late Wisconsin ice sheet from its terminal position (Figure 1). There are a few places where Middle Wisconsin drift is preserved, such as in the Genesee Valley (Young and Burr, 2006), that indicate there is a more complex, older glacial record yet to be explored. This field trip explores the dominant geomorphic features created by the final stages of ice recession across a scarp-dominated bedrock landscape from Batavia to the southern Lake Ontario shoreline.

The complex sequence of proglacial lakes that accompanied glacial recession across the Great Lakes region is well documented in the geologic literature of the last century, but some events continue to be subjects of detailed study and debate. For the purposes of this field trip, the creation of glacial Lake Whittlesey (Figure 2) at the time of the Port Huron advance (circa 13,000 ^{14}C yrs BP) is an appropriate place to begin a discussion of the major events in western New York that culminated in the formation of early Lake Ontario, Lake Tonawanda, and the development of Niagara Falls in the Niagara Gorge.

LATE WISCONSIN HISTORY

Glacial Lake Whittlesey, with its westward outlet through the Ubly channel, barely extended into western NY as a narrow embayment that ended 17 miles east-southeast of Buffalo near the Marilla moraine (Figures 2 and 3). Its shoreline elevations range from 850 feet above sea level (asl) near Cattaraugus Creek to 910 feet asl 12 miles north of Hamburg (Calkin, 1970). This phase of deglaciation was followed by a series of three Lake Warren stages, of which the last and lowest (Warren III) may have followed a brief drop to a lower Lake Wayne stage. The Lake Warren strands are 45 to 70 feet below the Whittlesey level in the region at elevations between 750 and 865 feet asl (Figure 10). The Lake Whittlesey strand transects the Hamburg moraine, but not the Marilla moraine (Figures 2 and 3). This indicates that the drop to the oldest Lake Warren I position occurred before significant ice retreat from the Marilla moraine (Muller and others, 2003).

Both the Hamburg and Marilla moraines are considered to be Port Huron in age. The ice subsequently retreated from its Port Huron position to form the Alden, Buffalo, and Niagara Falls moraines, all of which are transected by the lowest strand of Lake Warren III. The barriers controlling the spread of Lake Warren waters were a series of closely-spaced ice positions against the Onondaga Escarpment further east between Batavia and Hamilton, NY (Calkin, 1970; Calkin and Feenstra, 1985). The entire recession and accompanying 6 or more lake stages from the Marilla moraine to the Niagara (Lockport) Escarpment (19miles) may have taken as little as 400 to 600 years (Karrow and others, 1961). Minimum ages for the draining of Lake Whittlesey are between 12,610 and 12,730 ^{14}C yrs BP (Muller and Calkin, 1993). During the multiple Warren stages western glacial waters finally connected with co-existing central NY lakes through channels running eastward from the Genesee Valley to Syracuse. Lowest Lake Warren formed when ice readvanced to either the Batavia or the Alden moraine (Figure 3).

As the ice receeded north to the Barre moraine, Lake Tonawanda formed with the Barre moraine at its northern edge. During its early existence, Lake Tonawanda was a shallow extension of the upper Niagara River and early Lake Erie (Brett and Calkin, 1996). During its highest level (590ft) near Niagara Falls the lake was approximately 58 miles long and 30 feet deep. The flow from early Lake Erie probably fluctuated greatly, so the erratically flowing waters from Lake Tonawanda spilled into glacial Lake Iroquois, when the ice eventually retreated north of the Niagara (Lockport) Escarpment. Multiple Lake Tonawanda outlets eventually formed along the Niagara Escarpment in response to the highly fluctuating Lake Erie outflow into Lake Tonawanda (Figures 3, 16, 17 19). The outlets, from east to west, were near Holley, Medina, Gasport, Lockport, and at Lewiston (Niagara River). Greater isostatic rebound to the northeast gradually caused abandonment of the eastern outlets and shifted the main flow toward the Niagara River (Lewiston outlet). Radiocarbon ages on wood from the Lockport area indicates that the main discharge had shifted from Lockport to Lewiston by 10,920±160 ^{14}C yrs BP (Brett and Calkin, 1996). A more extensive discussion of the Lake Tonawanda history and the formation of Niagara Falls is contained in Brett and Calkin (1996).

It has been speculated that the catastrophic draining of glacial Lake Agassiz between 11,000 and 10,500 years BP caused major increases (3 to 10 times) in flows to Lake Tonawanda when Lake Agassiz discharged to Lake Superior and possibly through Lake Erie (Brett and Calkin, 1996). However, quiet-water marl deposits in the Lockport spillway (Figure 19) suggest that flow was limited during this inferred Lake Agassiz interval. Lake Tonawanda waters in the western part of the basin could have persisted until only a few hundred years ago, but the shift of full discharge of all the Great Lakes to the Niagara River is dated at 3780 years BP by a mollusk assemblage (Brett and Calkin, 1996).

Glacial Lake Iroquois, the final glacial feature visited on this trip, is probably the best known of the proglacial lake stages in the Ontario basin, because of its prominent relief and the location of the original Route 104, which follows the Lake Iroquois strand along much of the Ontario shore. Lake Iroquois formed around 12,200 ^{14}C yrs BP, and persisted for several hundred years, based on existing chronology and the strong development of its shoreline. Many of its features are as well developed as those seen along the modern shoreline of Lake Ontario. Early Lake Ontario was at a much lower level when the ice sheet finally retreated from the north flank of the Adirondacks (Covey Hill outlet) and allowed Lake Iroquois waters to escape to the Atlantic Ocean via the St. Lawrence Valley. Lake Ontario gradually has risen to its present elevation from this low stand, which was between 250 and 300 feet lower at Rochester, NY.

This trip will explore the landforms and field relationships seen from the Batavia moraine northward to the Lake Iroquois shoreline northwest of Lockport, NY. A major focus of the trip is the contrast between the bedrock and glacial relief. The route will permit comparison of the morphology of terrestrial and subaqueous moraines and the internal stratigraphy of sediments that may have been deposited by meltwater flowing into Lake Tonawanda immediately north of the Barre moraine. The influence of the Onondaga and Lockport Escarpments on the glacial history is closely linked to the evolution of the Niagara River and Niagara Falls.

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NYSGA 2006 – FIELD TRIP B6 – SUNDAY OCTOBER 8TH, 2006

QUATERNARY GEOLOGY AND LANDFORMS BETWEEN BUFFALO AND THE GENESEE VALLEY

Road Log (Follow route and numbered Stops on Figures 3-6)

0.0 Depart Adam's Mark hotel parking lot to the right onto Lower Terrace St (one-way street).

0.1 Left onto West Seneca St.

0.1 Left onto Franklin St.

0.3 Take a right onto Church St.

0.7 Turn left onto Elm St.

1.4 Take onramp to Hwy 33 East.

9.8 Turn right onto Genesee St. Pass Buffalo-Niagara airport on left.

12.0 Cross Transit Rd.

12.8 Cross Ellicott Creek.

16.6 Pull into driveway on left and park on one side (Figure 9). **STOP 1.**

The Buffalo/Niagara Falls moraine complex. Gravel quarries are cut into a large outwash fan [labeled Wisconsin outwash gravel (W_{og}) on Muller (1977)], which borders the Buffalo Moraine. Borings made for the study of the gravel aquifer in this area show two till units bracketing up to 40 ft. of sand and gravel, indicating that the outwash fan pre-dates a more recent ice advance. Thin red lake clay partings separate thick rippled sand units, indicating that the outwash sand/gravel was likely deposited subaqueously. The overlying till and sand/gravel unit shows evidence of flow and deformation. The till has been traced at least to the Alden Moraine (Figure 3), 8 km to the south.

20.9 Take a left at the stop sign onto Walden Ave – turns back into Genesee St. – still on Hwy 33.

23.1 Cross Crittenden Rd. and onto Lake Warren beach ridges (Figure 10).

Glacial Lakes of the Erie Basin. Because the Laurentide Ice Sheet (LIS) was sourced in the north (Figure 1), and Lake Erie drains north, large pro-glacial lakes were created as the LIS flowed into and out of the Lake Erie basin. The highest lake level recorded in the greater Buffalo region, termed Glacial Lake Whittlesey (Figure 2), created mainly an erosional escarpment between 850 and 900 ft. asl (260-274 m asl). Present Lake Erie elevation is 570 ft asl (174 m asl). As the LIS retreated to the Alden, Buffalo, Niagara Falls and Batavia moraines (Figure 3), the lake level dropped to a new level termed Glacial Lake Warren, which deposited beaches along the field trip route from 820 to 850 ft. asl (250-260 m asl) between 12,700 and 12,500 ^{14}C yr BP. Following Glacial Lake Warren, lower lakes termed Grassmere, Lundy, Algonquin and Dana existed; these lakes pre-date non-glacial Lake Tonawanda (see below), which existed from slightly before 12,000 ^{14}C yr BP up to ~9000 ^{14}C yr BP. See shoreline elevations and relative amounts of postglacial uplift on Figure 8.

Key chronological constraints on the glacial history of western New York (from Muller, 1977; Calkin, 1970).

22,800 ^{14}C yr BP	Uppermost organic materials in the buried St. David's gorge that pre-date the LGM
14,100 to 13,400	Valley Heads, Lake Escarpment, Gowanda Moraine deposits
13,000	Lake Whittlesey, Hamburg Moraine
12,700	Lake Warren (high stand), Alden Moraine
12,500	Lake Warren (low stand), Batavia Moraine, Niagara Falls Moraine
12,400 to 9,000	Existence of Lake Tonawanda
12,200 to 11,000	Lake Iroquois (main phase)

37.5 Take a left onto River St. – a small road that may not be marked.

38.3 Turn left onto Main St. just after crossing Tonawanda Creek.

38.4 Veer right onto Route 63 as it departs Route 5.

40.8 Turn left onto Galloway Rd. (Figure 11). This road closely follows the Batavia moraine.

The Batavia moraine. The Batavia moraine near Batavia has noticeably greater and more irregular relief than the moraines that will be visited to the north and west. This is due, in part, to the formation of some of the younger moraines at a submerged ice front (Figure 7). In this environment moraines are subject to being covered by lacustrine sediments and then to being winnowed and eroded by waves as lake levels fall. The west end of the Batavia moraine (930+ ft asl) projects above the elevations of Lake Warren, and thus it would have escaped wave erosion when an eastern finger of Lake Warren formed a narrow extension into the Genesee Valley with shorelines at current elevations close to 860 ft asl near Geneseo, NY. While driving along this Batavia moraine segment, note its dimensions and relief for comparison with the more subdued Barre moraine to be visited later (Figures 3, 5). According to Calkin (1970) the Batavia moraine is believed to have formed around the time of the lowest Lake Warren stage, which coincided with ice recession from that position (Figure 3). Lake Warren had three phases, Warren I and II are higher and older, Warren III is the lowest and youngest, and followed a short-lived drop to a "Lake Wayne" stage (Figure 8).

42.5 Turn right onto Downy Rd.

43.2 Turn left onto Batavia – Oakfield TNL Rd. At mileage 46.0 route crosses crest of Batavia moraine.

46.5 Turn right onto Maple Rd.

47.4 Traveling North, Maple Rd. drops off the Onondoga Escaprment. Old quarry and recycling center exposes Onondoga limestone (Middle Devonian). Note how, in general, the region's resistant bedrock units lead to relief (30-100 ft. escarpments, Figure 12) that dwarfs the glacial topography. **STOP 2.** Examine the nature of the resistant Onondaga Formation.

49.2 Turn left onto Ham Rd.

49.5 Turn right onto Knowlesville Rd.

50.3 Turn right onto Lewiston Rd.

51.0 Crossing Nichols Hill drumlin. Note height and trend of this prominent drumlin.

51.6 Drive straight through intersection as Lewiston Rd. veers right. You are now on Lockport Rd.

53.4 Turn left onto Albion Rd.

55.3 Begin ascent of drumlin along its axis.

55.7 Pullout on right side of road affords nice view of drumlin relief and inter-drumlin bog (Figure 13). The two drumlins here, trending WSW, show the direction of ice flow during the late stages of the last glaciation. **STOP 3.**

Tonawanda Lowland. This location is approximately one mile north of the southern edge of Lake Tonawanda (note surrounding wetlands), which was nearly 7 miles wide at this longitude. Lake Tonawanda's northern limit was determined by the rising dip slope of the Lockport Escarpment. The southern limit was formed, in part, by moraines (Figure 3) and by the gradual increase in elevation toward the Onondaga Escarpment (Figure 7). Lake Tonawanda formed following the lowering of Lake Warren toward the Lake Iroquois level, which is north of the Lockport Escarpment (Figure 3). Lake Tonawanda had several northerly outlets (Figure 3, Arrows), which are progressively older and higher toward the east. Lake Tonawanda waters flowed off the escarpment into Lake Iroquois, when it first formed. As shown on Figure 2, the easternmost outlet formed first and was succeeded in turn by the more westerly outlets as indicated by letters A through E on Figure 3. Current threshold elevations near the eastern (Holley outlet) are near 650 ft asl, whereas the threshold elevations near Gasport are closer to 600 ft asl. The westerly displacement of the outlets was dictated by the greater amount of postglacial rebound toward the northeast (Figure 8). Eventually the Lake drained completely after the outflow was shifted toward the modern position of the Niagara River.

56.1 Veer left as road descends northern drumlin flank.

58.1 Road now called Eagle Harbor Rd. Continue through intersection. Note stony till just south of intersection. Road crosses Barre moraine.

59.8 Turn left into Eagle Harbor Sand and Gravel quarry. Part of the Burma Woods esker and kame moraine complex. Dozens of small-scale quarries and this large one in this vicinity. **STOP 4**. Planned lunch stop. After visit, exit driveway and turn left.

Burma Woods Esker. This esker and outwash complex (seen in a modern excavation) may have formed during the earliest stage of Lake Tonawanda, when the ice had just retreated from the Barre moraine (Muller and others, 2003). Muller and others (2003) note that the position of this and similar eskers indicates early southward meltwater flow into Lake Tonawanda. Although the esker deposits are not currently visible in the modern pit, it is possible that the deformed sediments being excavated are contemporaneous with an ice position fronting on early Lake Tonawanda. The sediments visible in August 2006 at the working face (Figure 15) indicate that saturated sediments were deformed by intraformational slumping, and possibly by upward groundwater discharge immediately in front of the ice margin. This will be the longest stop on the trip (including lunch) and will allow time for discussion of how the currently exposed sedimentary structures may relate to the history of ice recession from the Barre moraine, or to the development of Lake Tonawanda.

60.3 Turn left onto Maple St. and pull off to the right at one tenth of a mile to see the roadcut through the Burma Woods esker. This esker is only several meters high, but contains well-rounded sand and cobbles with some boulders. **STOP 5.** Watch for traffic.

60.9 Turn left onto Pine Hill Rd.

62.3 Turn right onto Gray Rd.

63.0 Turn right onto Hemlock Ridge Rd. This road is built atop the Barre Moraine. Note how the Burma Woods esker trends perpendicularly to the Barre Moraine (Figures 3, 5).

65.3 Our route takes a quick jog. Turn left onto Shelby Rd, then immediately right onto Fletcher Chpl.

68.7 Cross S. Gravel Rd.

69.7 Turn left into the trailhead parking lot just after crossing the small creek. This little valley held one of the spillways of Lake Tonawanda. **STOP 6.** Depart parking lot turning left. Take an immediate left onto W. Shelby Rd.

Lake Tonawanda spillway. This is the location of one of the main spillways from Lake Tonawanda across the Lockport Escarpment south of Medina (see Figure 3 and discussion for Stop 3). At this location the lake outflow was able to cut down through the Barre Moraine, probably beginning at a low saddle in the moraine crest.

71.2 Cross Salt Rd.

74.1 Cross Griswold St.

75.1 Left at T-intersection onto Graham Rd.

75.6 Stay straight onto Route 77.

78.4 At junction of Rt. 77 and Gasport Road turn south.

79.3 Turn right on Lincoln Ave.

79.6 Cross Lake Tonawanda outlet channel.

80.9 Turn right on Hollenbeck.

82.5 Turn right on Rt. 77.

82.8 Turn left onto Mill Road (parallel to outlet).

84.2 Left back onto Gasport road.

84.5 Left into Victor Fitchlee Park for rest stop and short walk into Royalton Ravine (lake outlet at East branch of Eighteenmile Ck). **STOP 7**. This Park has restroom facilities and is located adjacent to a deep outlet channel passing through Gasport. A short walk down Royaltine Ravine leads to the broader Tonawanda channel, and the Ravine exposes shaley bedrock, which indicates how thin the glacial drift is at this location. Leave parking lot turning left back onto Gasport Rd.

84.7 Turn right onto Mountain Rd.

85.4 View to left off of Niagara Escarpment. Smoke stack in distance on shore of Lake Ontario, 12 miles away.

85.7 Turn left onto Bolton Rd. Road descends the Niagara Escarpment.

85.9 Cross Route 31.

86.1 Turn left onto Telegraph Rd., which follows the Erie Canal.

87.0 Turn right onto Main St.

87.5 Turn left onto Slayton-Settlement Rd.

92.5 Left onto Lake Ave.

93.8 Right onto Market St. at stop sign just after crossing the canal bridge. Note series of locks to get the Erie Canal across the Niagara Escarpment.

94.9 Right onto Main St.

95.3 Right onto N. Transit St.

96.0 Left onto Outwater Dr.

96.6 Outwater Park. Overlook to Lake Ontario from top of Niagara Escarpment. **STOP 8**. Proceed out exit of park onto residential street. Somewhere between here and Michigan St. park and walk to footpath 100 feet down Michigan St. from Craine St. – see Mi. 98.8. (Parking is difficult at this Stop and maybe impossible for a bus.)

96.7 Turn right onto Craine St.

97.0 Turn right onto Michigan St.

97.0 Hidden driveway on right leads to footpath to old quarry site where glacial drift has been stripped away revealing a striated fossiliferous surface of the Lockport Dolomite. **STOP 9.**

98.0 Turn left onto Gooding St.

98.7 Good exposure on left of Medina Group in Niagara Escarpment. Alternate **STOP 9**.

99.5 Turn left onto Old Niagara Rd.

102.3 Get onto Route 104 going west. Intersection is awkward; take a right at stop sign then a quick left onto Route 104. Route 104 follows the glacial Lake Iroquois beach.

Glacial lake Iroquois. Glacial Lake Iroquois is probably the best known of the proglacial lake stages in the Ontario basin, because of its prominent relief and the location of the original Route 104, which followed its course along much of the lake shore. The development of the Niagara Gorge and the retreat of Niagara Falls, and their relationship to Lake Tonawanda, are complex and interesting stories, described recently by Brett and Calkin (1996) in the guidebook for the Northeast Section Meeting of the Geological Society of America in Buffalo.

Lake Iroquois formed around 12,200 ^{14}C yrs BP, and persisted for several hundred years, based on existing chronology and the strong development of its shoreline. Many of its features are as well developed as those seen along the modern shoreline of Lake Ontario. It is not generally realized that Lake Ontario subsequently reached a much lower level when the ice sheet finally vacated the north flank of the Adirondacks (Covey Hill outlet) and Lake Iroquois waters escaped to the Atlantic Ocean (glacially lowered) via the St. Lawrence Valley. Since reaching this lower level, which was between 200 and 300 feet lower at Rochester, NY, the modern lake has been steadily rising to its present level. This is the result of postglacial tilting of the basin and uplift of the threshold that controls flow out through the St. Lawrence River. It was previously thought that Lake Ontario was a marine or brackish arm of the ocean for a short period (Champlain Sea), but modern studies have demonstrated that marine waters did not advance much west of Ogdensburg, NY, during that post-Iroquois, early Lake Ontario low stand. Further reading about the glacial history of the Ontario and Erie basins is contained in Geological Association of Canada Special Paper 30, *Quaternary Evolution of the Great Lakes* (Karrow and Calkins, 1985), and in Special Paper 35, *The Late Quaternary Development of the Champlain Sea Basin* (Gadd, 1988).

104.6 Turn right onto North Ridge Rd. Here, the road is still on a beach berm, but this beach is a spit-like feature that diverges westward from the main Iroquois beach strand.

104.9 Pull onto right shoulder. Stop to see sand and gravel of the Iroquois beach. **STOP 10.** Proceed westward on North Ridge Rd.

- IF TIME PERMITS, DRIVE TO LAKE ONTARIO -

106.0 Turn right onto Route 425. You immediately drive off Iroquois beach and onto flat Iroquois lake plain.

113.1 Turn left onto Route 18.

115.1 Turn right into Wilson Tuscarora State Park. Potential **STOP 11** (time permitting). Compare character of modern Ontario beach and its sediments with those most recently viewed at Stop 10.

END OF TRIP, Return by bus to Buffalo.

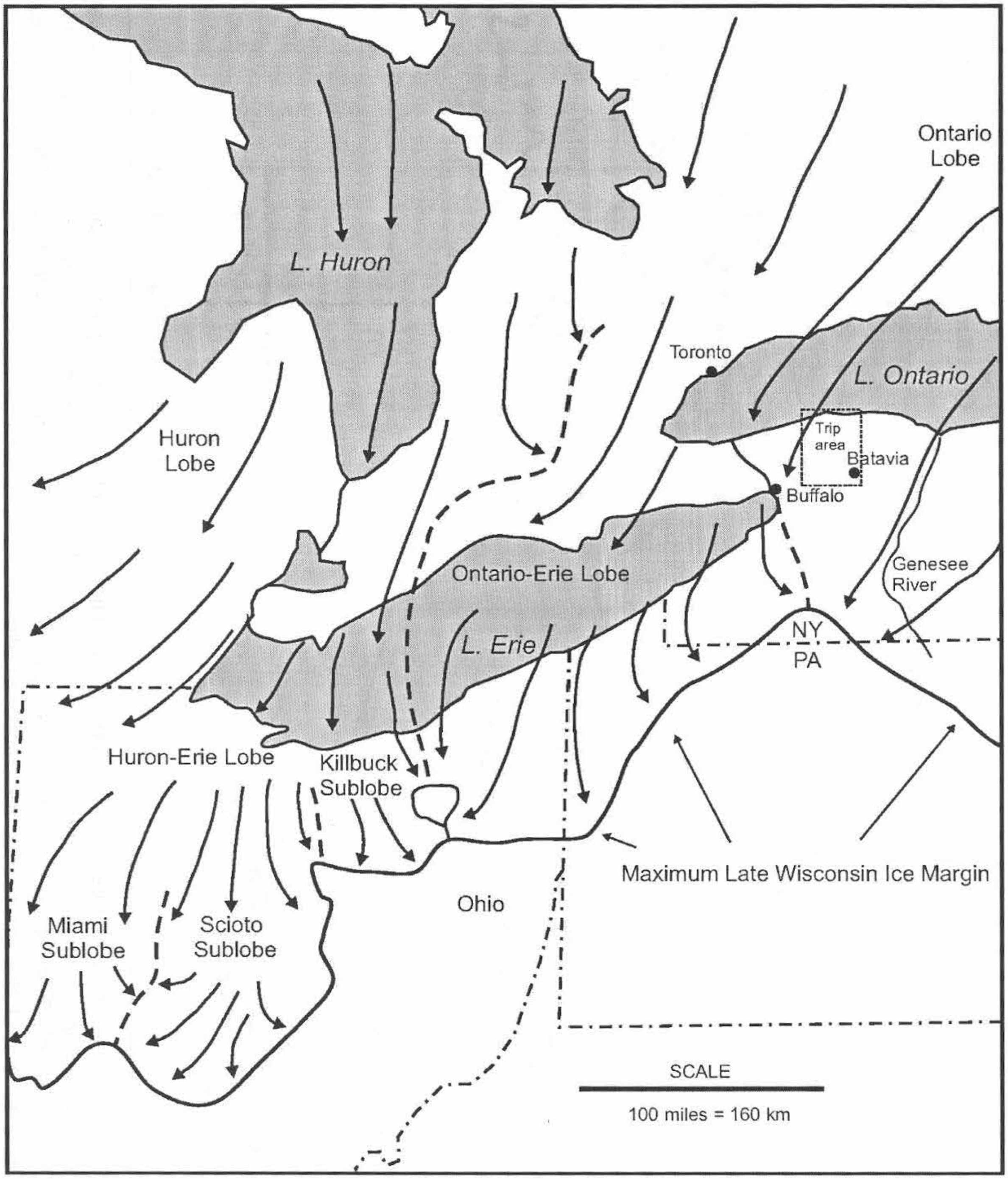


Figure 1. Position of Maximum Late Wisconsin ice margin in western Lake Ontario and Lake Erie basins. Large arrows show approximate flow lines in various sublobes. Trip area is outlined by box near Batavia, NY. See greater detail of trip area in Figures 2-6. Modified from Muller and others, 2003.

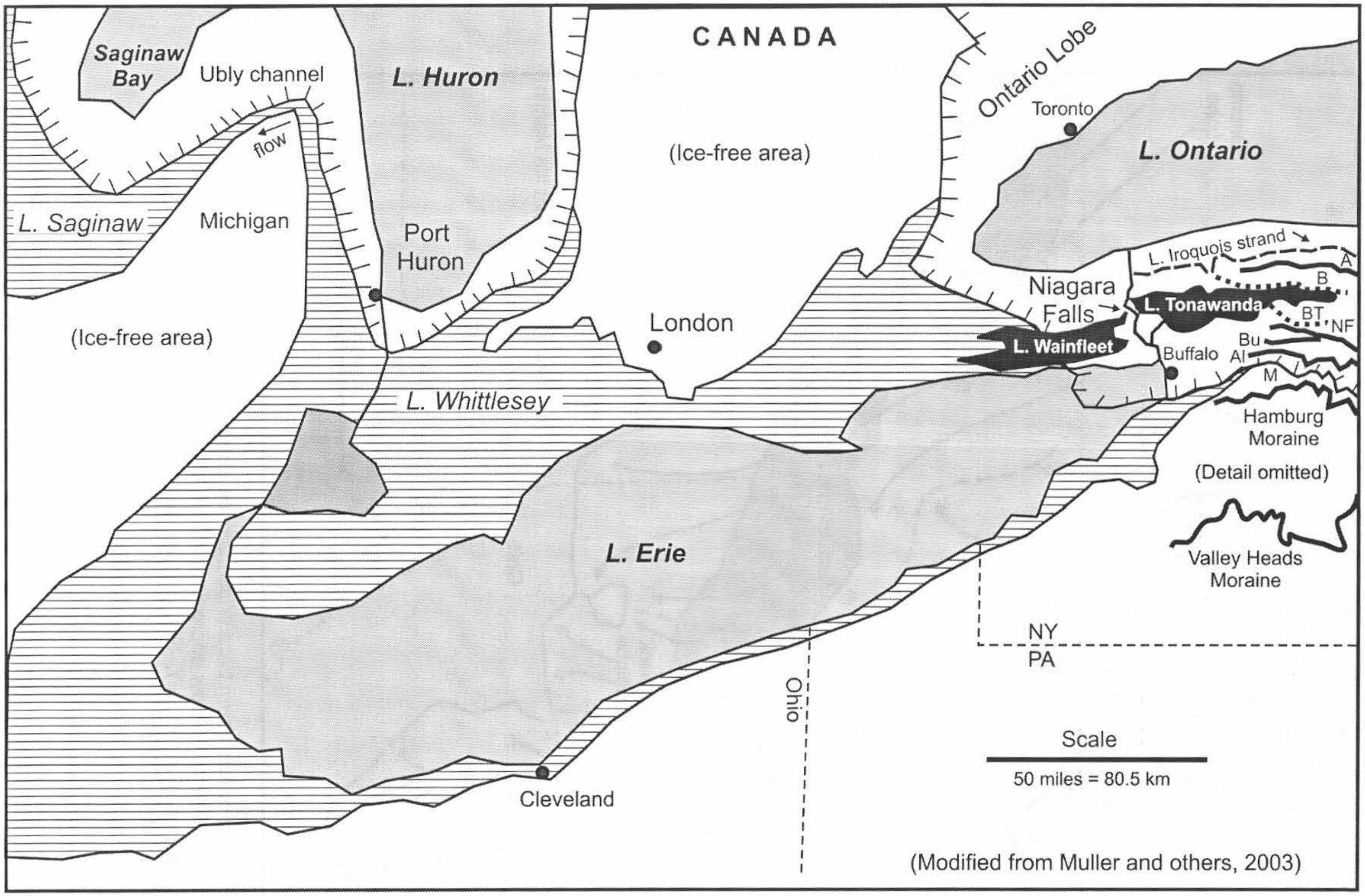


Figure 2. Ice position and selected geologic features near the end of Port Huron ice advance with extent of glacial Lake Whittlesey (13,000 to 12,700 ^{14}C years ago) ending at Marilla moraine. Lake Tonawanda-Wainfleet shown in black with positions of selected moraines [A=Albion, B=Barre (dotted), BT=Batavia (dotted), NF=Niagara Falls, Bu=Buffalo, Al=Alden, M=Marilla]. The conditions shown set the stage for the creation of Lake Tonawanda and Niagara Falls.

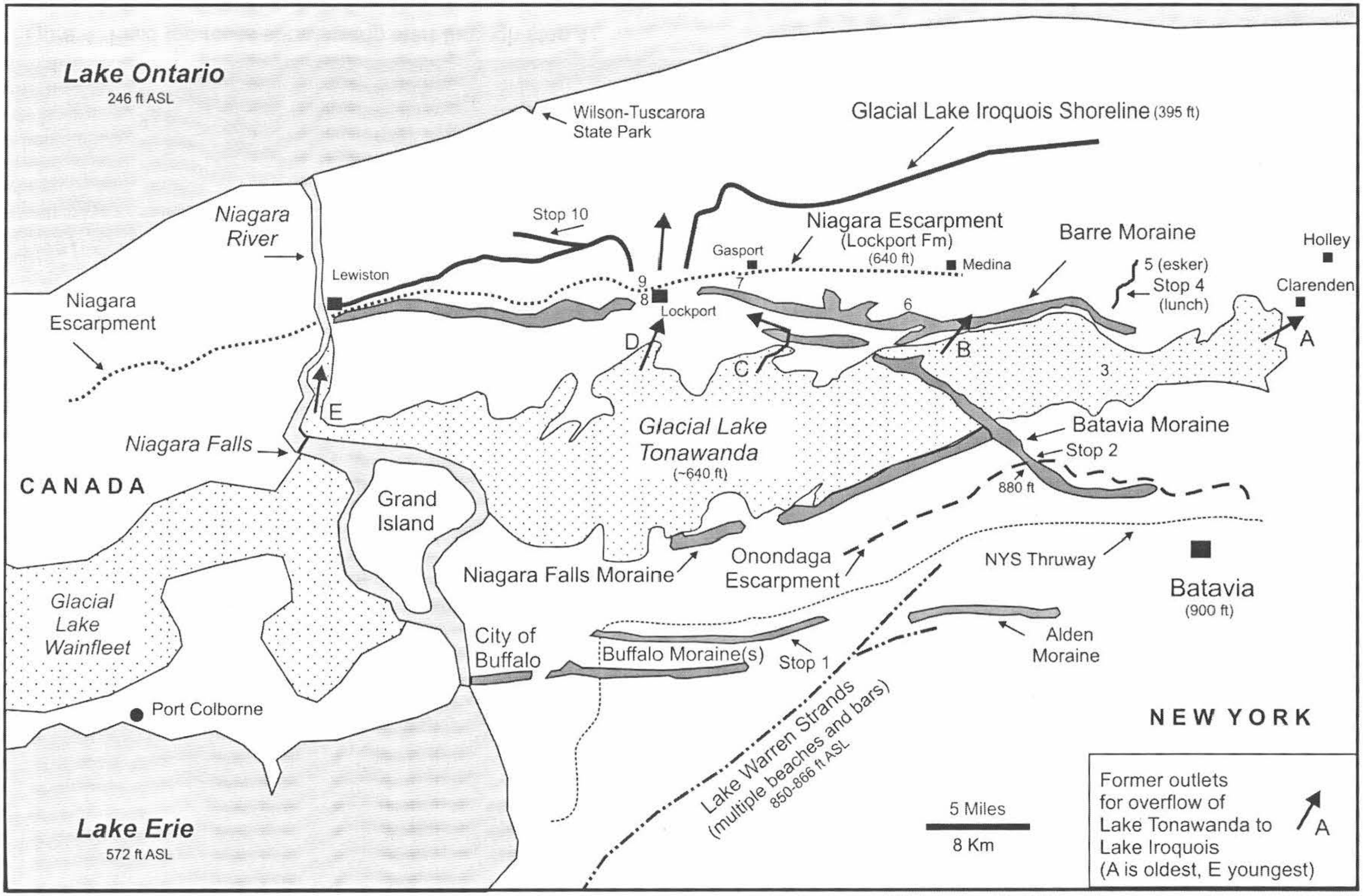


Figure 3. Geologic features to be visited and discussed on trip. Numbers 1-10 indicate locations of planned stops. Moraines, shorelines and other features compiled from map by Muller (1977) and figures in Muller and others (2003).

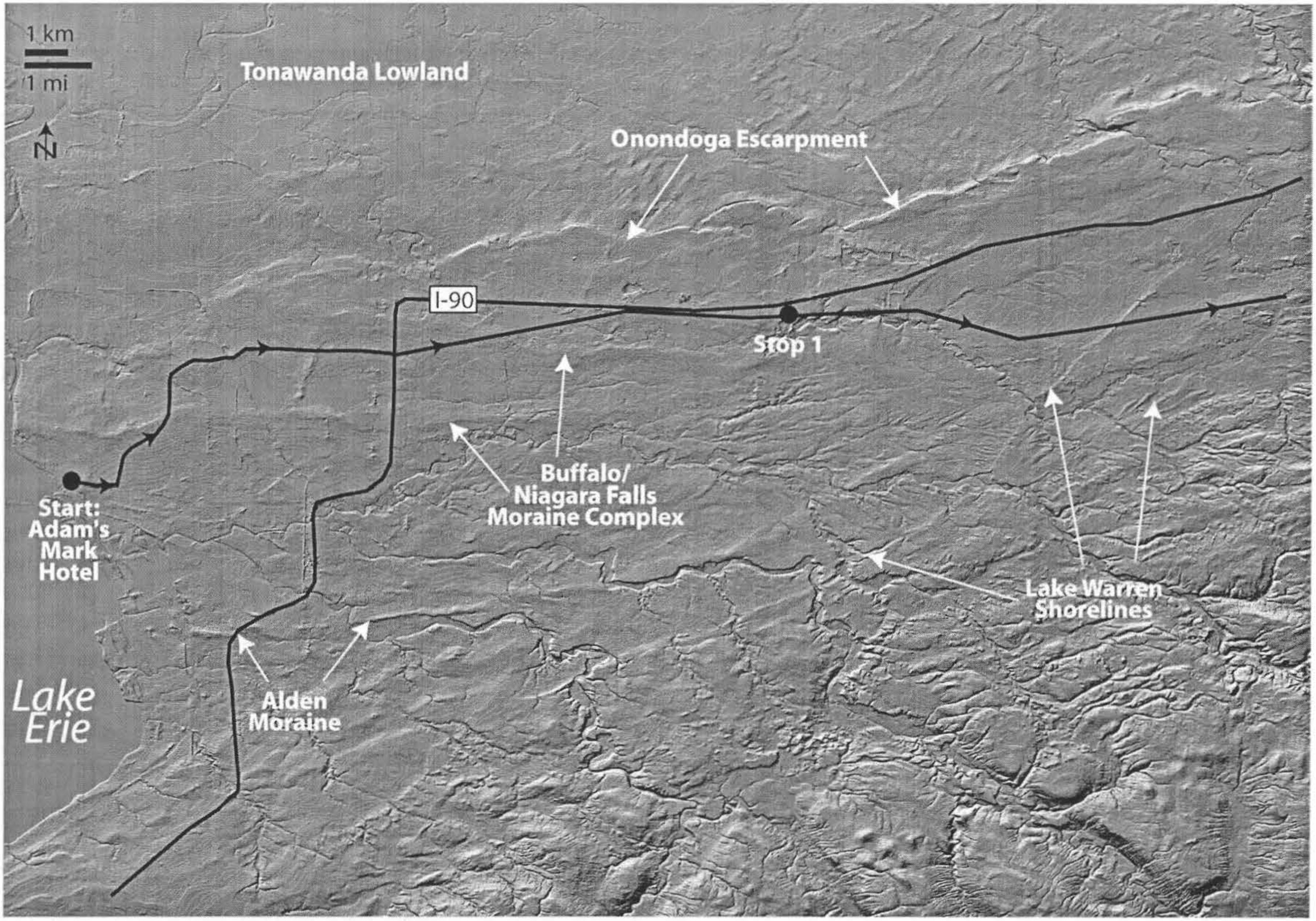


Figure 4. Field trip route showing trip start through stop 1.

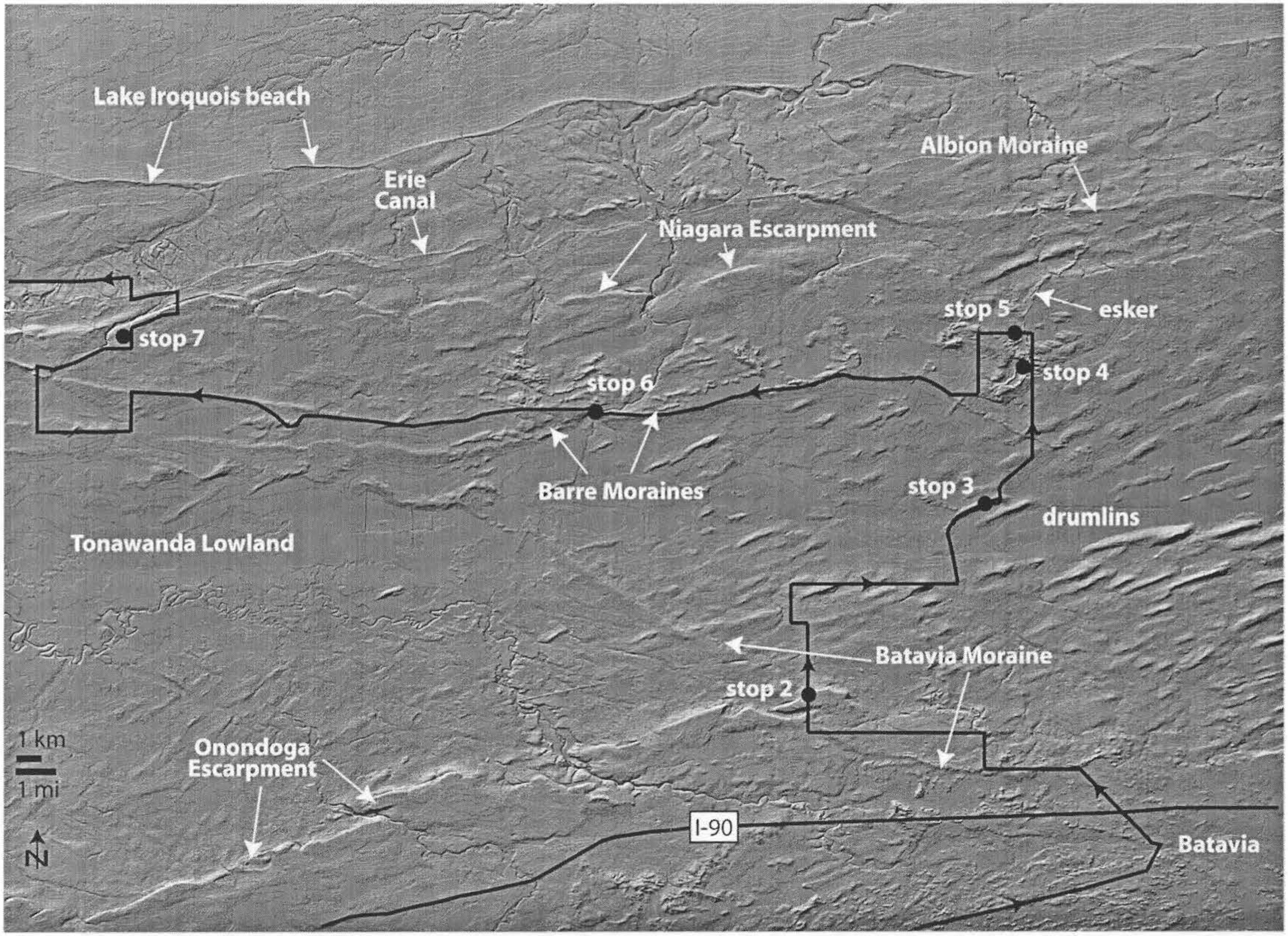


Figure 5. Field trip route showing stops 2 through 7.

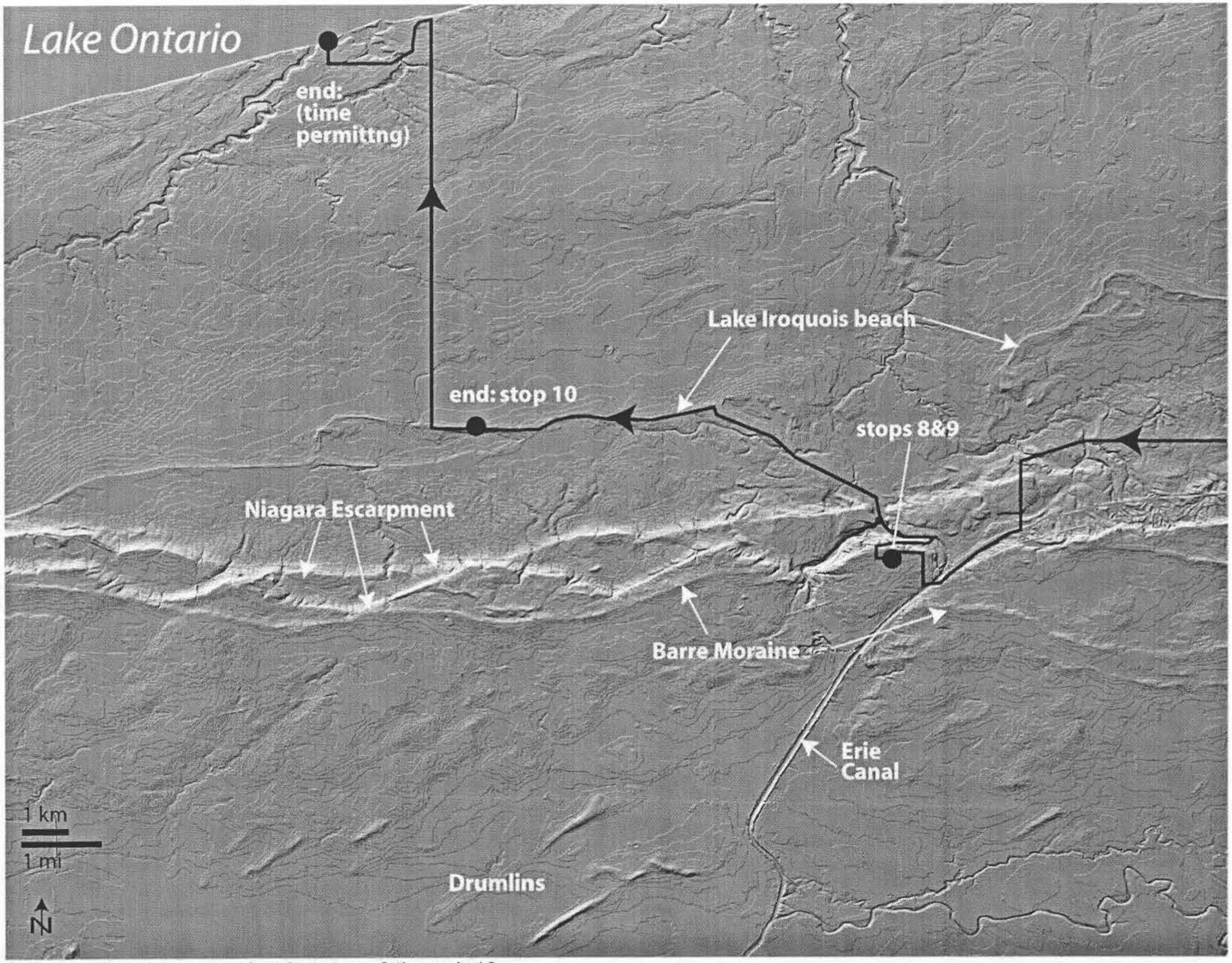


Figure 6. Field trip route showing stops 8 through 10.

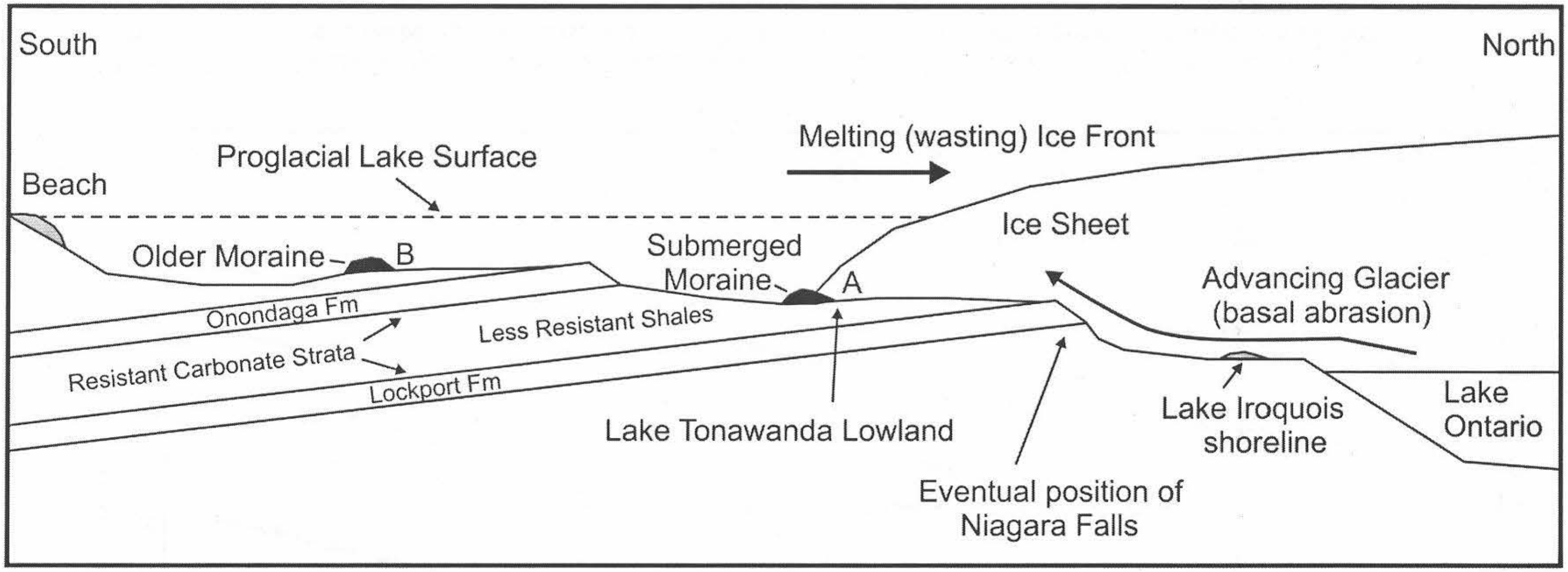


Figure 7. Schematic diagram of important topographic elements and resistant bedrock escarpments that controlled the development of existing geologic landforms as the Laurentide glacier retreated between 13,000 and 12,000 ^{14}C years ago. As the ice was melting back at the edge, the lower portion of the glacier was still advancing and eroding the bedrock. Beaches formed at the edges of glacial lakes also are referred to as "strandlines" or "strands". Some moraines formed under water in the shallow lakes (A), sometimes creating more subtle, low-relief landforms, also referred to as "morainal banks", "grounding-line wedges" or "DeGeer moraines". Some morainal deposits also were buried by fine sediments and then eroded by waves (B) as the lake surfaces were lowered to the next stage. This explains the subtle, smooth relief of some moraines. The ultimate level of modern Lake Ontario is shown and is still rising from a lower stand as a result of the postglacial tilting of the basin. Differential tilt of the Ontario basin is ongoing at about 1 ft/100 years from northeast (Ogdensburg) to southwest (Buffalo).

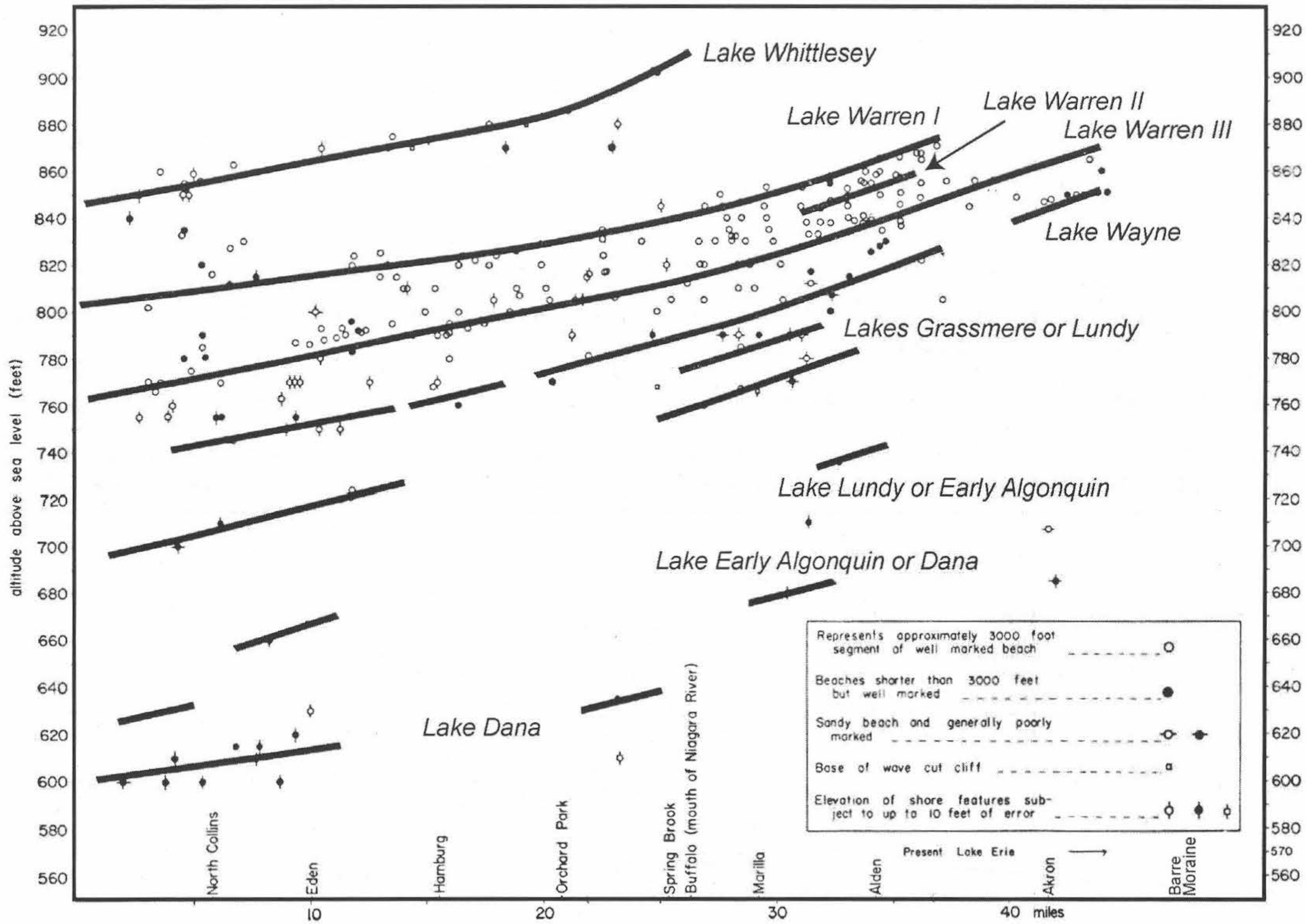
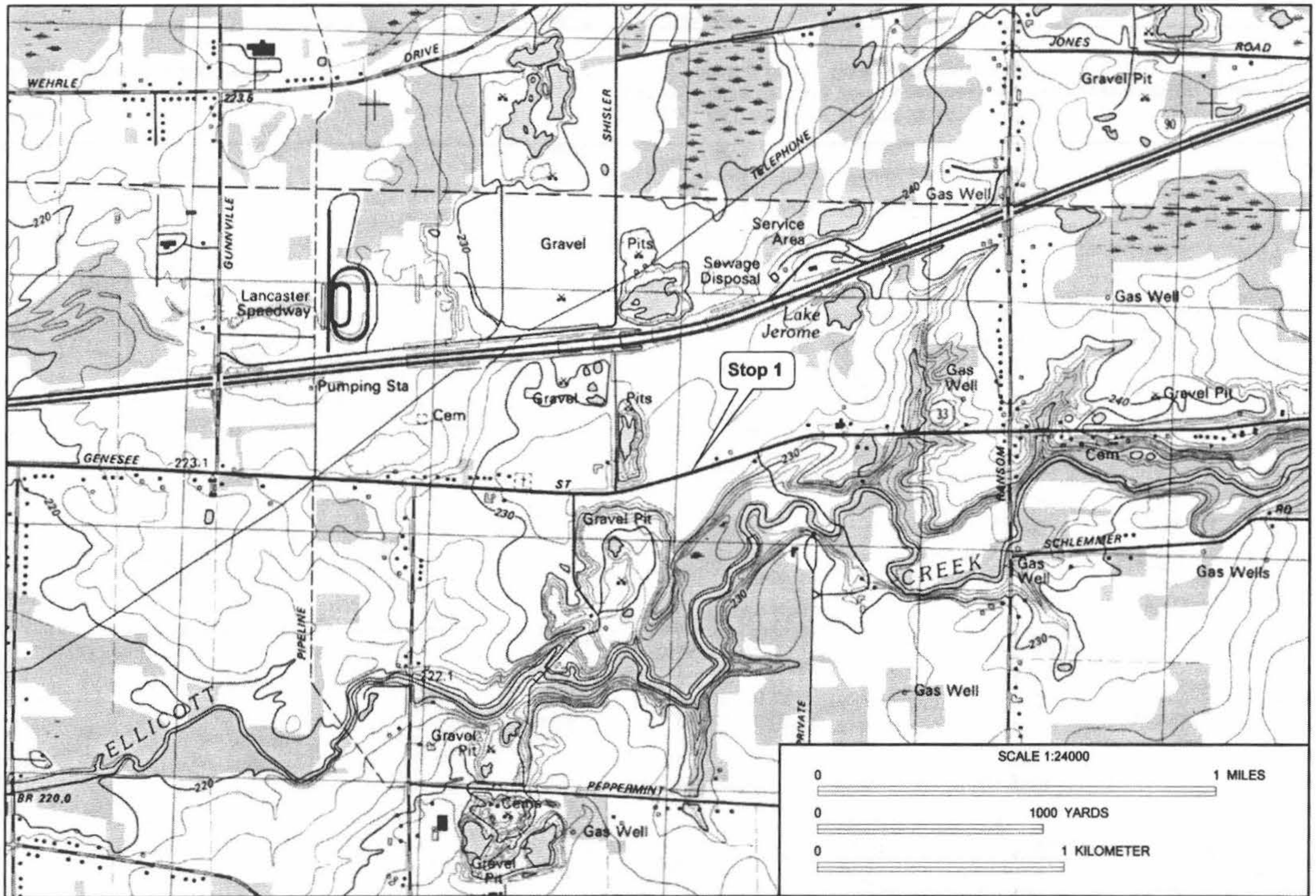
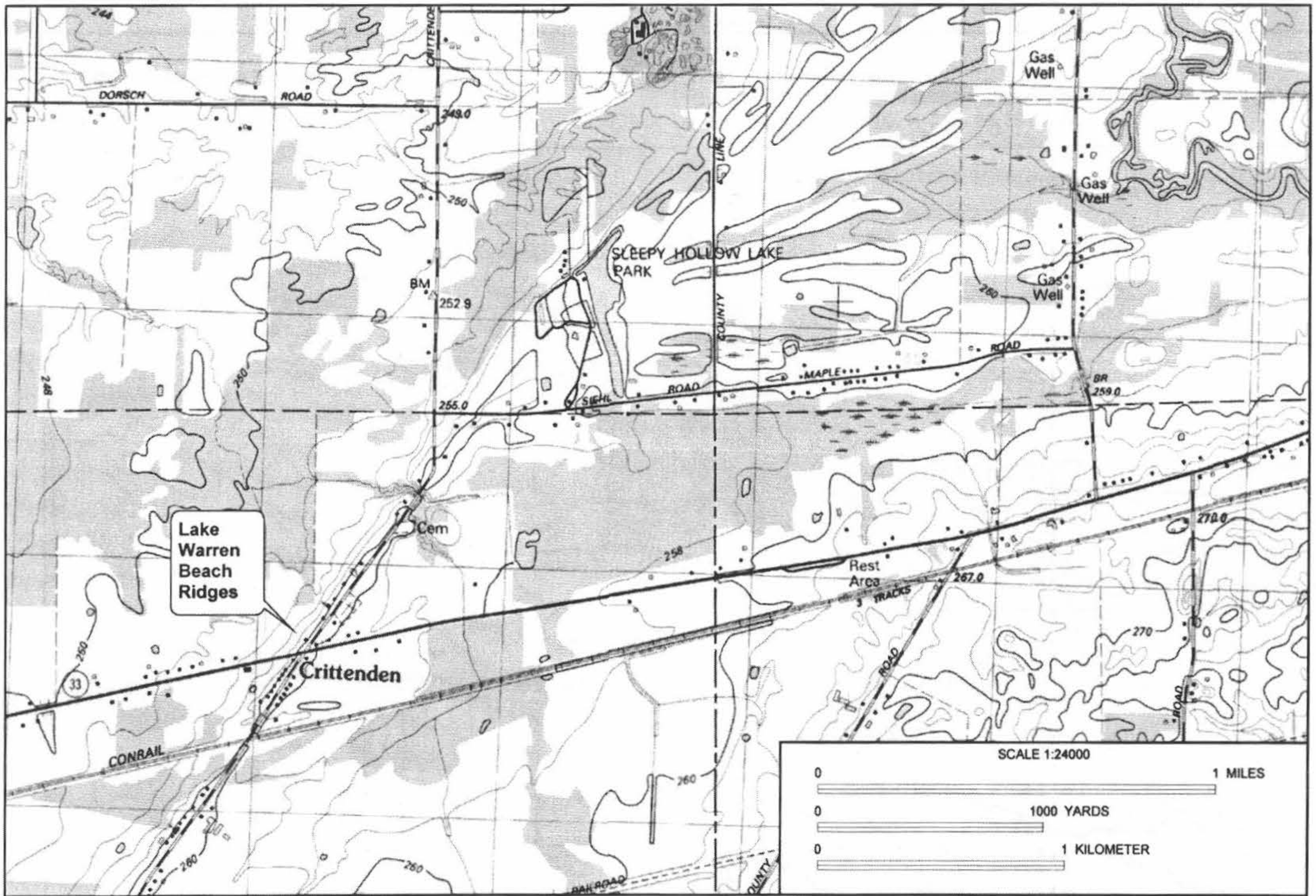


Figure 8. Glacial Lake shorelines in the Lake Erie Basin (modified from Calkin, 1970). The increasing elevation of the shorelines from southwest to northeast is a result of isostatic rebound from a greater ice load in the north.



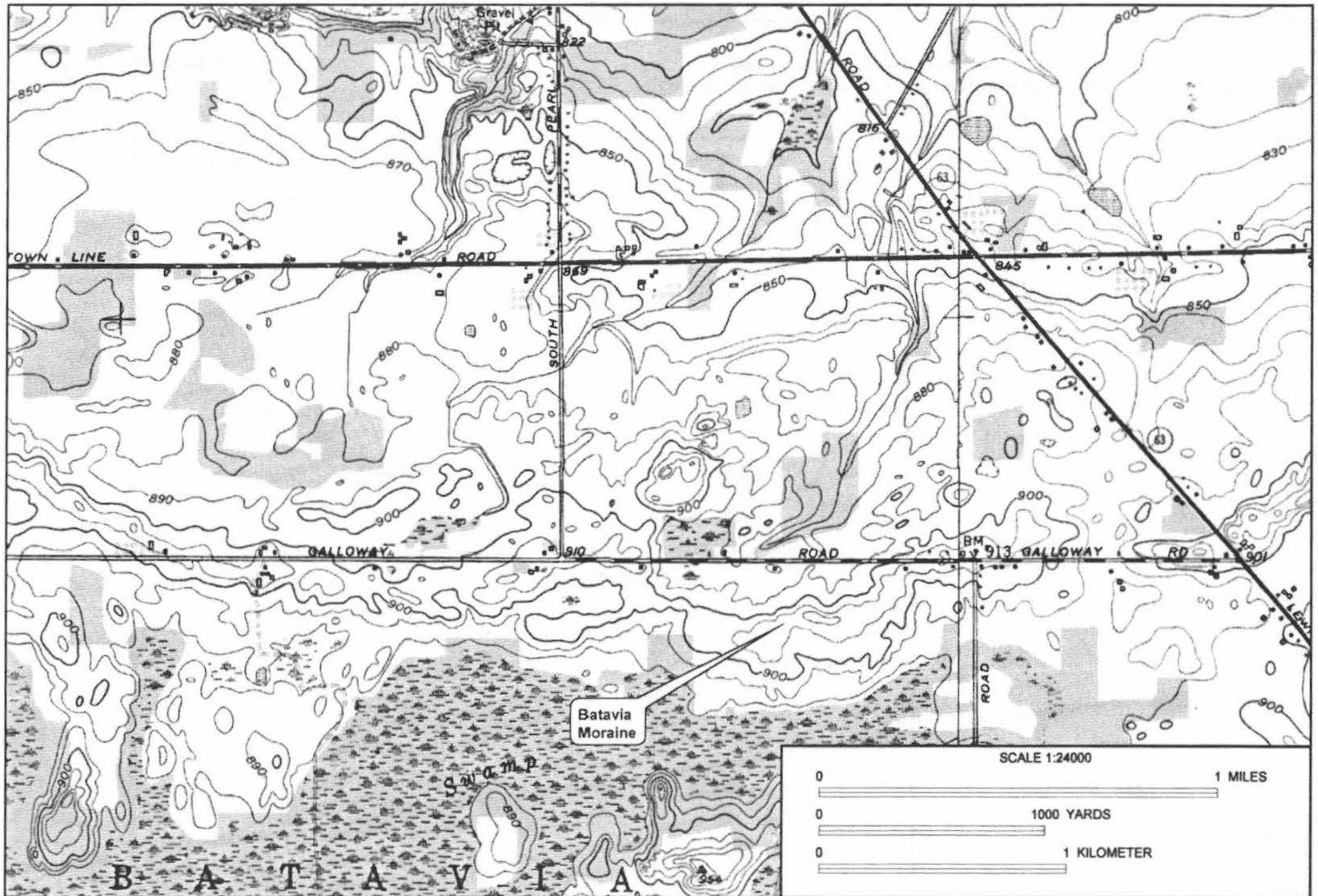
Name: LANCASTER
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 042° 56' 45.00" N 078° 36' 39.09" W
Caption: Figure 9. Buffalo-Niagara Falls moraine complex. Contour interval 2 meters.



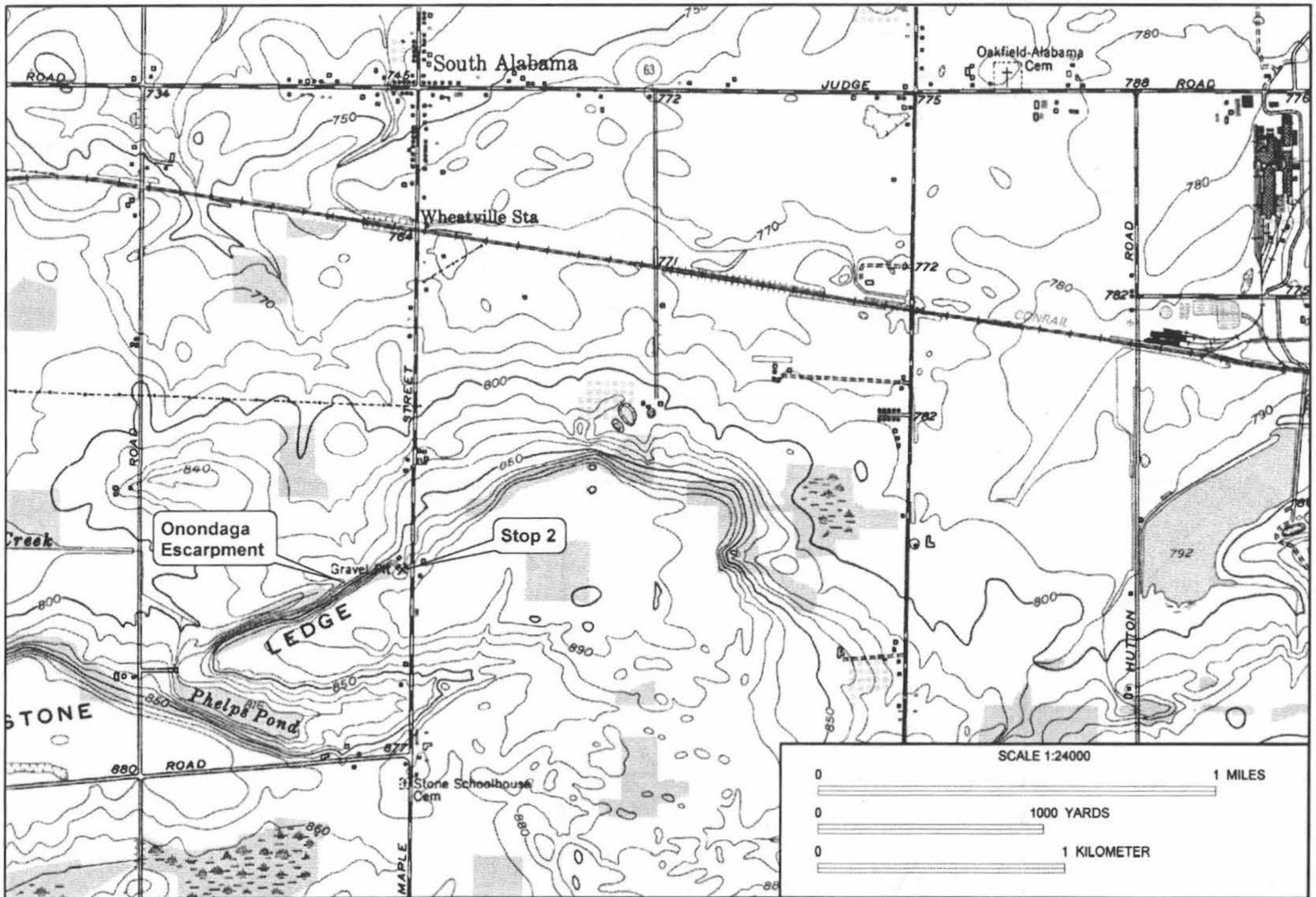
Name: CORFU
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 042° 57' 10.64" N 078° 28' 02.11" W
Caption: Figure 10. Lake Warren beaches fan out into several distinct ridges north of Rt 33. Contour interval is 2 meters.



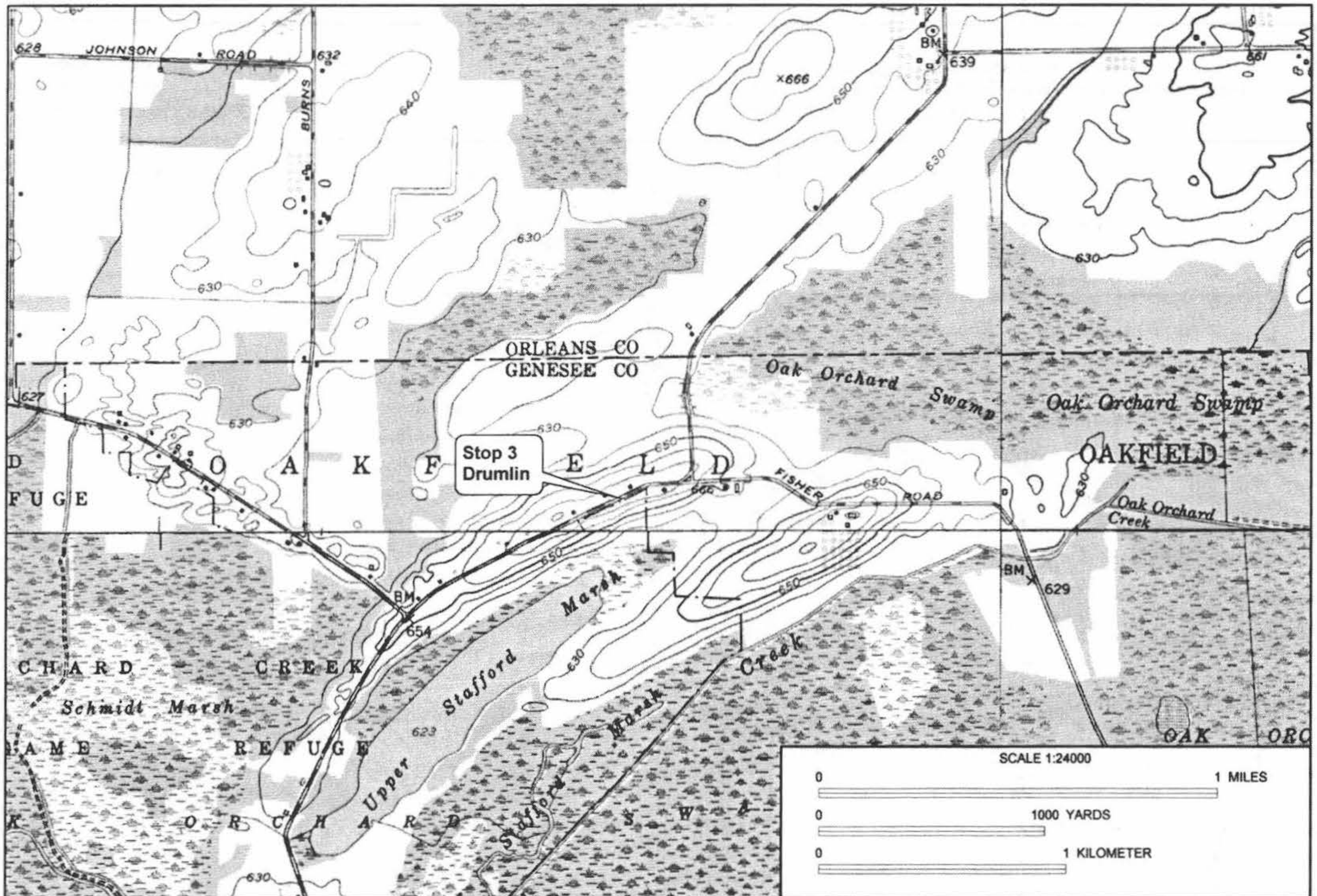
Name: OAKFIELD
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 02' 12.32" N 078° 15' 53.46" W
Caption: Figure 11. Batavia moraine along Galloway Road; mile 42.6 to 44.3. Contour interval is 10 feet.



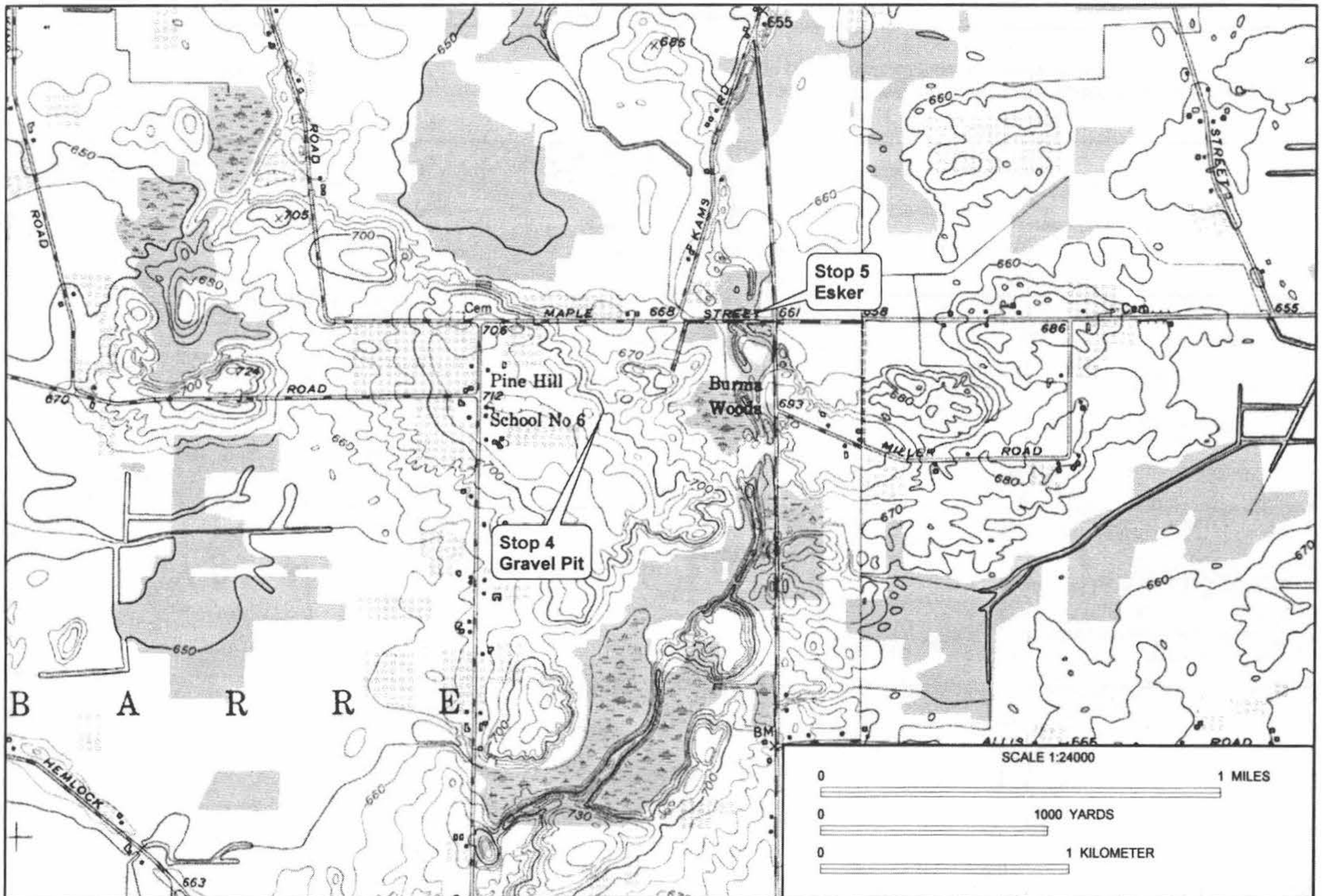
Name: OAKFIELD
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 03' 38.07" N 078° 19' 25.50" W
Caption: Figure 12. Onondaga Escarpment at elevation of approximately 880 feet. Contour interval is 10 feet.



Name: OAKFIELD
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 07' 39.96" N 078° 16' 01.17" W
Caption: Figure 13. Scattered drumlins idicate NE-SW flow of ice during late stage of glaciation. Contour interval is 10 feet.



Name: KNOWLESVILLE
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 10' 49.68" N 078° 15' 35.82" W
Caption: Figure 14. Gravel pit near esker preserves soft-sediment deformation structures (Fig. 15). Contour interval is 10 feet.

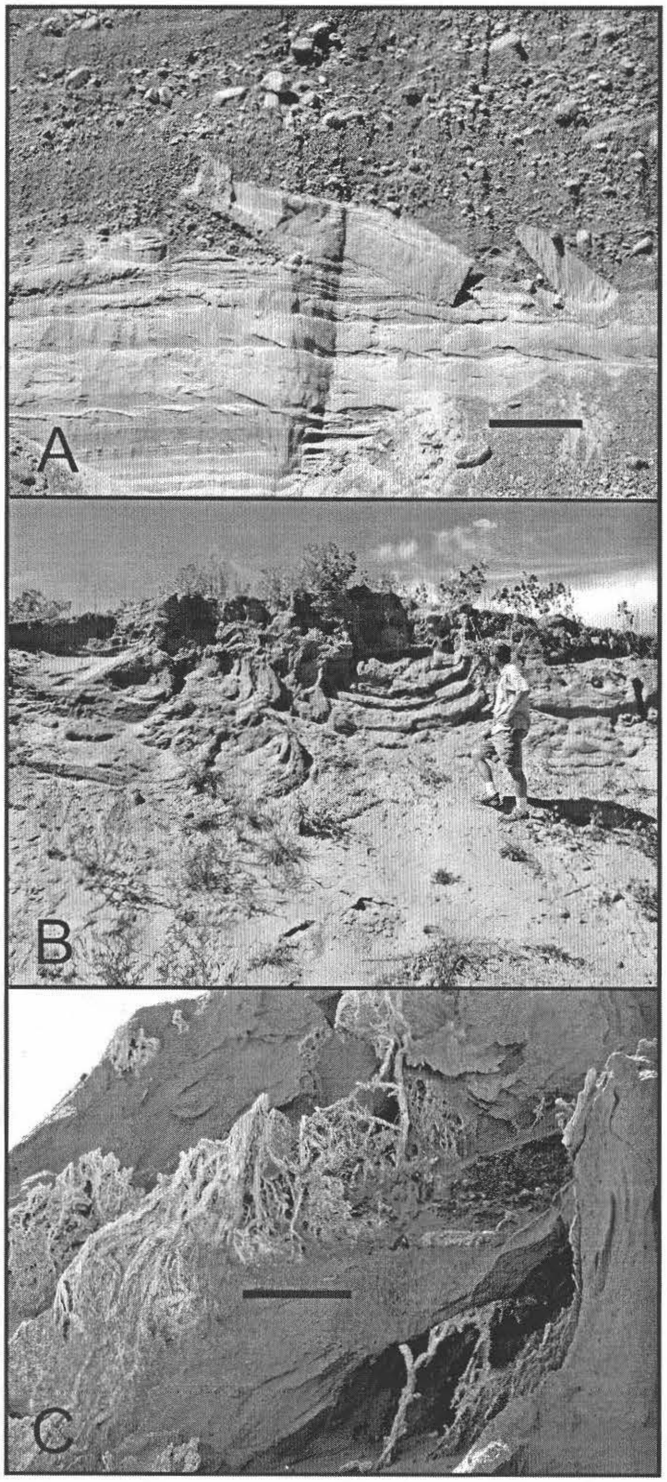
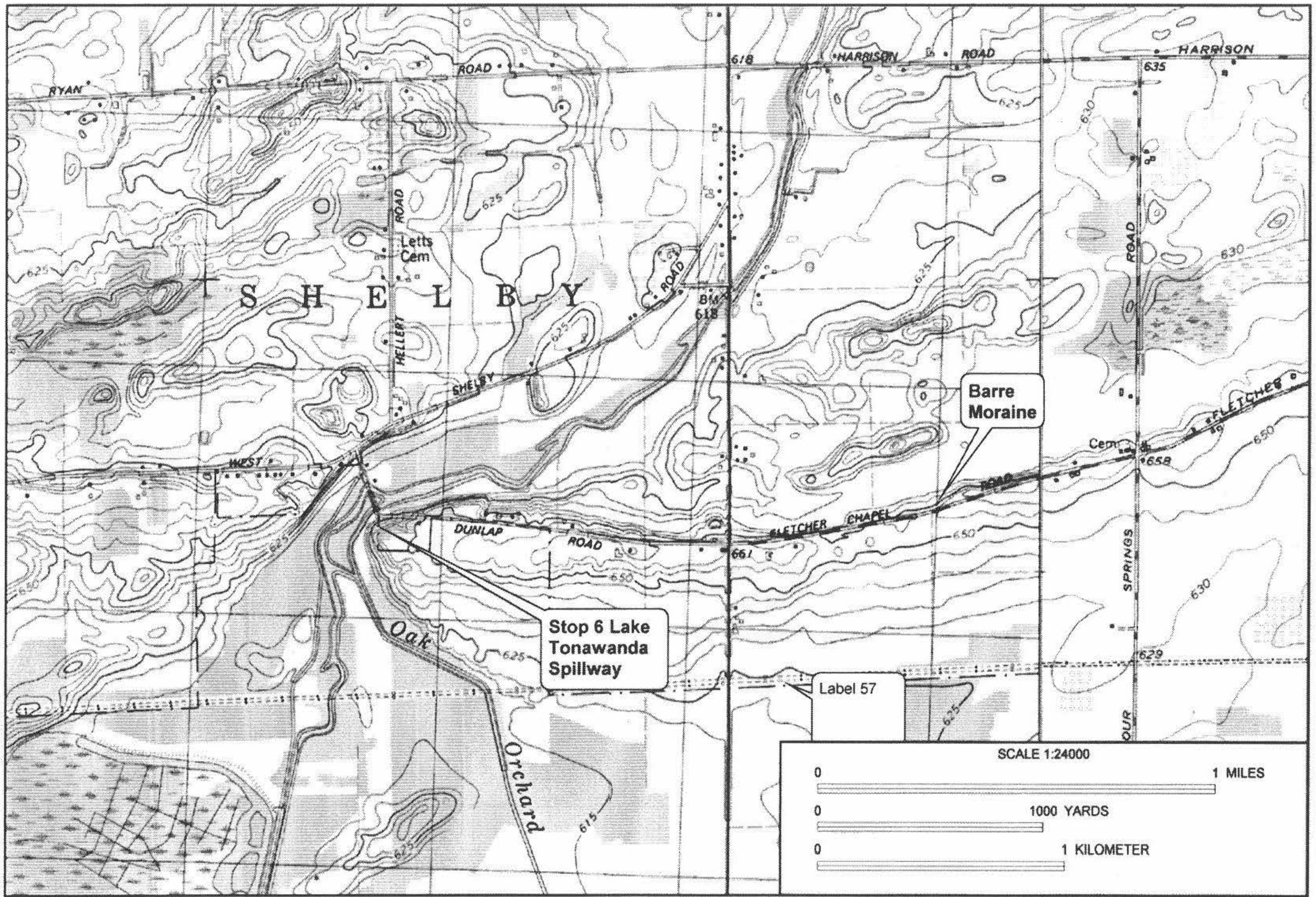
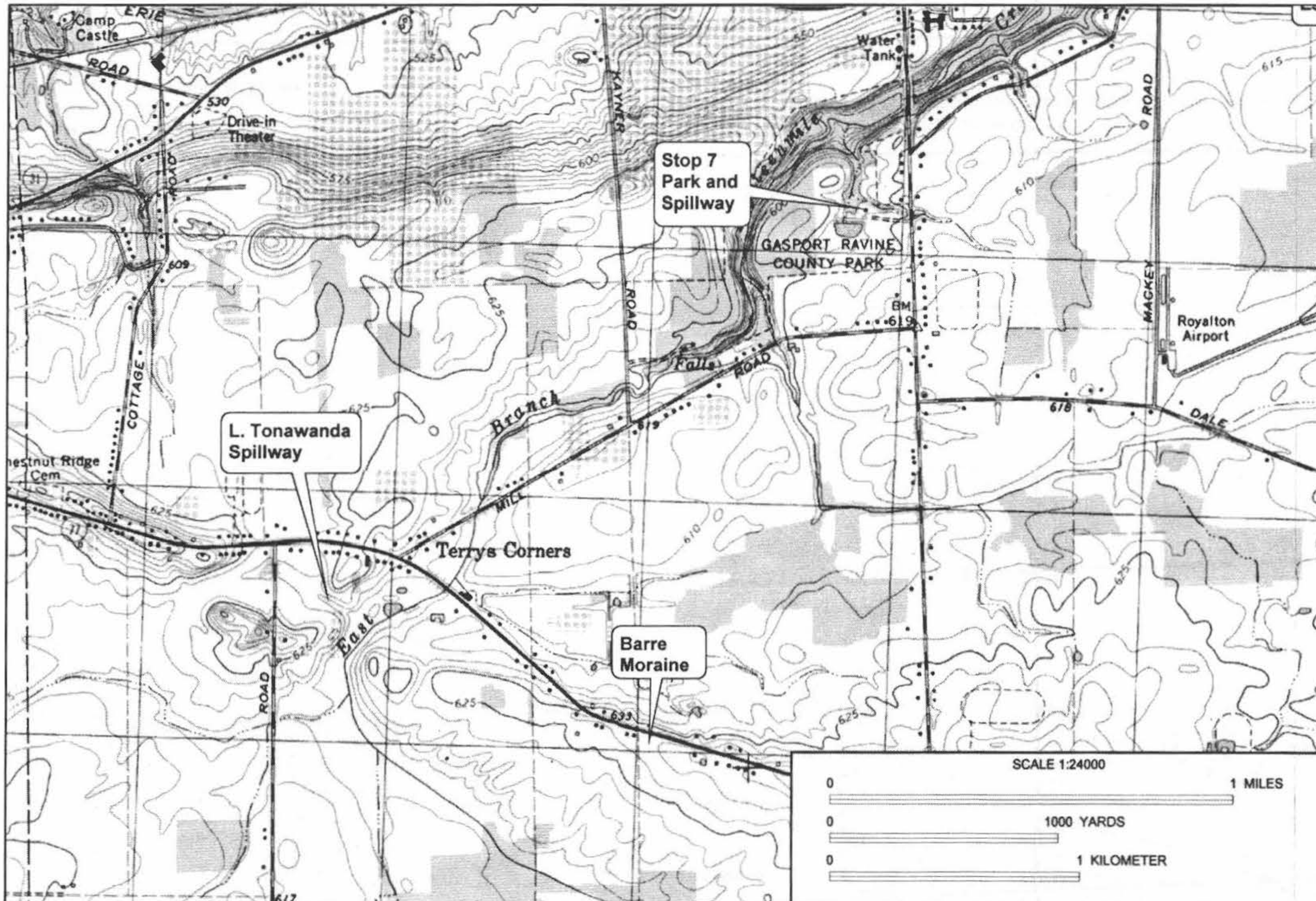


Figure 15. Gravel pit at Stop 4 preserves interesting sedimentary structures indicative of apparent brittle-style deformation (A) as well as soft-sediment deformation, possibly slumping and/or groundwater discharge (B). Root masses replaced by calcium carbonate are aligned along fracture traces(C). Scale: Bars in A and C are 1 foot.



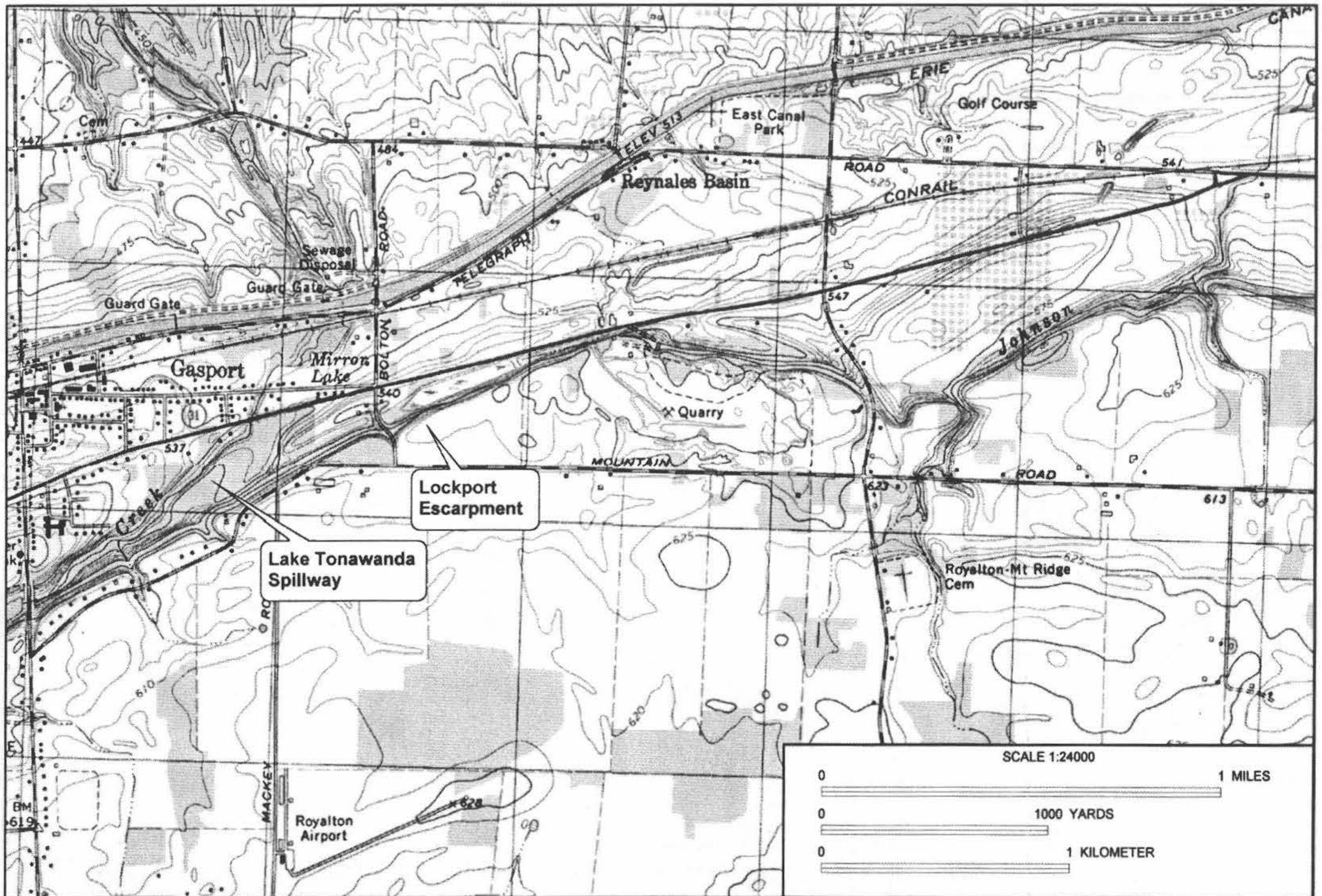
Name: MEDINA
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 09' 37.48" N 078° 23' 38.89" W
Caption: Figure 16. Lake Tonawanda spillway is occupied by Oak Orchard Ck. south of Medina. Contour interval is 10 feet.



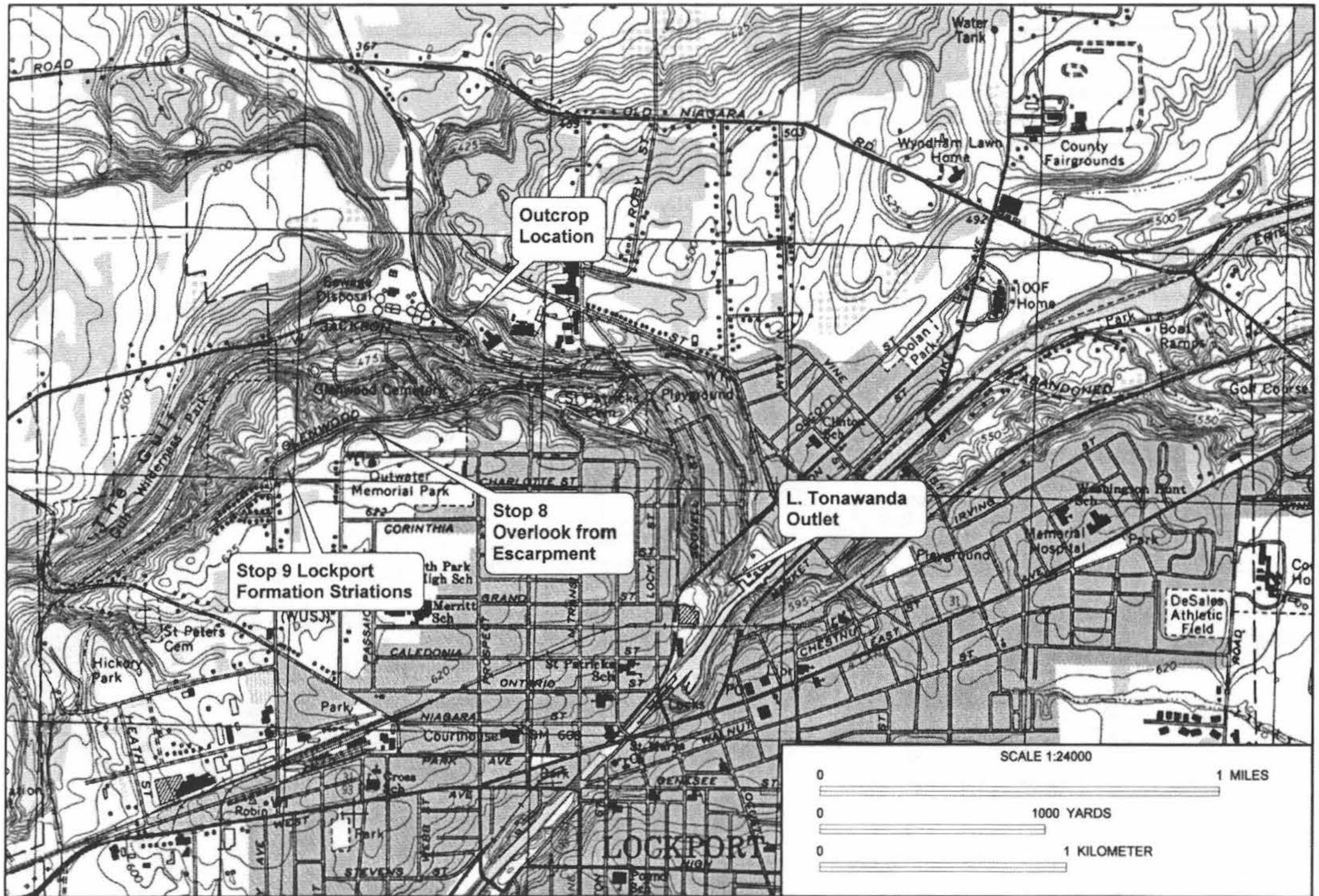
Name: GASPORT
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 10' 39.91" N 078° 35' 14.75" W
Caption: Figure 17. Lake Tonawanda spillway south of Gasport. Contour interval is 10 feet.



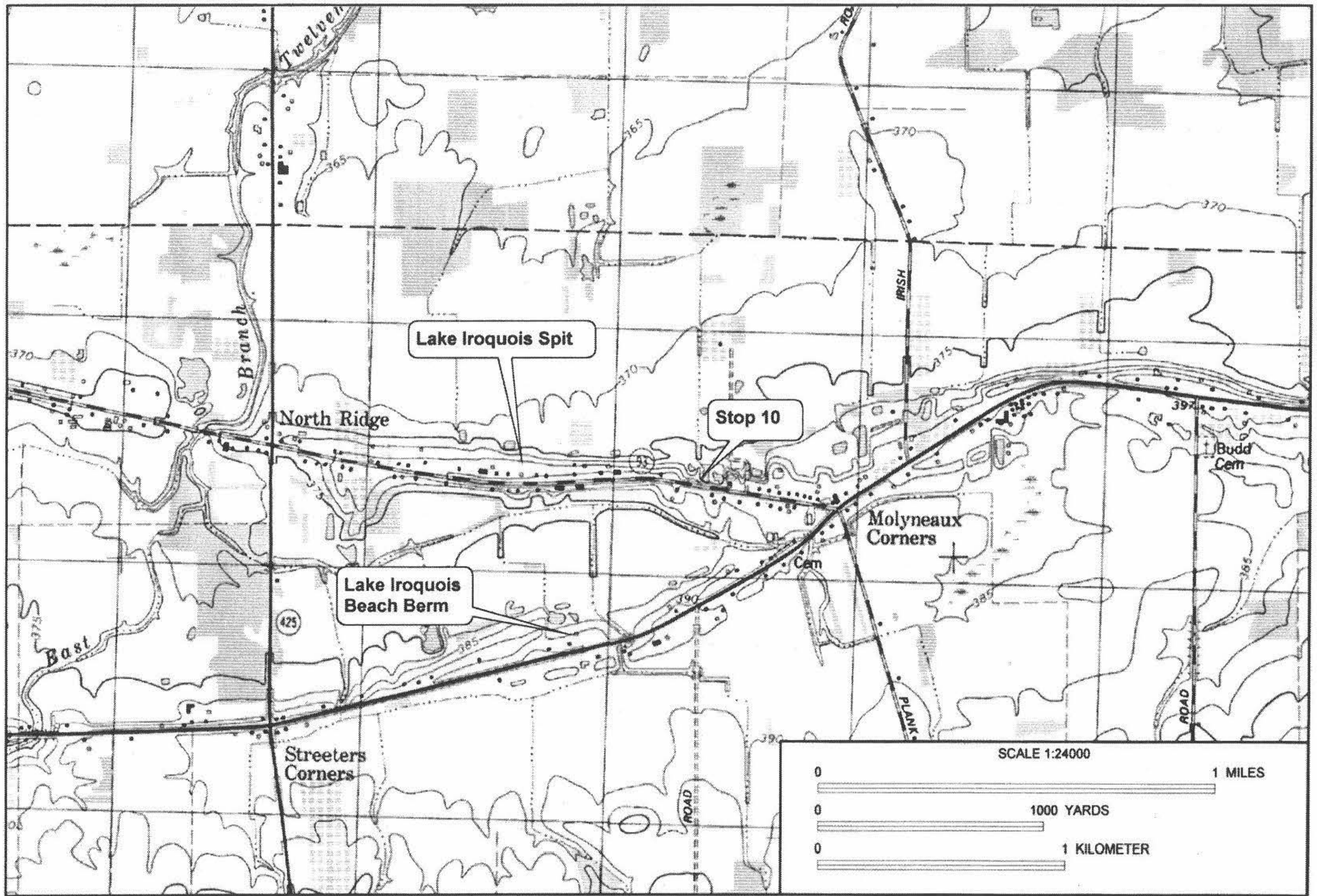
Name: GASPORT
Date: 9/6/2006
Scale: 1 inch equals 2000 feet

Location: 043° 11' 45.55" N 078° 32' 39.07" W
Caption: Figure 18. Lockport Escarpment near Gasport. Contour interval is 5 feet.



Name: LOCKPORT
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 10' 46.97" N 078° 41' 35.35" W
Caption: Figure 19. Escarpment and Lake Tonawanda spillway at Lockport. Contour interval is 10 feet.



Name: CAMBRIA
Date: 9/4/2006
Scale: 1 inch equals 2000 feet

Location: 043° 12' 43.98" N 078° 48' 23.59" W
Caption: Figure 20. Lake Iroquois shoreline features northwest of Lockport. Contour interval is 5 feet.

The Grenville Terrain, Canada
Field Guide

Pete Avery & Joaquín Cortés

October 7, 8, 2006

1 Description of the fieldtrip

This field trip will examine the geology of the Grenville Province along highway 62 between Madoc and Bancroft, Canada during Saturday the 7^{th} and Sunday the 8^{th} of October, with an overnight stay in Bancroft. There are excellent exposures of the progression of the metamorphic grade and deformation linked with the collision of the North America with a Continent now obliterated by the Appalachians. We will visit outcrops of marbles, metabasalts and siliciclastic/pelitic metasediments from facies Greenschist to Upper Amphibolite/ Granulite and associated intrusives (granodiorite/orthogneisses, nepheline–syenites, gabbros and norites). We will also visit the Princess Sodalite Mine, in which sodalite (an extremely rare Na–chlorine feldespatoid) occur.

2 Geographical Location

The area to be covered during this fieldtrip is located in Canada, in the southeast part of the province of Ontario (Figure 1); the town of Bancroft, is located approximately half distance between Ottawa and Toronto. In Figure (2) an extract of the official road Map of Ontario is shown (the distance between Peterborough and Bancroft is about 68 miles/110 Km on highway 28). Bancroft is the upper apex of the triangle defined by the highways 62, 28 and 7.

We will begin the trip driving from the University on Saturday morning arriving to the area about 2–3 p.m. We will have several stops

the Princess Sodalite Mine (see below), located 4 km (about 2 miles) further north following the highway 28, and during the afternoon we will visit several outcrops on the routes 62 and 7. We will return to Buffalo during the evening of the same day.

3 Geological Setting

The term Grenville terrain (*Grenvillian orogenesis*) is a major compressional event involving thrusting toward the craton Superior (Figure 3) and imbrication of the entire crust in a duplex structure, forming an imbricated fan (Figure 4). It was probably the result of the collision at *ca.* 1.1 Ga between the North american craton and a continent to the southeast, now largely obscured by the Appalachian orogen (Ordovician– Silurian ~ 440–410 Ma).

Figure 1: Map of the province of Ontario, Canada.

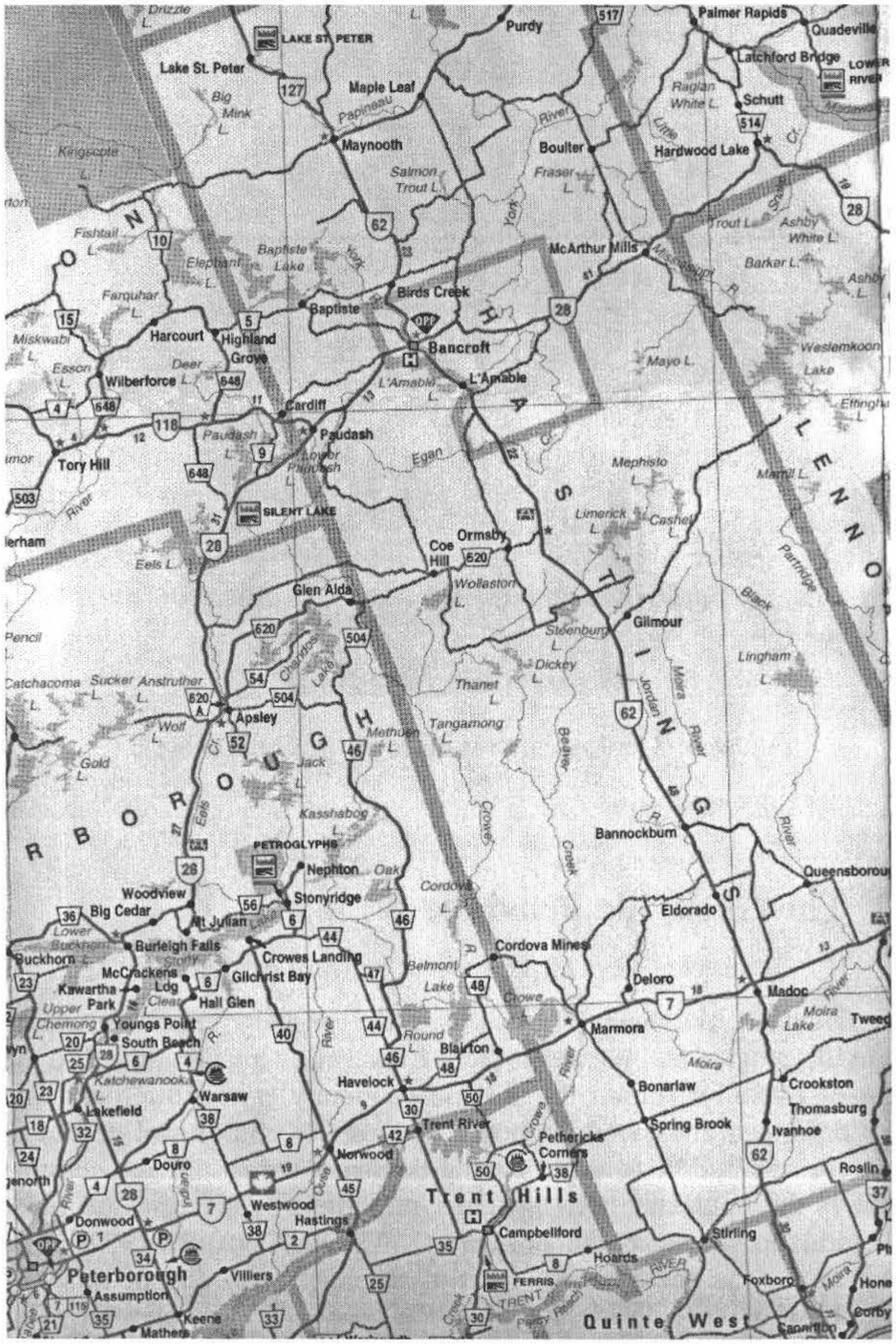


Figure 2: Road Map of the Bancroft Area.

on the highway 28 during the rest of the day to end the day in Bancroft in which we will spend the night. During Sunday morning we will visit

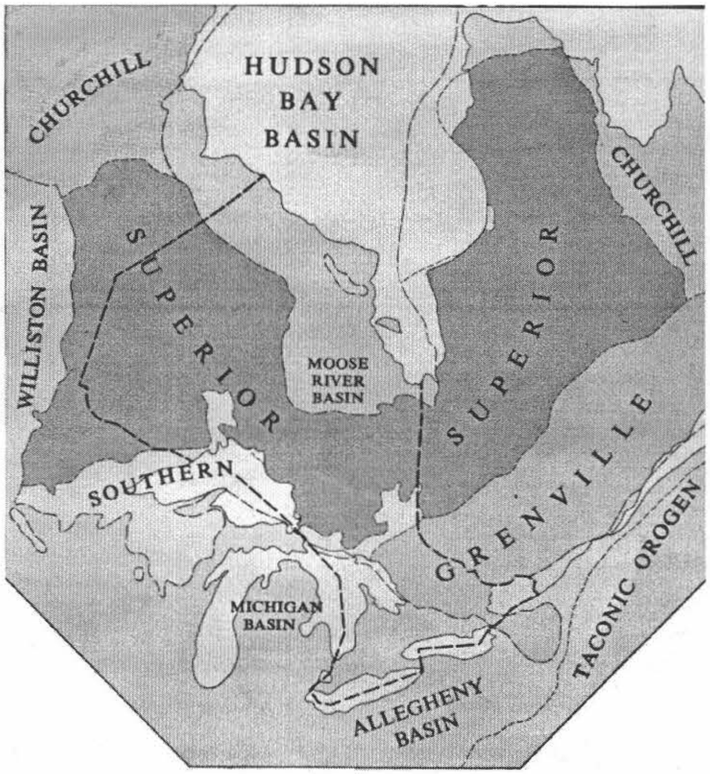


Figure 3: Geological provinces of the Ontario area.

3.1 The Grenville Province

(extracted from http://www.eps.mcgill.ca/~litho/)

The Grenville province (Figure 3) is the primary exposure of the Grenville orogen, and it extends from Lake Huron northeastward to the coast of Labrador (Figure 1). It exposes mainly crystalline rocks that were deformed and metamorphosed at lower to mid-crustal depths in the west, and middle to upper crustal depths in the east. The Grenville Front is the boundary between the Grenville province and the older structural provinces to the northwest (Wynne–Edwards, 1972). The Front marks the northwestern limit of tectonic reworking of rocks of the older provinces during the Grenville orogeny, a complex series of tectonic events generally considered to have taken place between 1300 and 1000 Ma.

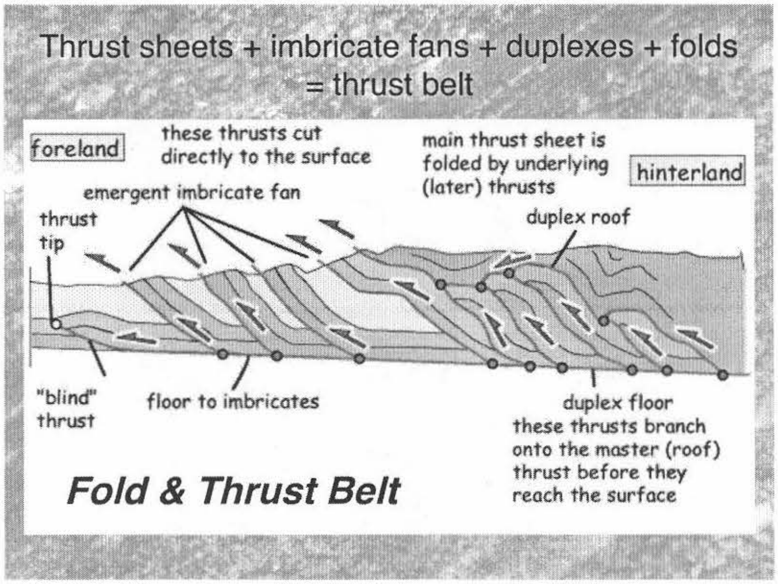


Figure 4: Structure of an orogenesis belt.

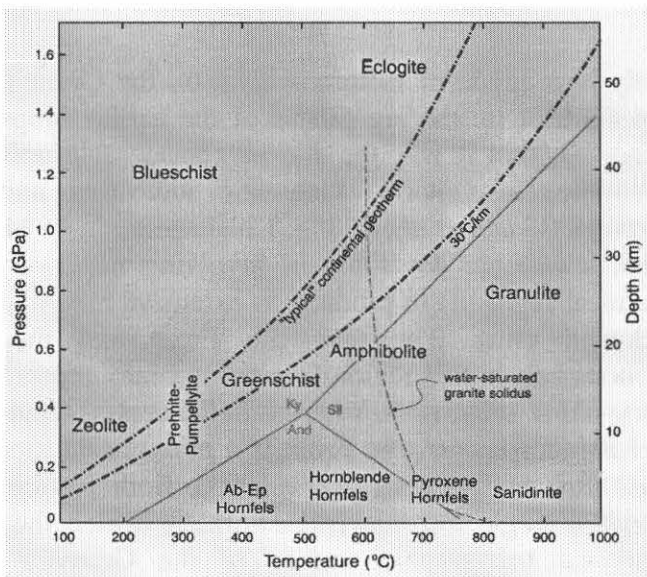


Figure 5: P–T diagram showing the metamorphic facies boundaries.

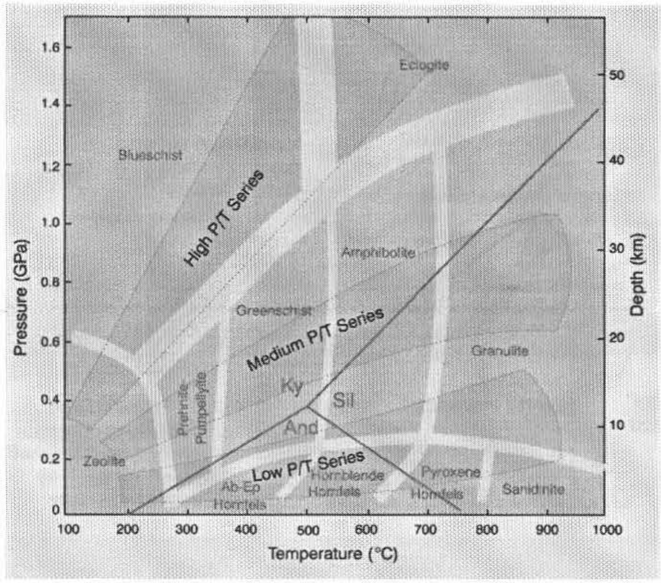


Figure 6: P–T diagram showing the metamorphic series and possible geothermal gradients.

The uniformly high grade of metamorphism on the Grenville side of the Front, in contrast to varying grades of the earlier orogens to the northwest, indicate that it was the site of considerable uplift. In the western Grenville, in Ontario, kinematic indicators and seismic reflection surveys (Green *et al.*, 1988b; Culshaw *et al.*, 1994; J. *et al.*, 1989) suggest a tectonic development involving northwest–directed stacking of crustal segments to produce an extensive mountain belt and an overthickened crust. The Grenville Front and the adjacent parautochthonous belt in this region (see below) may represent a deep crustal, reverse-sense shear zone, whereas in the central Grenville, near the Quebec–Labrador border, the Front is a mid–crustal metamorphic foreland fold/thrust belt (Rivers *et al.*, 1993). Both features may be reconciled with a single tectonic framework, in which the presently exposed east-west topographic surface of the Grenville province represents an oblique section through the crust.

The Grenville province has been subdivided into first–order belts that reflect the tectonic character of the orogen (Rivers *et al.*, 1989). The most northerly, located adjacent to the Grenville Front, is known as the parautochthonous Belt, and consists of units that can be traced into the foreland north of the Front. The more southerly Allochthonous Belt, comprising thrust slices of both monocyclic rocks and older polycyclic pre-Grenvillian units, tectonically overlies the parautochthonous Belt.

It contains most of the relics of recognizable supracrustal rocks and associated plutonic units. In the eastern Grenville province, the Allochthonous Belt can be subdivided into a thrust belt in the north and a syn-to post-tectonic granitoid magmatic belt to the south (Gower *et al.*, 1991). A major period of magmatism late in the Grenville orogeny flooded lower and mid-crustal sections with mafic magma that differentiated into large anorthosite complexes.

Although Grenvillian metamorphism in the Archean rocks makes it clear that significant thickness of transported rocks once overlay the Archean rocks (Childe *et al.*, 1993), only a thin *klippe* of transported material now overlies them, near the Grenville Front. In eastern Quebec, Archean rocks are exposed within the Grenville orogen only in a narrow zone near the Grenville Front. Seismic data make it clear, however, that Archean rocks occur in the subsurface at depths as shallow as 10 km, more than 100 km SE of the Grenville Front, corroborating the thin-skinned model for the geometry of the thrust belt in western Labrador proposed by (Rivers *et al.*, 1993).

3.2 Structure

Based in seismic profiles Green *et al.* (1988a) interpreted the structure of the Grenville province in terms of a simple tectonic model of an imbricated fan–duplex relating with thrust geometry.

For the Grenville front the lower crustal layer was interpreted as Archean crust, extended and thinned during the rupturing of the souther Superior craton and associated development of a south–facing continental margin at 2.4–2.5 Ga. On the basis of this interpretation, the upper crustal layer is either a thick wedge of highly deformed Huronian strata and displaced Archean basement or, more likely, the allocthonous mass.

During the initial stages of the Grenvillian orogeny, perhaps coeval with 1.25–1.30 Ga island–arc magmatism, there was a progressive stacking of microterranes against the Superior craton, depressing the rocks along the margin. Transport of microterranes would have occurred from southeast to northwest along ductile shear zones. The seismic profiles reveals the geometry at depth of at least two thrust planes (zones; Figure 9), that might be active at the time.

During a later phase of Grenvillian tectonism (1.15 Ga), intense deformation occurred near the stable craton. Thrusting cut deep into the crust and possibly the upper mantle, causing ductile imbrication of a wide band of rocks under high pressure (800–1000 MPa) and high temperature (600–800 °C) conditions; rocks buried deeply by the earlier stacking epizode were thrust back up to shallow depths. At this time much of the southeast–dipping fabric, clearly seen in surface outcrop was imposed. Although rocks would have deformed and metamorphosed by burial conditions it was proposed by Green *et al.* (1988a) that this thrusting under ductile conditions that created most of the mylonitic zones. In this region of the Grenville orogen the major component of reverse displacement was concentrated at the Grenville front, where abrupt increases in metamorphic grade occurred. Within the Grenville province, thrusting was largely responsible for progressive uplifting deeper level (higher metamorphic grade) rocks. The sharp increase in deformation of Superior and Southern province rocks at the Grenville front suggest that the thick lithosphere underlying most part of the craton and its southern margin acted as a rigid buttress throughout Grenvillian tectonism.

The thick crust now observed under the Grenville zone might have resulted from initial stacking of microterranes, or it might have been created during the later stages of ductile thrusting. If this is correct, the combining current 50 km crustal thickness with the 20 km of the material that must have been removed to exposed the lower crust (amphibolite–granulite facies rocks) yields a thickness of 70 km crust, comparable with the actual Himalayas.

3.3 Metamorphism in the Grenville province

Based in the maximum metamorphic conditions described in the previous section (800–1000 MPa and 600–800 $^{\infty}$C), it can be shown (Figure 5) that the maximum metamorphic conditions reached the boundary between the Upper Amphibolite –Granulite facies. Considering the high geothermal gradient consistent with a thick crust equivalent to the actual Himalayas the progressive metamorphism should have followed the path of medium P/T series (Barrovian) from Green Schists to Granulite facies (Kyanite ∅ Sillimanite transition). The subsequent thermal event due to intrussion of gabbro and norite are the causes of punctual areas of contact metamorphisms over–imposed on the regional metamorphism rocks. The interchange of fluids allowed the developing of restricted skarn zones.

3.4 Mineral resources

(extracted from: http://www.bancroftontario.com)

Due to the unique nature of the rocks exposed in the area of Bancroft, and the particular metamorphic processes that occurred during the Grenvillian orogenesis, a group of unique minerals can be found in the area.

Bancroft itself is regarded as the "Mineral Capital of Canada" because of the well known variety and quality of mineral species which occur here. A list of minerals that have been mined in the Bancroft area include: corundum, feldspar, uranium, graphite, iron, mica, lead, gold, molybdenite, apatite, beryl, fluorite, talc and sodalite. Additionally, nepheline syenite, marble, granite were also mined. Mining in most cases was carried out on a limited scale, between 1880 and 1935, and was largely confined to open cuts and quarries.

The Princess Sodalite Mine is an example of this kind of production. It is a quarry style mine with a century long history, which is still being worked at present. The Figure (7) shows the main seam of blue Sodalite $(Na_4(Al_3Si_3)O_{12}Cl)$, surrounded by the grey Nepheline Syenite in which it has formed.

Figure 7: A view of the main quarry.

During her visit to the 1901 World's Fair in Buffalo, New York, the Edward VII, then the Princess of Wales was captivated by the beauty of a gift of Bancroft sodalite. Because of this, arrangements were made to quarry enough to decorate her London residence, Marlborough House. In 1906, these hillsides were worked by the owners of the property for this mineral and shortly after, 130 tonnes of the rock were shipped to England to be used as a decorative stone in the Princess' royal home. And that is from where the name "Princess Sodalite Mine" comes. Today, sodalite is commonly used in the creation of jewelry and as an ornamental stone. It is also the mineral most people associate with the uniqueness of the Bancroft region's mineral heritage.

The pieces of sodalite you will find in the rock farm vary in the intensity of blue. Look for pieces that are a deep blue colour. Surrounding the mineral you will usually find a grayish–white mineral called nepheline. Nepheline syenite gneiss is the main rock which surrounds you (Figure 8).

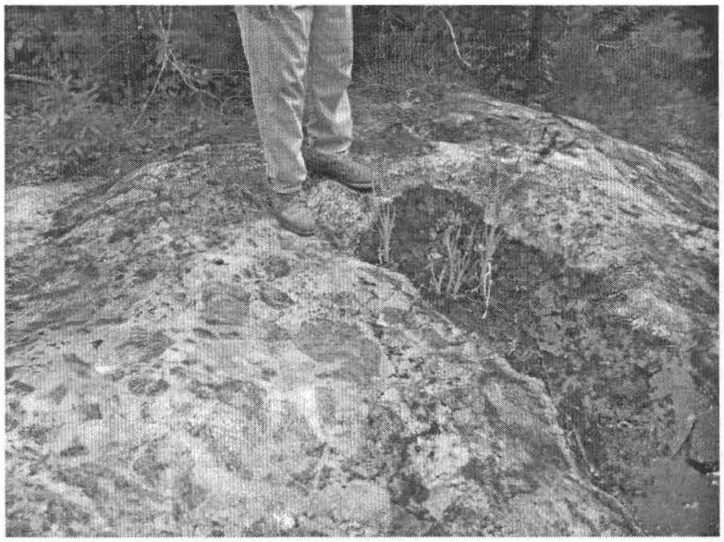


Figure 8: An outcrop of the Nepheline–Syenite in the Princess Sodalite Mine.

Since 1961, when the site was first opened to collectors, visitors have been made welcome. Andy Christie, the current owner, will do his best to make your visit a worthwhile experience. Good collecting!

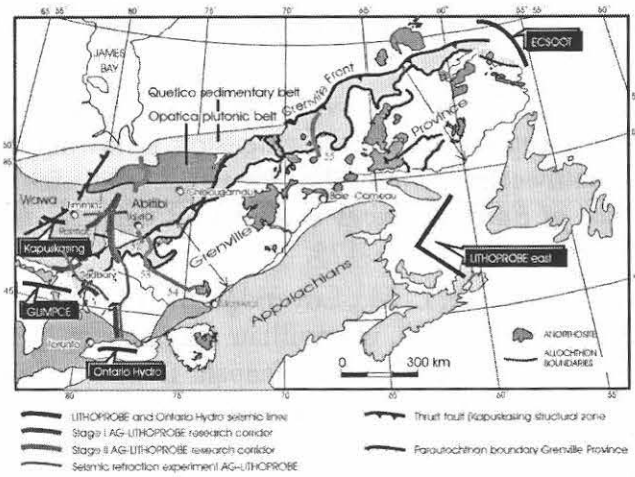


Figure 9: Tectonic Settings of the Grenville province

Sources of the Figures:

Figure 1: © 2000. Her majesty the Queen in right of Canada.Natural Resources. Canada.

Figure 2: The Official Road Map of Ontario. Canada.

Figure 3: From the Geological Map of Ontario. Ontario Geological Survey.

Figure 4: Taken from http://earth.leeds.ac.uk/learnstructure/index.htm

Figures 5, 6 : From Winter, J. D., 2001. An Introduction to Igneous and Metamorphic Petrology. Prentice Hall, 697 p.

Photographs 7, 8: Taken from Alan Goldstein's Site Photo Gallery, http://www.mindat.org/sitegallery.php?u=2749

Figure 9: Taken from: http://www.eps.mcgill.ca/~litho/structure.html

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