92nd ANNUAL
NEW YORK STATE GEOLOGICAL ASSOCIATION
FIELD CONFERENCE

Field Trip Guidebook

NYSGA

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Edited by
Gary S. Solar and
David W. Valentino
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PREFACE

Celebrate! We are fortunate that the annual NYSGA Field Conference has returned!
We had high expectations for the 91st NYSGA to be held in 2020 when planning began in 2018. Many terrific NYSGA conferences had been hosted in the Buffalo, New York area over the years, and this one looked to be no different. Many potential trip leaders were identified, both 'classic' and new. Many of them agreed to run trips as soon as asked, and some offered without solicitation. Some other trip leaders needed some nudging (😉). New trips and leaders were identified. There was some excitement. Professionals and students alike were showing great interest. Fifteen trips were planned. Alas, it was not to be.

We were full steam ahead in July 2020 in organizing the conference for the first week of October that year. However, it became increasingly clear over that summer that the logistics and risks of running a field conference at that time were proving difficult and unwise, respectively. Many leaders were understandably worried for their own safety and for the safety of the trip participants – being at outcrops, naturally huddled closely to observe what the leaders were explaining, etc. Some leaders experienced resistance from landowners as well. The writing was on the outcrop wall – with great pain the decision came to postpone the 92nd NYSGA Field Conference to 2021 (and hope for the best). After having to cancel the 2020 conference due to the COVID-19 Pandemic, the future was uncertain to say the least.
We then set our sights on 2021, and here we are. The number of trips is less than hoped, but for many reasons several trips planned originally just were not going to be feasible.

HOWEVER, WE HAVE TRIPS FOR 2021, THE 92nd FIELD CONFERENCE!

Thank you to the 2021 NYSGA Trip Leaders. Without you, naturally we would not be able to run the conference, and for that we are all truly grateful. Thank you for your time and experience. For those of you who had intended to run trips, but could not for safety concerns, we thank you too, because without your example over the years we would not have trips now. Thanks to the NYSGA Advisory Board and web master. Without you there is no NYSGA. And finally, thank you to the trip participants, and the future users of these field guides. Again, without you, it is just the leaders talking to themselves about their passions in the field (as usual). Are there challenges in running a field conference during a global pandemic? Absolutely. Are we ready for this challenge? Yes.

"Be safe" usually means "in the field" for field scientists, but now it takes on new facets of meaning. We are truly fortunate to have these trips for the 92nd Field Conference. Let's get out in the field, and be safe doing it. Perhaps the future will see a return to "normal" for the field conferences out there, but there *are* conferences out there.

Gary Solar, President
NYSGA 92nd Annual Field Conference

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MARINE CRISIS EVENTS AND FOSSILS OF THE LATE DEVONIAN APPALACHIAN BASIN DELTAIC SEQUENCE OF WESTERN NEW YORK

KIMBERLY C. MEEHAN
University at Buffalo, Department of Geology, Buffalo, New York 14260

JAMES T. BOYLE
University at Buffalo, Department of Geology, Buffalo, New York 14260

INTRODUCTION

The Devonian Period experienced many changes within the environmental and biological realms and terminated with one of the largest mass extinctions of the Phanerozoic. The Appalachian Basin of western New York contains several alternating successions of gray and black shale, many of which mark marine crises culmination in the end-Devonian mass extinction. The causes for the formation (e.g., anoxia, flooding, biological production changes) of the black shale sequences has long been debated within the basin. This period was characterized by major changes in both the terrestrial and marine biospheres and terminated with one of the largest mass extinctions of the Phanerozoic Era (Kaiser et al., 2016). Middle Devonian terrestrial environments saw tremendous increases in biomass and complexity with the evolution of vascular and seed-producing plants, trees, and the formation of deeply weathered and thicker soils (Beerbower et al., 1992). The Devonian marine realm has long been suspected to have been heavily affected by bottom water anoxia, enhanced organic carbon burial rates, dramatic shifts in primary production, and an extended biotic crisis. The biotic crises and related marine extinction events have been attributed to many factors including bolide impacts (McLaren, 1982), tectonism and climate change (Ettensohn et al., 1988), oceanic overturn and/or euxinic conditions (Kelly et al., 2019; Haddad et al., 2016; Boyer et al., 2021), cold water oceans and dysaerobic conditions (Copper, 1986), marine ecosystem collapse (McGhee, 2013), eustatic change (Johnson and Sandberg, 1988), and more recently linking the marine phenomena to coeval developments in the terrestrial realm (Algeo et al., 1995; Algeo and Scheckler, 1998; Algeo et al., 2000).

Algeo et al. (1995) presented the hypothesis that the Middle-to-Late Devonian marine biotic crisis and mass extinction of benthic communities were precipitated by the evolutionary development of vascular land plants; terrestrial floras appeared in the Middle Ordovician, and these land plants were small, either non-rooted or shallowly rooted, and ecologically limited to moist lowland habitats (Cascales-Miñana, 2016). Evolutionary innovations of these floras in the Devonian allowed them to interact with substrates and strongly influence weathering processes, hydrologic cycling that would have changed the amount of run-off and peak discharge (Schumm, 1977; Algeo and Scheckler, 1998), and has been suspected by some researchers to have resulted in geochemical fluxes and the formation of carbon-rich black shale beds within the Appalachian Basin. While flooding events in the Appalachian Basin have received much attention recently (Kelly et al., 2019; Haddad et al., 2016; Lash, 2019; Bartlett et al., 2021 (in press)), by no means is there a consensus on the causes of formation of black shale sequences and other major shifts in lithology during this global event (see Kaiser et al., 2016).

Over 100 years of research of the Devonian Appalachian Basin has led to the construction of one of the most detailed litho-stratigraphic frameworks of a Paleozoic foreland system that has allowed for detailed interpretations of sea level history and shifts in sedimentation rates (Dana, 1894; Wanless, 1947; Brett and Baird, 1986; House and Kirchgasser, 1993; Brett, 1995; Ver Straeten and Brett, 1995; Brett et al., 2011; Ver Straeten et al., 2011). As a result, a tremendous amount of geochemical proxy data from these gray and black shale sequences has shown that the perception that the Devonian black shales were deposited under anoxic conditions holds true, thus far, for only one black shale unit (Werne et al., 2002), the Oatka Creek Shale, and that intervals of terrestrial fresh water flux were not as prevalent as previously thought and that primary production plays a strong role in many, if not all black shale beds of the Appalachian Basin (Arthur and Sageman, 2005).

More recent research on intercalcated deposits described in ancient and modern marine settings (e.g., Cretaceous Western Interior Seaway and Eastern Atlantic Ocean, Pleistocene North Atlantic, Neogene Mediterranean, modern Black Sea) has focused on eutrophic conditions and the effects of organic input, climate variations, primary production changes and seasonal riverine input and has made clear that the formation of carbonaceous and organic-carbon-deficient layers is anything but straightforward. The same is true of investigations attempting to derive the mechanisms behind the gray and black sequences in the Devonian Appalachian Basin (Werne et al., 2002; Sageman et al., 2003; Arthur and Sageman, 2005; Ver Straeten et al., 2011; Wilson and Schieber, 2015; Kelly et al., 2019; Smith et al., 2019; Haddad et al., 2016; Lash, 2019; Boyer et al., 2021; Bartlett et al., 2021 (in press)).

GEOLOGIC BACKGROUND

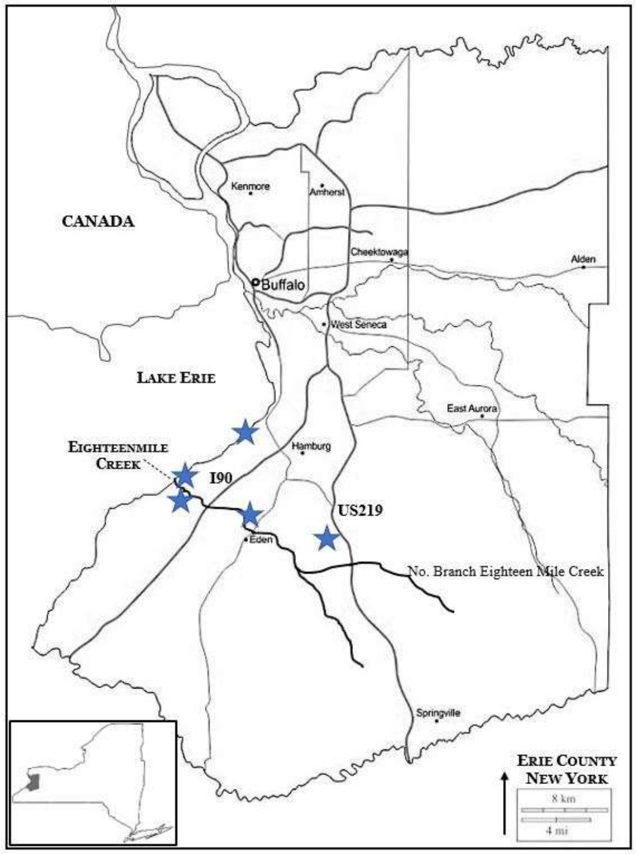

Figure 1: Inset map: Location of Erie Count, New York; Detail map: stars denote locations of field trip stops in and around Eighteenmile Creek, Erie County, NY.

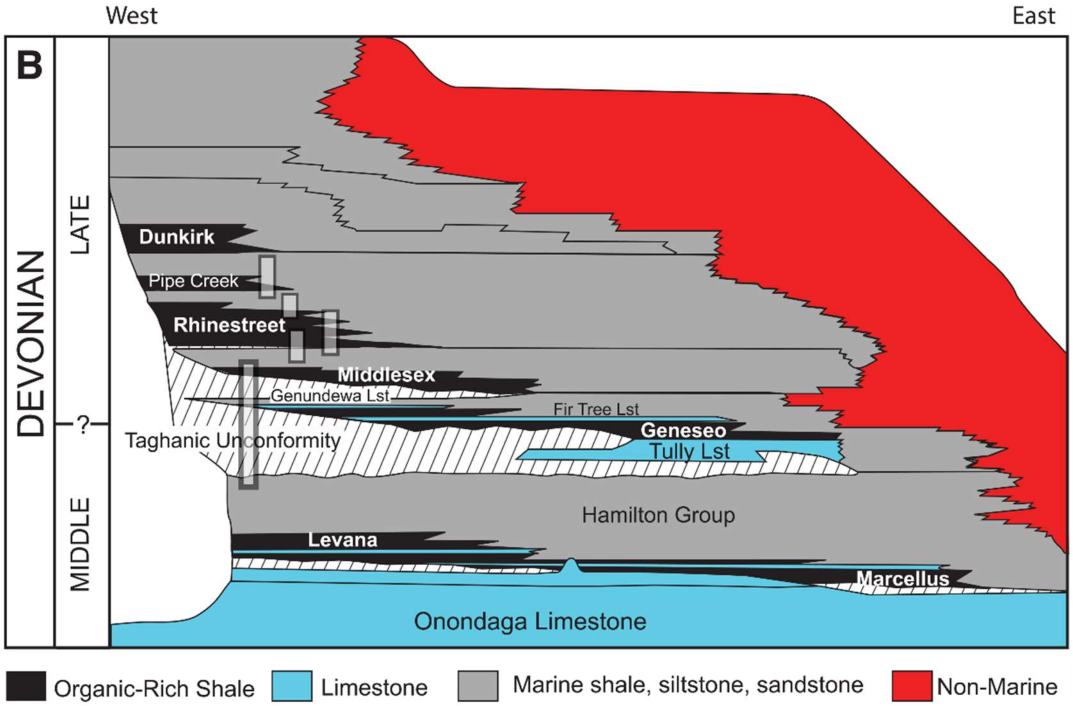

Figure 2: Hyper-generalized deltaic sequences of western New York from Smith et al., (2019). White highlights denote general portions covered along field trip stops.

The nearly one hundred years of research on the basic lithology of the deposits in Erie County, New York were summarized and compiled by Buehler and Tesmer (1963). More recently, work attempting to constrain the ecological conditions (e.g., freshwater inundation, primary production, and oxygen conditions) and relative sea level change have been published (Werne, 2002; Arthur and Sageman, 2005, Ver Straeten et al., 2011; Wilson and Schieber, 2015; Hupp and Weislogel, 2018; Lash 2016, 2019; Kelley et al., 2019; Fritz, 2019; Haddad et al., 2016; Smith et al., 2019; Bartlett et al., 2021 (in press); Li et al., 2021). These distal marine shales recorded relative sea level changes in alternating beds of deeper water, dark gray to black organic-rich shale, suspected to be deposited in dysoxic or anoxic bottom waters and light gray organic-poor shale, deposited in shallower, more oxygenated water (Ver Straeten et al., 2011; Fritz 2019).

<u>Lithology</u>

Givetian - Hamilton Group

Three units of the lower Moscow Formation, the Deep Run, Menteth, and Kashong members, present in the Finger Lakes region, are absent at Eighteenmile Creek due to non-deposition or erosion (Maletz, 2008). The Moscow Formation is fairly thin at the shores of Lake Erie, measuring about 11.5 feet, but increases in thickness to ca. 55 feet in the eastern part of Erie County (Buehler and Tesmer, 1963). At Eighteenmile Creek, the Moscow Formation starts with the Tichenor Limestone Member, overlain by the Windom Shale (Deep Run Shale, Menteth, and Kashong Shale are absent). A very thin upper Moscow Formation section is present comprising about 4.3 m of Windom Shale, despite its thinness, the Windom Shale is more stratigraphically complete (condensed), and is divided into a number of persistent faunal assemblage zones (Grabau, 1898; Brett, 1974; Brett and Baird, 1986). The top of the Moscow Formation is defined at the base of the Tully Limestone in the Seneca Lake region (south), but this unit is not represented in Erie County. Instead the Leicester Pyrite Bed and the North Evans Limestone represent the local base of the Genesee Group (Maletz, 2008).

The Tichenor Member, historically referred to as the 'encrinal limestone' (Grabau, 1898), is a reworked layer (~30- 60 cm) of medium gray, buff to rusty weathering calcarenite that forms a single massive unit at the shores of Lake Erie (Brett, 1974). It consists predominantly of echinoderm skeletal debris with a mixture of lime mud and sparry calcite cements and the lower surface of the unit hosts a distinct disconformity. Fossils from the upper part of the Wanakah Shale are often found incorporated into the base of the Tichenor Limestone. The top of the Tichenor Limestone is a discontinuity surface with a locally developed hardground. This surface shows a relief of up to 20 cm. Locally the upper surface of the Tichenor Limestone exhibits dark, possibly phosphatic staining, anchor-faceted fossils and directly cemented crinoid holdfasts, bryozoans, and auloporid corals (Brett, 1974).

The Windom Shale Member is only 4.3 m thick in the Eighteenmile Creek section along the shores of Lake Erie, but is fairly complete and a number of faunal associations can be differentiated. The Windom Shale generally thickens to the east and north but with increasing stratigraphic truncation of the upper portion of the unit to the north (Brett and Baird, 1986). The basal Windom Shale Member consists of about 50 cm of very soft, medium gray, richly fossiliferous shale that contains a brachiopod rich assemblage, similar to the middle Wanakah Shale (exposed only downstream of the crossing of Route 5 and Eighteenmile Creek). Abruptly overlying the brachiopod rich shale is a 10-15 cm thick band of soft, blocky crumbly weathering gray mudstone containing an exceedingly rich and unique fossil assemblage, termed the Bay View Coral Bed (Baird and Brett, 1983). A prominent and traceable carbonate concretion band a few centimeters below the contact with the Genesee Group is consistently found along Eighteenmile Creek. However, it alternates between being expressed as isolated concretions and a semi-continuous limestone band. The Windom Shale is sharply and disconformably overlain by the North Evans Limestone of the Genesee Group.

Frasnian – Genesee Group

The North Evans Limestone Member consists of lenses of <2-18 cm thick of bioclastic limestone, with sedimentary features like cross lamination, intraclasts, and abundant fossils. This limestone is a classic example of an erosional lag deposit or bone bed (Maletz, 2008).This unit was important in the development of conodont biostratigraphy and has been found to be composed of 50% conodonts by weight in some samples (Bryant, 1921)**.** It was deposited in a dysaerobic environment upslope of and coeval to the lower part of the Genundewa Formation and represent the erosional remnants of the Penn Yan Shale (largely absent in western New York, but see below) and the uppermost Windom Shale (Brett and Baird, 1986). This lag deposit thus covers six conodont chronozones and spans the Givetian-Frasnian boundary. The upper surface of the North Evans Limestone is most commonly expressed as an undulating unconformable contact with the Genundewa Limestone.

A thin (<5 cm) black shale unit is intermittently exposed between the North Evans Limestone and the overlying Genundewa Limestone along Eighteenmile Creek. This unit is probably a local remnant of the Penn Yan Shale Member that is more consistently present further east (Baird and Brett, 1986). However, given its spotty occurrence this correlation is tentative (Baird et al., 2006).

The Genundewa Limestone Member is a thin limestone unit (~10 – 20 cm) that marks a break in sedimentation associated with a regression (for detail see: Baird et al., 2006). The unit's base appears undulatory and erosional in western New York and is composed primarily of the problematic fossil *Styliolina* (Baird et al., 2006; Baird and Brett, 1986). This unit is relatively thin in Eighteeenmile Creek compared to exposures further east and south (Baird et al., 2006) as deposition in the local region started later. The lower Genundewa Limestone beds found further east are represented, in part, by the North Evans Limestone. Internally the Genundewa shows evidence of large burrows, hummocky cross-stratification, and soft sediment deformation structures. The unit becomes progressively more clastic and grades into the overlying West River Shale.

The West River Shale Member is ~2.5 m thick at the Lake Erie shore. It consists of dark gray shale with beds of argillaceous siltstone. These siltstone bands become more common to the east as the unit thickens (de Witt and Colton, 1978). The West River Shale has a conformable contact with the underlying Genundewa Limestone and an abrupt contact with the overlying Middlesex Shale of the Sonyea Group. The unit is finely laminated and darker at the base and grades upward into chippy grey calcareous shales.

Sonyea Group

The Sonyea Group is differentiated into the Middlesex and the Cashaqua Formations in the Lake Erie shore sections of Erie County. It consists mostly of dark gray to black shales and is about 12 m feet thick (David et al., 2004). The Middlesex Formation consists of black shales and is approximately 2.9 m thick in Erie County, New York (Sutton et al, 1970; Schieber, 1999; Prevatte, 2020). Freshly broken surfaces of this unit often smell of natural gas and an 'oil slick' is also often apparent (David et al., 2004). A few gray bands and concretionary layers are present in the unit.

The Cashaqua Formation is a gray shale, approximately ~9.1 m in thickness, with an abundance of often flattened ellipsoidal limestone nodule bands, and a few thin layers of black shale (Sutton et al., 1970; Prevatte, 2020). The nodule bands that are prevalent in this unit and in the overlying Rhinestreet are interpreted to represent times of exceptionally reduced rates of deposition and probably formed within a few meters of sediment-water interface (Lash and Blood, 2004a). The Cashaqua Formation grades eastward into a thickening sequence of siltstone and silty shale, and is part of a common turbidite facies of the Catskill Delta (Wilson and Schieber, 2015). This deposit is sharply overlain by the organic-rich Rhinestreet Formation

(Sageman et al., 2003) which marks the onset of the West Falls Group. The Sonyea Group is overlain by the West Falls Group comprised of the Rhinestreet, Angola, and Java Formations. The Java Formation itself contains the Pipe Creek and Hanover units.

West Fall Group

The West Falls Group is represented by two shale members in western Erie County: Rhinestreet and Angola Formations (Sutton, 1985). The Rhinestreet Formation, the basal shale unit of the West Falls Group, is a thick, mostly fissile, black shale that thickens rapidly from west to east (~30 m) and interfingers with, and grades into the overlying gray Angola shale (Wilson and Schieber, 2015) with a distinct oil content (Jaffe, 1950). The petroliferous Rhinestreet Formation, a heavily fractured black and gray shale, at its base along the Eighteenmile Creek section contains a total organic carbon content of 8.09% which diminishes upward in the column to 2.3 % at the Angola shale contact (Sageman et al., 2003). The Rhinestreet Formation contains a few thin gray siltstone beds and thin-bedded argillaceous limestones and when freshly broken smells distinctly of natural gas (Lash and Blood, 2004b). Carbonate concretions are common at many levels and often contain pyrite, but macrofossils are rare. The limestone concretions may be septarian limestones with veins filled with calcite, dolomite, albite and siderite. Distinct large (>2 m diameter) carbonate concretions, commonly referred to as the scraggy layer, marks the transitional zone between the upper-most black Rhinestreet Formation and the gray, fissile Angola Formation (Lash and Blood, 2004a; Wilson and Schieber, 2015).

Angola Formation is more than 67 m thick and alternates between intervals of gray and black shales with impure limestones and calcareous siltstone beds. It is characterized by about 50 distinct beds containing calcareous concretions and limestone nodules of up to 1 m in diameter (Lash, 2016). This unit is relatively barren of macrofossils.

Java Group

At the base of the overlying Java Group the thin, black Pipe Creek Formation (~1m), marks the onset of the Lower Kellwasser Event (Over, 1997, Bush et al., 2015; Kelley et al., 2019; Haddad et al., 2016; Boyer et al., 2021). The Pipe Creek Formation is a persistent, organic-rich black shale throughout its lateral extent (Lash, 2016; Wilson and Schieber, 2015) that intertongues its adjacent shale formations. This shale contains prolific thin and oblate carbonate tubules reminiscent of pipe systems from mud volcanoes and methane cold-seep structures, however, the true origins of the fluids' origination has yet to be determined as a vast majority of the formation lies below Lake Erie and is not well exposed in the area.

The Hanover Formation overlies the Pipe Creek Shale, and is a gray shale with some interbedded black shale beds. The Hanover also thickens to the east grading into silty shale, siltstone, and sandstone (Over et al., 1997; Over, 2002). The total organic carbon content rises from <1 wt.% in the Angola to an average of 4 wt.% in the Pipe Creek, and then falls back to <1 wt.% in the Hanover Formation (Sageman et al., 2003).

PALEONTOLOGY

Macrofossils

The macro-paleontological record within this deltaic complex is superb, thoroughly investigated (e.g., Sutton et al., 1970; Thayer, 1974; McGhee and Sutton, 1981; Sutton and McGhee, 1985; Over 1997), though considered mostly sparse in western New York, and reflects a series of marine crises associated with black-shale deposits throughout the mid to late Devonian (McGhee, 1996). In Erie County, some of the most prolific macrofossil beds are found within the Wanakah Shale, Tichenor Limestone, and through the majority of beds within the Windom Shale Member (e.g., Bayview Coral Bed, Smoke Creek Trilobite Bed. Penn Dixie Pyrite Beds, etc.). Beautifully summarized in the official Penn Dixie Field Guide by Stokes and Schrieber (2017) these beds contain abundant brachiopods (*Ambocoelia, Mediospirifer, Mucrospirifer*, etc.), bryozoans (A*tactotoechus, Fenestella*, etc.), corals (e.g., *Amplexiphyllum, Stereolasma, Favostires, Pleurodictyum*), crinoids, bivalves, gastropods, ammonites (e.g., *Spyroceras, Tornoceras*) and trilobites (e.g., *Eldredgeops, Greenops, Dipleura, Bellacartwrightia*, and *Pseudodechenella*). Large platy fish (placoderm) remains, such as *Eastmanosteus*, *Dinomylostoma*, smaller arthrodires and ptyctodonts, are found with many members of the Hamilton Group (e.g., Bryant, 1921; Buehler and Tesmer, 1963; Stokes and Schreiber, 2017).

Flora

Within the Genundewa, Middlesex, and Rhinestreet there are large fragments of plant remains. In the Middlesex and Rhinestreet shales these tend to be compresses carbon residues with either a frond or woody stem appearance. In the Genundewa these plant fossils retain some volume, though often the state of preservation otherwise is poor. These plant fossils were presumably washed in from terrestrial sources similar to the Gilboa forests deposit of similar age in eastern New York (Stein et al. 2012).

Microfossils

Conodonts

The North Evans Limestone contains a highly diverse conodont fauna of at least 41 species of conodonts (Bryant, 1921; Buehler and Tesmer, 1963). The mixed conodont zone faunas suggest that the unit might represent a multiply reworked deposit that accumulated over a long time period spanning the Givetian-Frasnian boundary (Maletz, 2008). Other units contain much less diverse assemblages of this group. Over (1997) produced one of the most detailed conodont biostratigraphic records of the area.

Tentaculites

The Genundewa Limestone contains abundant tentaculites and styliolina fossils; *Styliolina fisurella* being the most common fossil. Buehler and Tesmer (1963) provided a complete list of the rich conodont and fish fossils also found in relative abundance within this unit.

Foraminifera

Middle to Late Devonian (Givetian through Famennian) black and gray shale beds of Western New York contain hundreds of diminutive calcareous and agglutinated foraminifera (Wilson and Schieber, 2015; Li et al., 2021). The genera within these beds are reminiscent of shallow modern predominance facies. These foraminiferal assemblages and their associated predominance facies (Li et al., 2021) correlate well with prior lithologic and geochemical investigations that establish this portion of the Appalachian Basin as a deltaic setting but suggest are likely an inner neritic zone with depths most likely not exceeding 50m (Smith et al., 2019; Li et al., 2021). Dominant genera include several species of *Ammobaculites* and *Saccammina* which suggest that paleodepths did not exceed 50 m throughout the Frasnian. Opportunistic genera reflect a muted crisis associated with the *punctata* isotopic event (aka Rhinestreet Bioevent) and Lower Kellwasser (Pipe Creek) events. While there are definite shifts in the diversity of assemblages between gray and black shale, the foraminiferal type and feeding mode, indicative of depth and oxygen availability respectively, there is little variation between the distinct beds. No significance was found between total organic and foraminiferal type of feeding mode. Identification at the species level is problematic but suggest that the depositional environment was stressed. However, the effects of the end-Devonian mass extinction were not significant for these foraminifera in comparison to those frequently reported; the data suggests that there was no local extinction for benthic foraminifera at least through the lower-most Hanover Shale, just prior to the Upper Kellwasser and Hangenberg marine crisis events, within this portion of the deltaic complex of the Appalachian Basin of the western New York.

Algal Cysts – Tasmanites (?)

Over 50 years of palynological research of Late Paleozoic deposits have helped to determine that the lithologic sequences and paleoenvironments based on fossil dinoflagellates, acritarchs, foraminifera, silicoflagellates and radiolarians. The miospore biostratigraphy of the Late Devonian Appalachian Basin is well established (Streel, 1972; McGregor, 1979; Richardson and Ahmed, 1988; Loboziak and Melo, 2002). Within the Devonian Appalachian basin, varying palynological components have been reported (Over, 1997; Feist and Van Aller Hernick, 2014; Lash, 2016; Chamberlain et al., 2016; Kelley et al., 2019; Haddad et al., 2016; Boyer et al., 2021) but most commonly as secondary findings in these investigations.

In the past, *Tasmanites* have been commonly misinterpreted as terrestrial plant spores that were transported by the wind or currents into the black-shale seas and basins primarily due to their lack or limited external texture and internal structure. Although Dawson (1871) first described similar spore-like discs as Sporangites, Newton (1875) provided a more thorough description of the fossil and named it *Tasmanites* after the Tasmanian white coal, which is composed of these discs. Newton (1875) thought that *Tasmanites* were, "vegetable organs" or spore cases, but he was not certain from which plant they had come and suggested that they were related to lycopod macrospores. Subsequent investigations found that *Tasmanites* were of algal origin, bearing a resemblance to modern green alga (e.g., Schopf et al., 1944; Wall, 1962; Brooks, 1971). While debate remains as to the true marine algal assignment in this portion of the deltaic wedge, these structures have been identified as *Tasmanites* in the Tully Formation in Pennsylvania (Chamberlain et al., 2016) and several other coeval basins across Laurentia (e.g., Iowa Basin, Michigan Basin, Ohio Basin, etc.).

On a world-wide basis, marine algal cysts are found in deposits ranging in age from the Silurian to the Cretaceous (Brooks, 1971) and their remains occur in relative higher abundance throughout black shales ranging in age from Middle Devonian to Early Mississippian in both Gondwana and Laurentia (Revill et al., 1994; Vigran et al., 2008; Wicander and Playford, 2008; Haddad et al., 2016; Mouro et al., 2016; Lelono, 2019). Many such enrichments have been interpreted as reflecting algal blooms in areas supplied with meltwater from surrounding glaciers or other terrestrial run-off sources (López-Gamundí, 2010; Lelono, 2019). The localities investigated in this study are known to have been located within the southern tropics (Fig. 1a) and mostly likely did not have direct meltwater influence. There is strong sedimentological evidence that this deltaic wedge may have been inundated with fresh water run-off due to increased seasonal riverine input shifting the hydrochemistry and productivity levels (Arthur and Sageman, 2005; Lash, 2016; Kelly et al., 2019; Haddad et al., 2016; Boyer et al., 2021; Bartlett et al., 2021 (in press)) leading to a series of black shale formation.

Charophytes

Charophyta fossil materials include oogonia (female reproductive organs; most common), antheridia (male reproductive organs, less common), and portions of the Characeae plant itself (e.g. branches, thalli, stipules, leaves, etc.). Well-preserved oogonia and thalli have been reported in the Givetian-aged Ludlowville Formation in eastern New York (Feist and Van Aller Hernick, 2014) and more recently in Bartlett et al. (2021, in press) within several deposits in western New York. Outside of this single publication, charophyte fossils are yet to be reported from any other portion of the Upper Devonian Appalachian Basin.

Charophyta are a group of extant aquatic alga found most commonly in freshwater but have been found to be occasionally abundant in brackish areas (Shepherd et al., 1999; Soulié-Märsche, 1999, 2008). Generally occurring in quiet or gently flowing waters, charophytes have specific light and specific oxygen level requirements (de Winton et al., 2004; Küster et al., 2004) and so they are usually found in very shallow depths (several cm) but have been found in deeper waters, where nutrient input is reduced (<30m; Middelboe and Markager, 1997) as long as light sources are adequate. Some have been found in swiftly flowing rivers (McCourt et al, 2016), but such occurrences have been rarely noted in the literature.

A clear and distinct inverse relationship has been identified between the presence of marine algal cysts and charophyte fossils within Frasnian beds (Bartlett et al., 2021 (in press)). Two of beds have discrete contacts and mark the onset of a marine crisis and known transgressive events (Johnson et al., 1985; Veevers and Powell, 1987; Algeo et al., 1995; Algeo and Scheckler, 1998; Ver Straeten et al., 2011); the Cashaqua-Rhinestreet contact of the *punctata* event, the Angola-Pipe Creek contact for the onset of the Lower Kellwasser event. The Upper Kellwasser event occurs within the Hanover Formation.

Trace Fossils

Burrows are by far the most obvious trace fossils in this succession and are mostly clearly present at the base of the Genundewa. Burrows can sometimes be several centimeters across and cross-cut the underlying North Evans Limestone in some locations. The surface of the Tichenor limestone may also show burrows of substantial size and smaller, mostly horizontal, burrows can be found in the grey shale units such the Windom. These small burrows weather to a rusty brown color that is easy to spot against the gray background shale.

DEVONIAN MARINE CRISES EVENTS

The Devonian Period contains evidence of a series of marine crises that are recorded globally, a few which, in western New York in particular, correlate to the deposition of black and gray shale sequences or limestones, and accentuated pelagic faunal turnovers (e.g., House, 1996; DeSantis et al. 2007; Haddad et al., 2016; Kelly et al., 2019, Lash, 2019). The events which straddle the beds of this investigation are the punctata event (Rhinestreet event), the Lower Kellwasser Event (Frasnian-Famennian, with the Lower Kellwasser Event at the Pipe Creek, ~372 Ma and second and more severe Upper Kellwasser Event in the uppermost Hanover Shale). The latter event impacted up to 80% of marine species (Hallam and Wignall, 1997). The subsequent and terminal Devonian Hangenberg event (359 Ma; not covered in this field trip), saw losses of an additional ~50% of invertebrate diversity (upper Hanover Shale, correlated beds not included in this investigation; Streel et al., 2000).

Taghanic Event

A major global eustatic transgression is recognized in the Late Devonian. Pre-Taghanic Event Givetian marine areas of eastern and western North America were not connected across the southern United States and were connected only briefly between western Canada and the north-central United States (Indiana and Michigan). upper Tully Formation, a major transgression accompanied upward change from clean carbonate to increasingly muddy limestones, and culminated in the overspread of anoxia and resulting black mud deposition within the foreland basin; this deepening partly reflects eustatic highstand conditions (Johnson et al. 1985), but it also was greatly enhanced by flexural loading of the craton by thrust slices during a collisional pulse (Third Tectophase) of the ongoing Acadian Orogeny (Ettensohn et al., 1998). Pre-Taghanic Middle Devonian brachiopod faunas of eastern and western North America belong to different faunal provinces, that is, to the Appalachian and Old World provinces respectively. The Taghanic onlap of the continental backbone in the southwestern United States provided shallow-water marine areas for dispersal of bethonic animals and the resulting intermigration brought an end to brachiopod provinciality that had prevailed since the Early Devonian. By analogy, provincial shifts in established faunal successions should provide dates for other sedimentary-tectonic events.

Because Appalachian Province fossils are known to range as far as Colombia and Venezuela in Emsian-Eifelian time, but not at the same time to the American west, a large land barrier is postulated for times of provinciality, that is, during the intervals Ludlow-early Siegenian, and Emsian-mid-Givetian. Other intervals during the Silurian and Devonian were times of breaching of the land barrier by marine seas.

During deposition of the Additionally, Tully strata record major faunal fluctuations associated with the regional demise and/or geographic restriction of the long lasting, diverse, and endemic Hamilton Fauna (Brett 1995; Sessa et al., 2002; Baird and Brett, 2003; Baird et al., 2003); this pattern of faunal overturn and global extinctions, known as the "Taghanic Event" (Johnson, 1970; House, 1981), is increasingly recognized as actually representing two or more temporally closely-spaced global bioevents which may have been as severe, or more so, than the widely-known Frasnian-Famennian extinction (Aboussalam and Becker, 2001; House, 2002).

punctata Event

At the contact of the Middlesex-Cashaqua a muted expression of the well documented global *punctata* carbon excursion event in North America begins and terminates just below the Cashaqua-Rhinestreet contact (Lash 2019). Magnetic susceptibility readings published by Lash (2019) shows that the final peak of the ^{13}C excursion in this locality begins at 40 cm below the contact Cashaqua-Rhinstreet, following this final peak at 10cm below the contact. Recent foraminiferal investigations has shown that the number of species sampled drops from 22, at 10 cm below, to 11 (50%), at the contact (Li et al., 2021). The lowest number of genera within the Rhinestreet Formation, a notable second plummet, occurs at 60 cm above the contact (Li et al., 2021). The reduction of genera occurs where the lithology shifts to a well lithified black shale. While several species 'disappear' within these formations, their genera do persists throughout the Frasnian deposits of western New York. Marine algal cysts (possibly *Tasmanites*) have recently been identified and found to be highly concentrated at 40 and 21 cm below the Cashaqua-Rhinestreet contact (Bartlett et al., 2021 (in press)). These concentrations also coincide with carbonate concretion layers, well preserved freshwater algal reproductive organs (e.g., oogonia, antheridia) and associated branchlets and stipules, and microtektites (Lash, 2019; Bartlett et al., 2021 (in press); Meehan and Lash, 2021).

Lower Kellwasser Event

The Pipe Creek Shale marks the onset of the Lower Kellwasser Event and contains the highest peak of foraminiferal species counts for Frasnian beds in western New York (Li et al, 2021). Being flanked by gray shales, the Angola and Hanover Shale Members, with periodic increases in epifaunal and mixed populations suggests that within the time just prior to, during, and immediately after the Lower Kellwasser Event, this portion of the deltaic wedge was either well oxygenated or the organic influx remained high. There is essentially no biotic turnover within foraminiferal populations; genera/species remain the same, abundance, diversity, and richness are statistically unchanged within the Pipe Creek Shale.

Up-section, the Hanover Shale contains a continuously diminishing number of foraminiferal genera (Li et al., 2021). While the authors suggested that this decrease in population may be reflective of the biotic crisis, the sampling did not extend high enough within the member to determine whether this is the case absolutely. Within the reaches of the investigation conducted by Li et al. (2021) the expression in benthic foraminiferal may not be as distinct as in the macrofaunal due to the opportunistic nature of the populations in these deltaic environments and the lack of taxonomic literature with which to identify species, thus, potentially muting any changes in diversity at the species level.

Upper Kellwasser Event

The Upper Kellwasser event is represented in western New York by a thin black shale deposit within the Hanover Formation, less than a meter below the contact with the overlying Dunkirk Formation (Lash, 2016). While the Kellwasser events have traditionally been associated with consistent widespread anoxia, and even euxinic conditions this does not appear to be the case in western New York based on geochemical (Lash, 2016; Kelly et al.2019; Haddad et al., 2016) and trace fossil (Boyer et al., 2021) evidence. Instead, findings suggests rapidly fluctuating oxygenation between dysoxic to anoxic and perhaps occasionally euxinic conditions The Upper Kellwasser bed is associated with a negative $\delta^{13}C$ excursion and high TOC followed immediately by a positive excursion and drop in TOC for the remainder of the Hanover Formation (Lash, 2016). It is worth noting though that the overlying grey shales of the Dunkirk Formation continue to have elevated TOC levels while carbon isotopes return to background levels prior to the Upper Kellwasser event. This recovery interval corresponds to an increase in siliceous microfossils (i.e. radiolarians). Several additional papers have recently been published adding much data to support these hypotheses. Recent data include marine agal cyst abundance, lipid geochemical proxies, trace metals, and ichnofossil assemblages; suggesting that this environment was highly stressed and not likely to be continuously anoxic, but periodically euxinic (Kelly et al., 2019; Haddad et al., 2016; Boyer et al., 2021).

FIELD GUIDE AND ROAD LOG

Meeting Point: Buffalo State Campus; Burchfield Penny Art Center Parking lot Rockwell Road and off Elmwood Avenue Entrance (across from the Albright-Knox Art Museum).

Meeting Point Coordinates: 42°52'53.11"N 78°52'42.10"W

Distance (miles) Cumu-lative	Point to point	Route Description

Stop A: Eighteenmile Creek Access Site at the end of Basswood Drive, Lake View, New York. 42°42'27.0"N 78°57'22.1"W

20.1	0.3	From Rockwell Road Entrance at Elmwood Avenue, turn left, head north on Elmwood Avenue toward Iroquois Drive.
	0.1	Turn left onto NY-198 (Scajaquada Pkwy).
	1.2	Merge onto NY-198 W.
	3.6	Use the left lane to merge onto I-190 S toward Downtown Buffalo.
	7.6	Take exit 7 to merge onto NY-5 W/Buffalo Skway toward Outer Harbor/ Lackawanna. (Bad merge – take caution of right lane oncoming traffic from Downtown).
	7.1	At fork (after Woodlawn State Park Overpass/Ford Stamping Plant) keep right to stay on NY-5 W.
	0.2	Continue to North Creek Road on NY-5 W (BEFORE the bridge; turn around available off South Creek Road).
	0.1	Turn Left onto North Creek Drive.
	0.1	Turn Right onto Basswood Drive. Destination at end of the street
	0.2	Walk down dirt path to Eighteenmile Creek.
	0.3	Walk downstream. Outcrop on north side of the creek.

Regardless of creek level reaching this outcrop will require at least a brief dip into shallow water (usually less than a foot deep), so dress appropriately. This large amphitheater-style exposure stretches ~0.3 miles along the north side of the creek with the Tichenor limestone member of the Moscow Formation at the base providing a resistant platform to walk along. At this site members of the Hamilton (Tichenor, Windom), Genessee (North Evans, Genundewa, West River), and Soneya (Middlesex, Cashaqua) Groups can be observed. Large blocks of the more resistant North Evans, Genundewa, and Middlesex formations are common along the surface and can be readily examined. The underside of the North Evans is most accessible in situ at the upstream end of this outcrop where its unusual character is clearly visible.

There is a second exposure just upstream of where we will enter the creek but it is more difficult to view without crossing deeper water. However, for those interested this secondary location shows the full exposure of the

Genundewa, West River, and Cashaqua units and even shows the lowest few meters of the Rhinestreet Formation at its top. The ubiquity of carbonate nodule bands in the Cashaqua is on full display at this exposure as well.

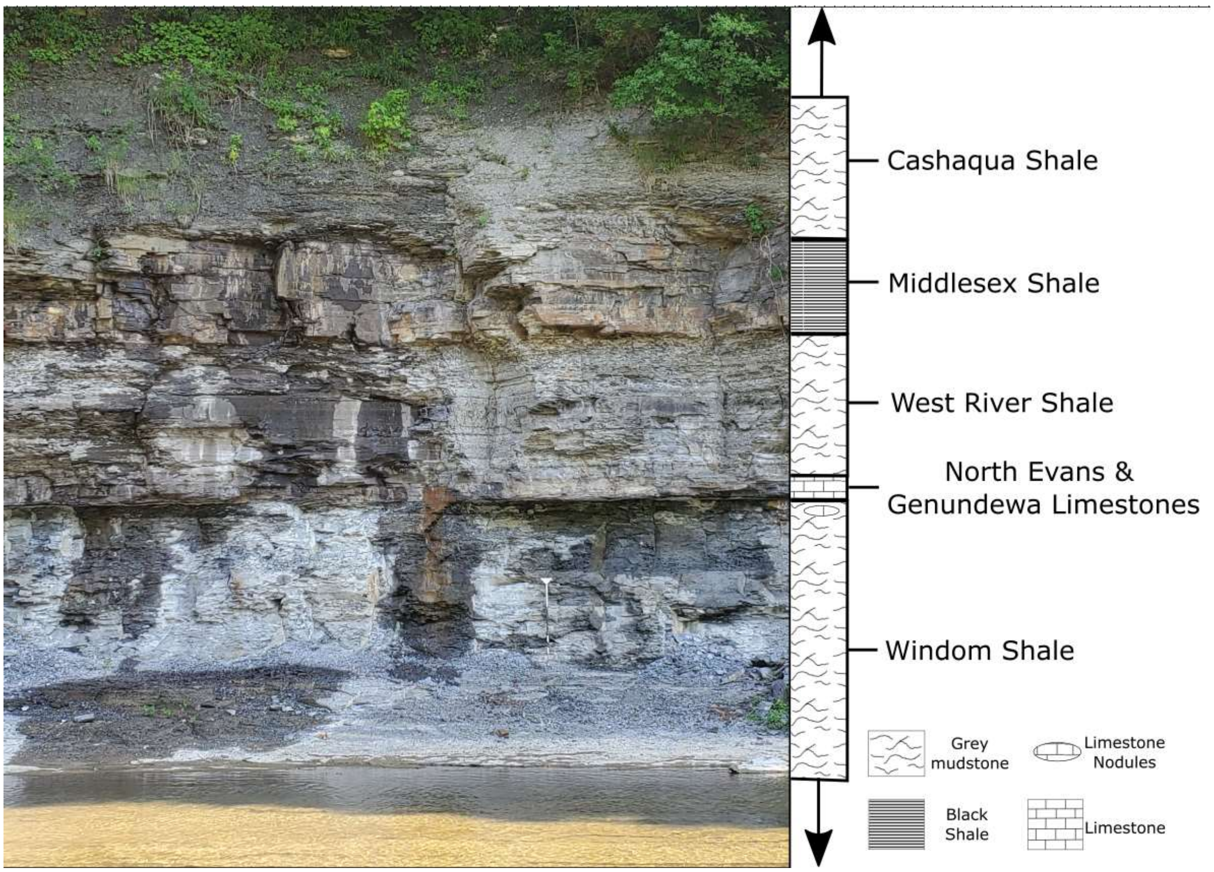

Figure 3. Photograph of stop A along Eighteenmile Creek with lithological units labeled. Arrows on the stratigraphic column indicate units that are truncated at this location.

Stop B: Eighteenmile Creek Hobuck Flats at the end of Versailles Road, Derby, New York. 42°42’28.07” N 78°57’22.30”W

2.6	0.2	Head north on Basswood Drive toward North Creek Road.
	0.1	Turn left onto North Creek Road.
	0.5	Turn left onto NY-5 W.
	1.6	Cross the Eighteenmile Creek Bridge and make left at South Creek Road.
	0.2	Turn left onto Versailles Road. Destination at the end of the road.
	0.2	Walk downstream. Outcrop along north side of creek.

A short walk upstream there is an exceptional exposure of most of the Cashaqua and Rhinestreet Formations. There are numerous carbonate concretion layers accessible on foot from the Cashaqua and nodules originating from the Rhinestreet can often be found as talus along the base of the outcrop. Some of these nodules contain ammonite fossils and more rarely other macrofossils, such as placoderm fishes. There are also several soft pockets of organic matter within the upper Cashaqua that can often be spotted by their rusty weathering pattern. While the Rhinestreet is dominantly a black laminated shale there are also silty layers of grey shale within it that are visible at this locality.

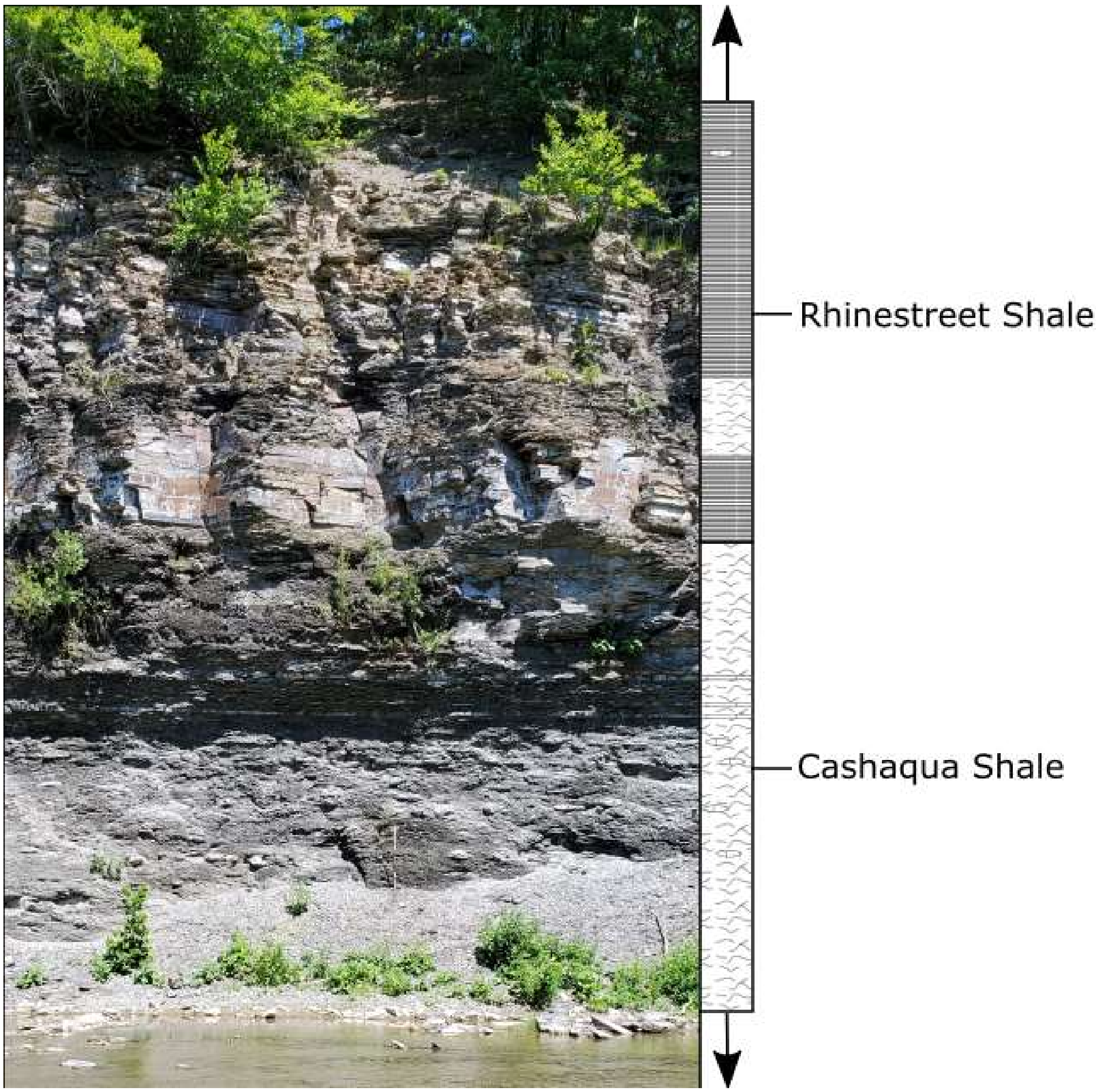

Figure 4. Photograph of stop B along Eighteenmile Creek with lithological units labeled. Lithological key as in figure 3.

Stop C: Eighteenmile Creek at NY-20 (Southwestern Blvd), fisherman's trail across from North Evans Cemetery. Gravel path descent, may be washed out in places. Take caution on this foot path). 42°41'42.20"N 78°56'10.34"W

0.5	0.2	Head southeast on Versailles Road toward South Creek Road.
	0.2	Turn left onto South Creek Road (becomes Shadagee Road)

0.2 Continue on Shadagee Road (also So. Creek Road). Park on right at North Evans, St. Vincent De Paul Cemetery.

Uppermost Cashaqua Formation and beginning of the Rhinestreet Formation (Frasnian). This is the location where Lash found the ^{13}C/TOC excursion for the *punctata* event and also at 20 and 40 cm below Cashaqua-Rhinestreet contact, at concretion layers, microtektites presumed to be from the Alamo event were found in macerated shale. Flooding event; hyperpycnal/monsoonal suspected.

Some ammonoids and their casts have been found here, however, this location is mostly barren of macrofossils. Though there are abundant microfossils – not visible with naked eye.

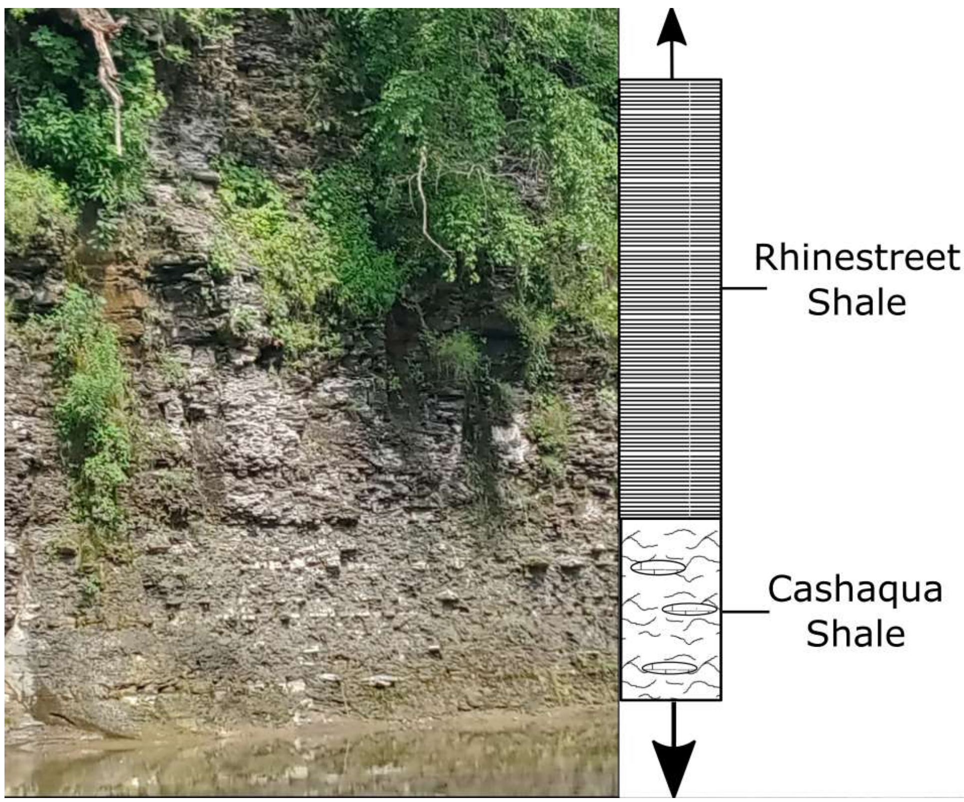

Figure 5. Photograph of stop C along Eighteenmile Creek with lithological units labeled. Lithological key as in figure 3.

Stop D: Preischel's Farm. 2993 Belknap Road, Eden, New York. 42°41'1.06"N 78°52'47.30"W

3.3	0.9	Head southeast on Shadagee Road toward Bauer Road.
	0.9	Turn left on Bauer Road.
	1.5	Continue Straight on Belknap Road. Meet in lot on left at Preischel's Farm.

Optional drive/walk

*** All wheel or 4x4 drive <u>strongly</u> suggested. By foot this is a 0.5 mile walk:

Take dirt road next to tractor garage and milk silos on eastside of street. Last I was there the road was pretty washed out and steep, strongly recommend parking before the final drop off (you WILL see it), field with household appliances there for parking.

Rhinestreet and Angola Formation transitional zone, scraggy layer. Amazingly large concretions (average 1 m but can be as wide as >3m). This location contains geochemical signals and microfossils (charophytes, oogonia, antheridia), suggest seasonal to intermittent but regular flooding. Some macrofossils can be found but are rare. The Rhinestreet Formation is rich in methane/oil and contains a very distinct and strong odor.

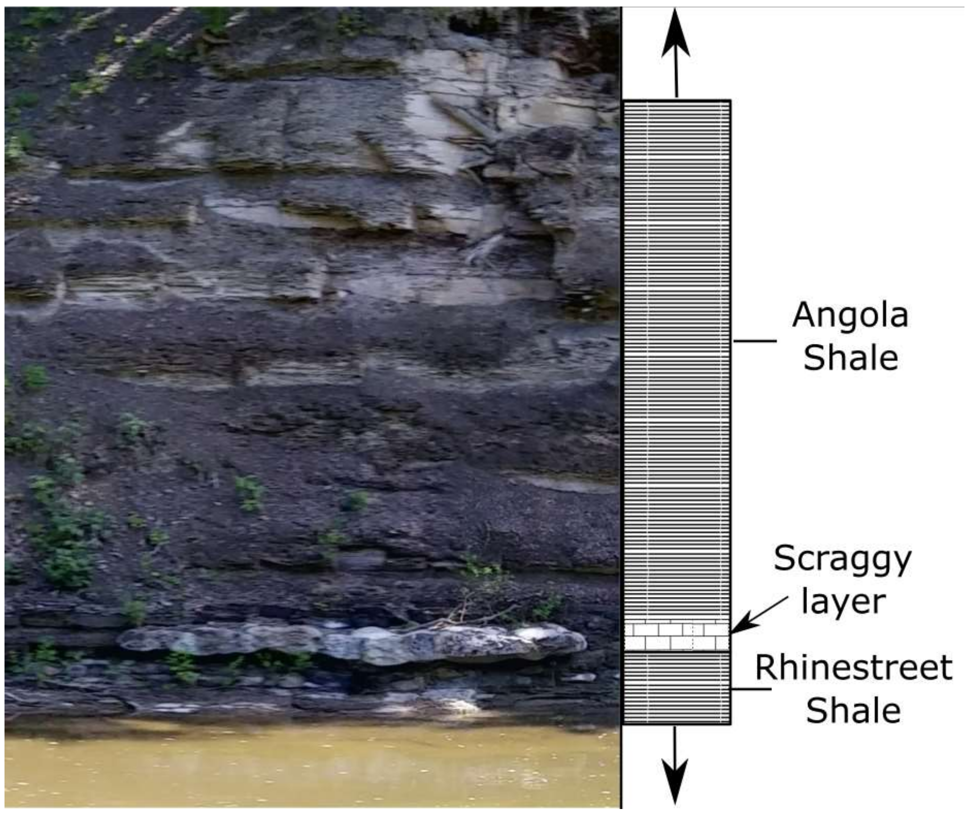

Figure 6. Photograph of stop D at Preischel Farm with lithological units labeled. Lithological key as in figure 3.

Stop E: US- 219 and Zimmerman Road overpass (broad shoulder for parking). 42° 40' 40.35"N 78° 47' 01.63"W

7.3	0.8	Head south on Belknap Road toward Bley Road.
	0.4	Turn left on Bley Road.
	443 ft	Continue onto Eden Valley Road to the right (NY-62)
	4.3	Make immediate left onto North Boston Road.
	0.3	Turn left onto Heinrich Road
	0.4	Continue to the right onto Eckhardt Road
	0.1	Turn right onto NY-391 S/Boston State Rd/Hamburg-Springville Rd.
	1.2	Use the right lane to merge onto US-219 S via the ramp to Springville.

Park on wide shoulder near Zimmerman overpass.

*** be mindful of oncoming traffic. Speed limit is 65 mph, use due diligence should you cross the highway. It is strongly recommended that you use proper exit/entrance ramps to view the east side of the outcrops.

Angola, Pipe Creek (1m thick), and Hanover Formations on the west, middle, and east side shoulders. Oldest to youngest trending north to south along the highway. Pipe Creek marks the Lower Kellwasser Event. Uppermost Hanover marks the Upper Kellwasser Event.

Pipe Creek finger concretions 13C bear evidence of short-lived flow – disputed as to whether microbial, hydrates, or mud volcano … nonetheless, methane was discharged either due to common weather related mechanical dissociation.

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EIGHTEEN MILE CREEK: SOUTH BRANCH TO LAKESHORE; MIDDLE TO UPPER DEVONIAN SHALES

PAUL A. AGLE
Department of Atmospheric and Geological Sciences, SUNY Oswego

INTRODUCTION

For logistical purposes, we will run this trip "backwards" in geologic time. We will start at the headwaters of Eighteen Mile Creek South Branch on the Appalachian Plateau. We will then begin to descend the plateau as the creek incises into the Upper Devonian shales. In these exposures, we will be able to observe important markers within the Late Devonian, including both the Lower and Upper Kellwasser extinction horizons. As we journey further back into geologic time, we will observe shales deposited during times of less environmental stress and, finally, we will end at the Lake Erie shoreline which provides excellent fossil collection opportunities. This morning trip serves as a companion trip to the afternoon "Penn Dixie Fossil Park & Nature Reserve: A Window into the Devonian Period of western New York" and the final portion of the road log will take us to the quarry where that trip will begin.

Eighteen Mile Creek is named not for its length, but for its distance along the Lake Erie shoreline from the mouth of the Niagara River. The mouth of the Buffalo River, which would eventually become the terminus of the Erie Canal is about 2miles closer. These major navigable waterways were vital to the indigenous peoples, later colonial communities, and the industrial history of the area. On this trip, we will be visiting the locations of two of the earliest mills along the creek (Stops 2 & 3). In addition to powering mills, Eighteen Mile Creek and its valleys and floodplains provide fertile soils which continue to support a vibrant agricultural industry, particularly sweet corn, which is celebrated annually during Eden's Corn Festival (1st weekend in August).

The lower gorge and lakeshore sections have been a popular site for previous NYSGA field trips. These trips mostly sought to analyze a specific stratigraphic section (e.g., Zambito and Mitchell, 2006 – Wanakah Member; Lash and Blood, 2006 – Rhinestreet Member) or to address specific geological / paleontological problems (e.g., Over et al., 1999 – Disconformities; Baird et al., 2006 – Geologic and Biotic Event horizons). This trip is meant to serve as a comprehensive overview of the creek itself, primarily the South Branch, but it also provides some optional stops along the main (northern) branch. The Eighteen Mile Creek gorge can be intimidating and may seem unapproachable at first. This trip will highlight areas that provide relatively easy accessibility for other researchers and students who may wish to study these intervals in more detail. For this purpose, the trip also provides a broad sampling of the stratigraphy (ranging from Middle Devonian (Givetian) Wanakah Member up through the Late Devonian (Famennian) Dunkirk Member. From a paleontological perspective, we will see fossil communities thriving along the lakeshore section (and later at Penn Dixie Fossil Park in a separate afternoon trip) and higher in the section, we will explore outcrops surrounding the Kellwasser Events, where these fauna were stressed or eliminated.

Geological Setting

On this trip, we will be examining rocks (mostly shale) from the Middle to Late Devonian. These sediments were deposited into the Appalachian foreland basin, which is associated with compression related to the Acadian Orogeny (Figure1). At the time of deposition, the study area was positioned approximately 20° south of the Equator (Scotese and McKerrow, 1990). The sediments were primarily derived from the erosion of the Acadian Mountains with sediments generally fining to the west in a prism known as the Catskill Clastic Wedge (or Catskill Delta). Figure 2 illustrates the recurrent unconformities associated with the cratonward migration of the peripheral bulge and the tongues of black shale that are associated with the subsequent flexural deepening of the basin during more active pulses of tectonic activity and crustal loading in the Acadian Orogen (Ettensohn, 1994). This flexural deepening of the basin and the westward distal position within the basin results in the stratigraphy of the study area (Figure2) being dominated by shales and silty shales with only the occasional pulse of coarser material (usually pinches out east of the study area) and carbonates during times of lower relative sea-level and/or higher biologic productivity.

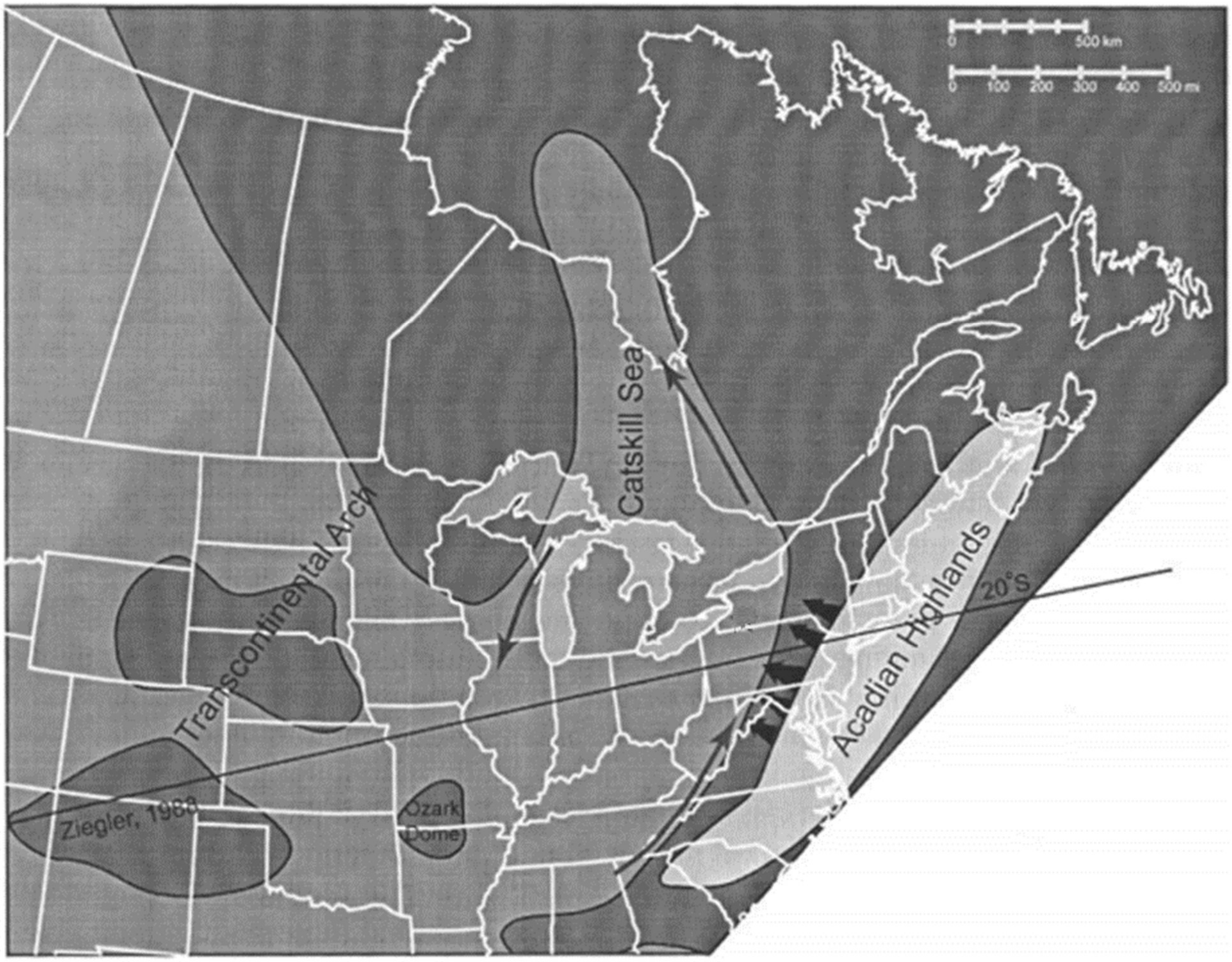

Figure1. Devonian Paleogeography showing Catskill Sea (Appalachian Basin) and Acadian Highlands. 20°S latitude parallel based on paleomagnetics of Ziegler (1988). From Smith and Jacobi (2006).

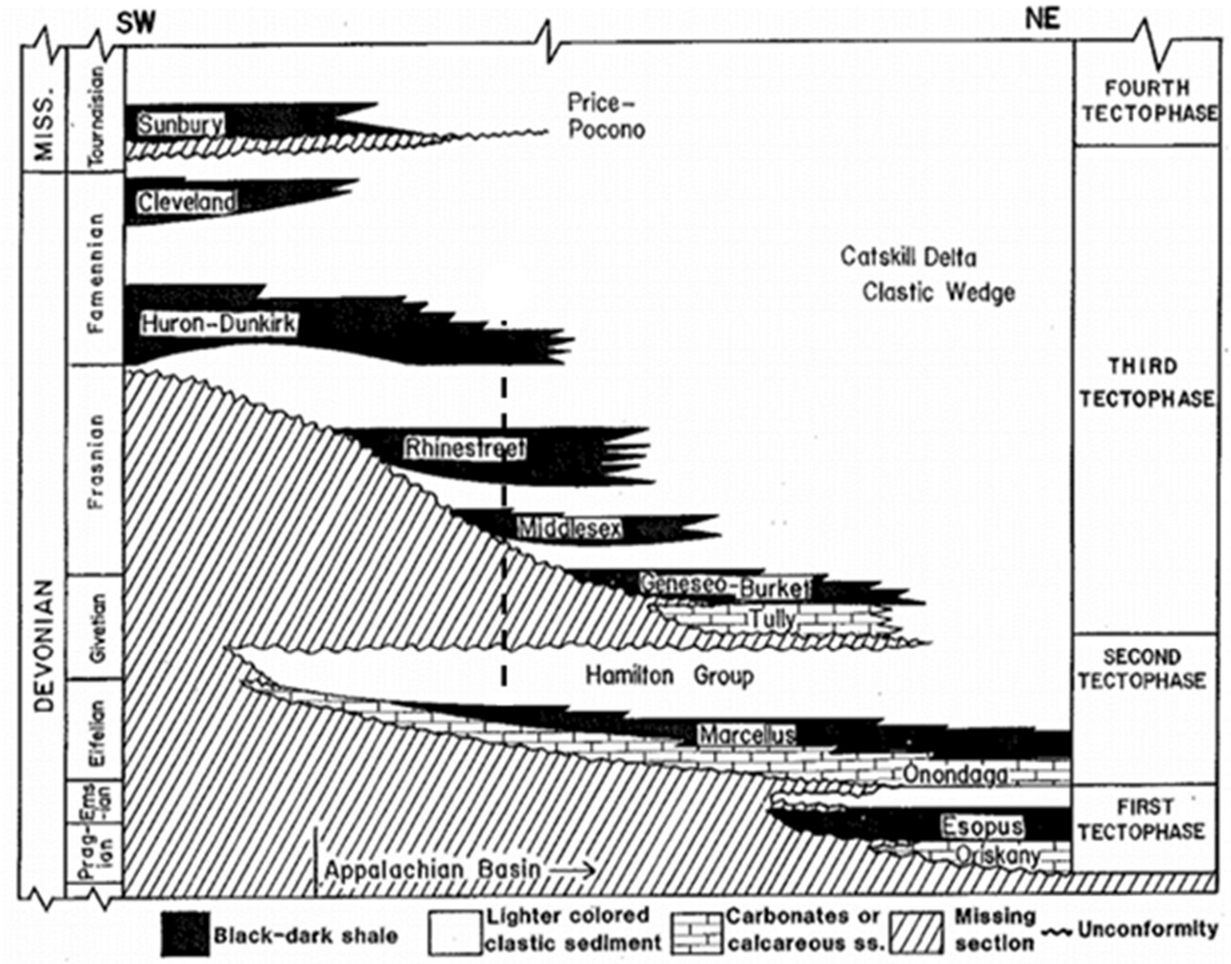

Figure2. Composite stratigraphic section from north-central Ohio (SW) to east-central New York (NE) through the northern portion of the Appalachian Basin. The dashed black line indicates a very approximate location and interval covered in this trip. From Ettensohn (1994).

The Late Devonian is also a time of major faunal disruption. Raup and Sepkoski (1982) assembled family-level fossil data from both vertebrates and invertebrates and identified 4 Phanerozoic intervals when extinction exceeded the background rates, deemed mass extinctions (Fig3A). A 5th event (Devonian) was also above background, but not statistically significant, as this extinction is "smeared" across three different stages: Givetian, Frasnian, and Famennian (Raup and Sepkoski, 1982). The Devonian mass extinction is more accurately a series of major extinction events. Despite being less abrupt than the end Ordovician mass extinction, Droser and others (2000) assert that the Devonian extinction(s) were more ecologically impactful. Sepkoski and Miller (1985) indicate that from the Ordovician onwards the Paleozoic fauna were gradually being displaced from the shallow shelf environment and into deeper water by the Modern fauna (solid line, Fig3B), but the Devonian extinctions reverse this trend until well into the Carboniferous. However, these extinctions do essentially mark the end of the trilobite-dominated Cambrian fauna (dashed line, Fig3B). Although trilobites linger to the end Permian mass extinction, they never return to the abundance and diversity observed in the Devonian (as at Stop5 and in the afternoon Penn Dixie Fossil Park trip).

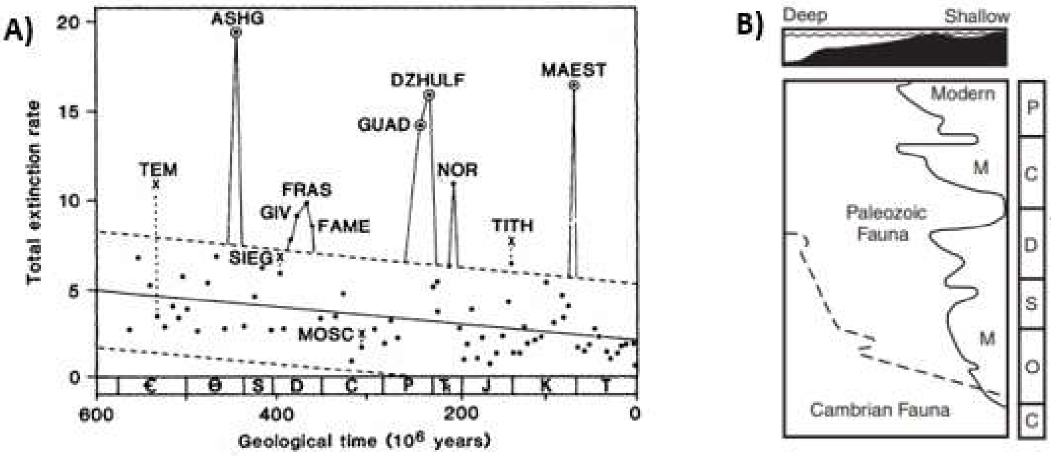

Figure3. Impact of Phanerozoic Extinction Events. **A)** Family-level extinction rates by geologic Stage from Raup and Sepkoski (1982). GIV – Givetian, FRAS – Frasnian, FAME – Famennian. Dashed lines represent range of "background" extinction. Spikes above this background rate are mass extinctions. **B)** Ecological reordering through time from Sepkoski and Miller (1985). Dashed line separates Cambrian Fauna and Paleozoic Fauna, Solid line separates Paleozoic Fauna from Modern Fauna

In addition to their geologic and paleoecologic importance, these Devonian shales are also significant sources of hydrocarbons in New York State and elsewhere within the Appalachian Basin. In fact, the first gas well in North America was drilled near Fredonia in 1821 (Hill et al., 2004) in Devonian shale. This shallow well likely produced out of the Dunkirk Member, which we will see at Stops 1 and 2. Chestnut Ridge Park and the Eternal Flames Falls are atop the next topographic ridge to the northeast from the Eighteen Mile Creek valley. Etiope and others (2013) geochemically finger-printed the source of this naturally occurring gas seep as the Rhinestreet shale, which we will see at Stop4. These hydrocarbons seep up along dominant fracture sets (Jacobi and Fountain, 2000; Schimmelmann et al, 2018) which also guide the topography of the area. Farther to the south and higher in the stratigraphic section the first oil "well" in North America was the Seneca (Cuba) Oil Spring in Cuba, NY as described in Van Tyne (2006) and the first actual oil well (the Drake Well) near Bradford, PA (Owen, 1975).

Previous Investigations

Amos Eaton (1830) published his geologic map of New York State just 15 years after William Smith's famous geologic map of England ("The Map that Changed the World"). According to its website, the New York State Geological Survey was established in 1836 and James Hall (one of Eaton's students) was named the first state Paleontologist and director of The New York State Museum. Since these early days, the rocks of New York State have been an outdoor laboratory to develop and test emerging geological and paleontological principles.

More specific to Eighteen Mile Creek, Amadeus Grabau (1898) published an extensive review of the geology and paleontology of the main branch of Eighteen Mile Creek starting at the Lake Erie shoreline and moving upstream to the present-day railroad bridge between Rts 5 and 20. Brett (1974) contains a basemap with Grabau's original numbered sections located on a topographic map. Over and others (1999) provide a detailed stratigraphic examination of this area (Figure4). We will be examining the Wanakah and Tichenor Members at Stop5 of this trip (and during the afternoon trip to Penn Dixie Fossil Park & Nature Reserve).

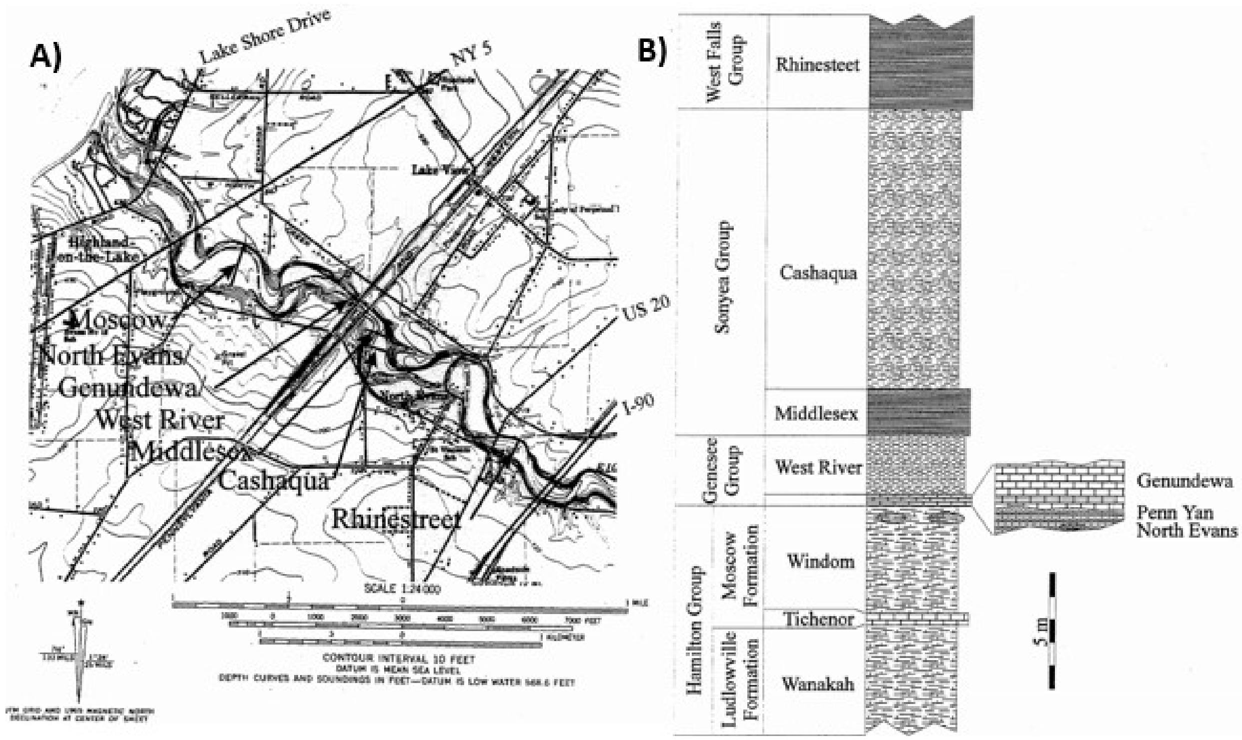

Figure4. Stratigraphic units of the lower reaches of Eighteen Mile Creek (main branch). **A)** Stratigraphic contact location (at creek level). From Over and others (1999). **B)** Measured section of this same interval. From Over and others (1999).

Grabau (1898) states that "the upper gorge and branches have not been examined", but the gorge and exposures continue well upstream from the Thruway bridge and extend up both the main branch and the south branch. Lash and Blood (2006) provide a focused stratigraphic (Figure5) and petrographic analysis of the Rhinestreet black shales near the confluence of these branches ("The Forks", Stop4 of this trip). Their guide focuses on the hydrocarbon potential for these organic-rich black shales. The older Marcellus shale and other hydrocarbon-bearing shales around the United States (e.g., Eagle Ford, Permian Basin, Barnett, etc.) currently garner most of the attention, but Hill and others (2004) provide a comprehensive survey of other potential targets in New York State. One of the more notable features within the Rhinestreet are the often very large 1m+ diameter internally laminated septarian carbonate concretions, which Grabau (1898) referred to as "turtle stones", "turtle backs", or "petrified turtles". These concretions form in a few discrete layers within the unit, which Lash and Blood (2006) cite as one piece of evidence that they likely formed near the seafloor shortly after deposition. Many concretions are visible *in situ* in the banks of the gorge and some excellent examples have weathered out and can be readily examined at creek level. The rigid concretions create stress shadows at their margins that preserve the initial "house of cards" depositional fabric of the clay grains, which is usually quickly lost throughout the complicated burial and compaction history of these shales (Lash and Blood, 2006).

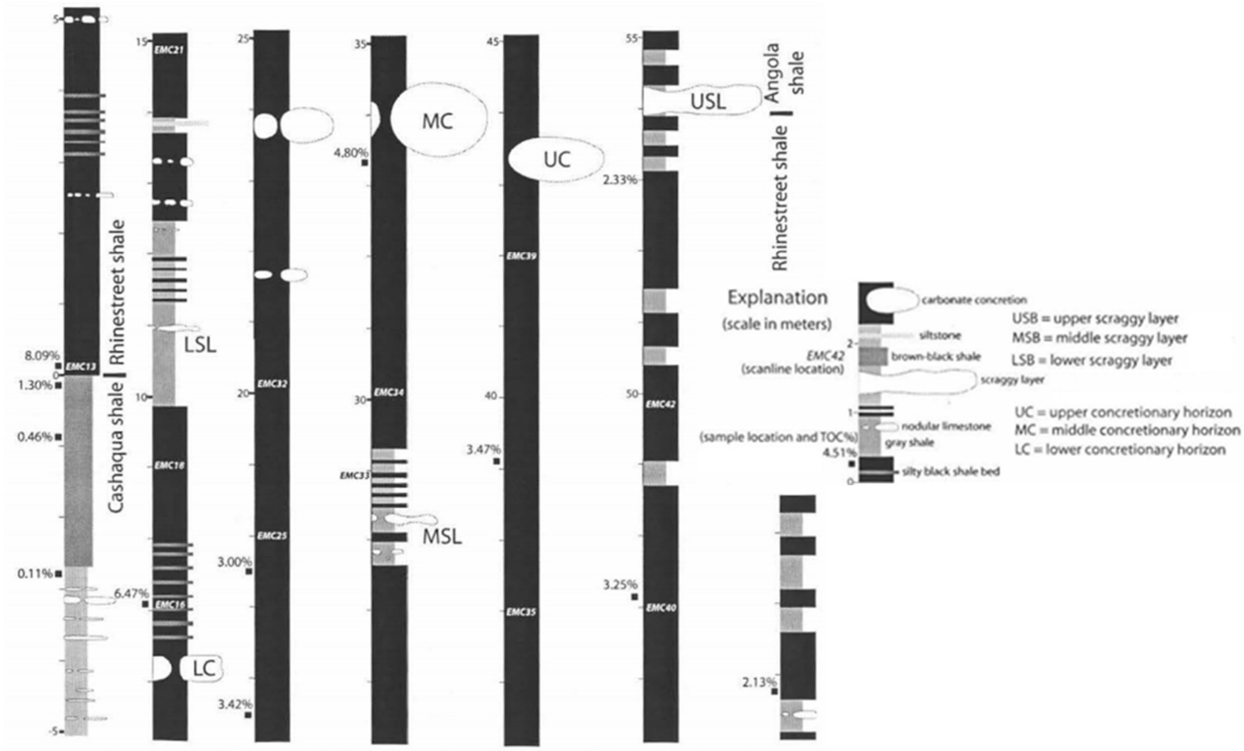

Figure5. Stratigraphic column of the Rhinestreet interval at "The Forks" area of Eighteen Mile Creek. Note that the column wraps around with the lowest portion of the section at the lower left and the highest portion at the far right. From Lash and Blood (2006).

Upstream from "The Forks", the creek divides into the main (northern) branch which flows through the Village of Hamburg and the South Branch which flows through the Eden Valley. For purposes of this trip, we will be focusing on the South Branch as the exposures are more consistent, however, we have also provided a list of relevant places along the main branch at the end of Stop4. Heading upstream on the South Branch, the gorge becomes less steep and widens near the historic Eden Valley Mill. The outcrop in this area is somewhat limited until the creek approaches Rt75 (Sisson Highway).

At Kromer's Falls near Old Mill Run Rd, we will examine the Lower Kellwasser Bed, which is the first of three major ecological disruptions in the Late Devonian: Lower Kellwasser Event, Upper Kellwasser Event, and the Hangenberg Event which occurs at the Devonian-Carboniferous boundary. The Kellwasser Beds were first defined by Schindler (1990) from the Late Devonian sections of Europe and Morocco. These events were initially recognized in western New York by Over's (1997) analysis of conodont biostratigraphy. Bush and others (2015) have recently revised the correlations of the basinal black shales (such as along Eighteen Mile Creek) with the more proximal coarser-grained deposits to the SE and their later field trip (Bush and others, 2017) extended these correlations into Pennsylvania. It can be difficult to recognize these ecological crises in the black shales since even outside of these crises the fauna of the deep basin are low abundance/diversity. Boyer and others (2014) analyzed burrow size and density (particularly of *Chondrites*) and trace metal content (molybdenum) as a proxy for bottom-water oxygen levels and the resulting environmental stress across the Upper Kellwasser interval. Both the Lower and Upper Kellwasser Events were stops on Baird and others (2006) field trip and their detailed stratigraphic sections are provided below (Figure6). Their Stop6 corresponds to our Stop3 and their Stop7 corresponds to our Stop1. We will also visit the post-Upper Kellwasser Dunkirk Member (Stop2).

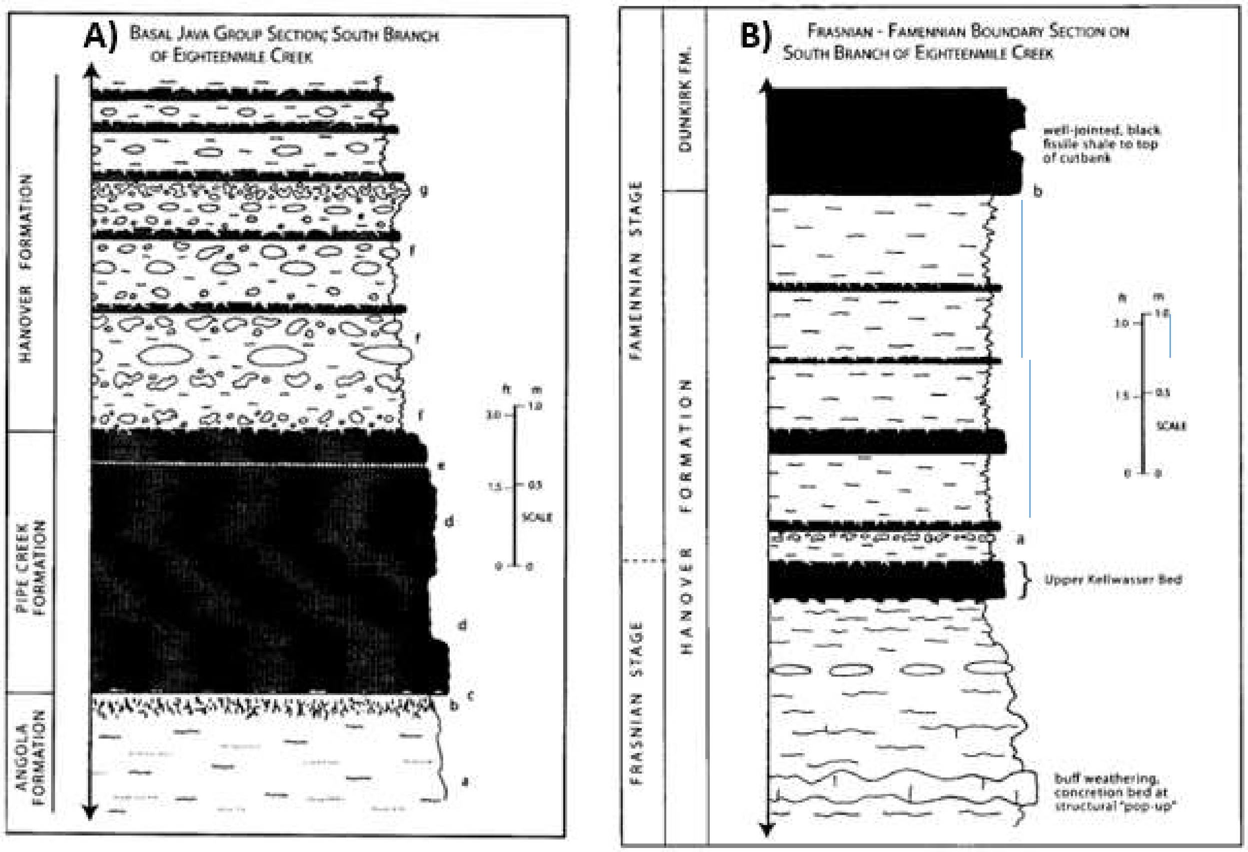

Figure6. Stratigraphic sections through the Kellwasser Events along Eighteen Mile Creek South Branch. **A)** Lower Kellwasser Event as seen at Kromer's Falls (Stop3) **B)** Upper Kellwasser Event as seen near New Oregon Rd Bridge (Stop1). From Baird and others (2006).

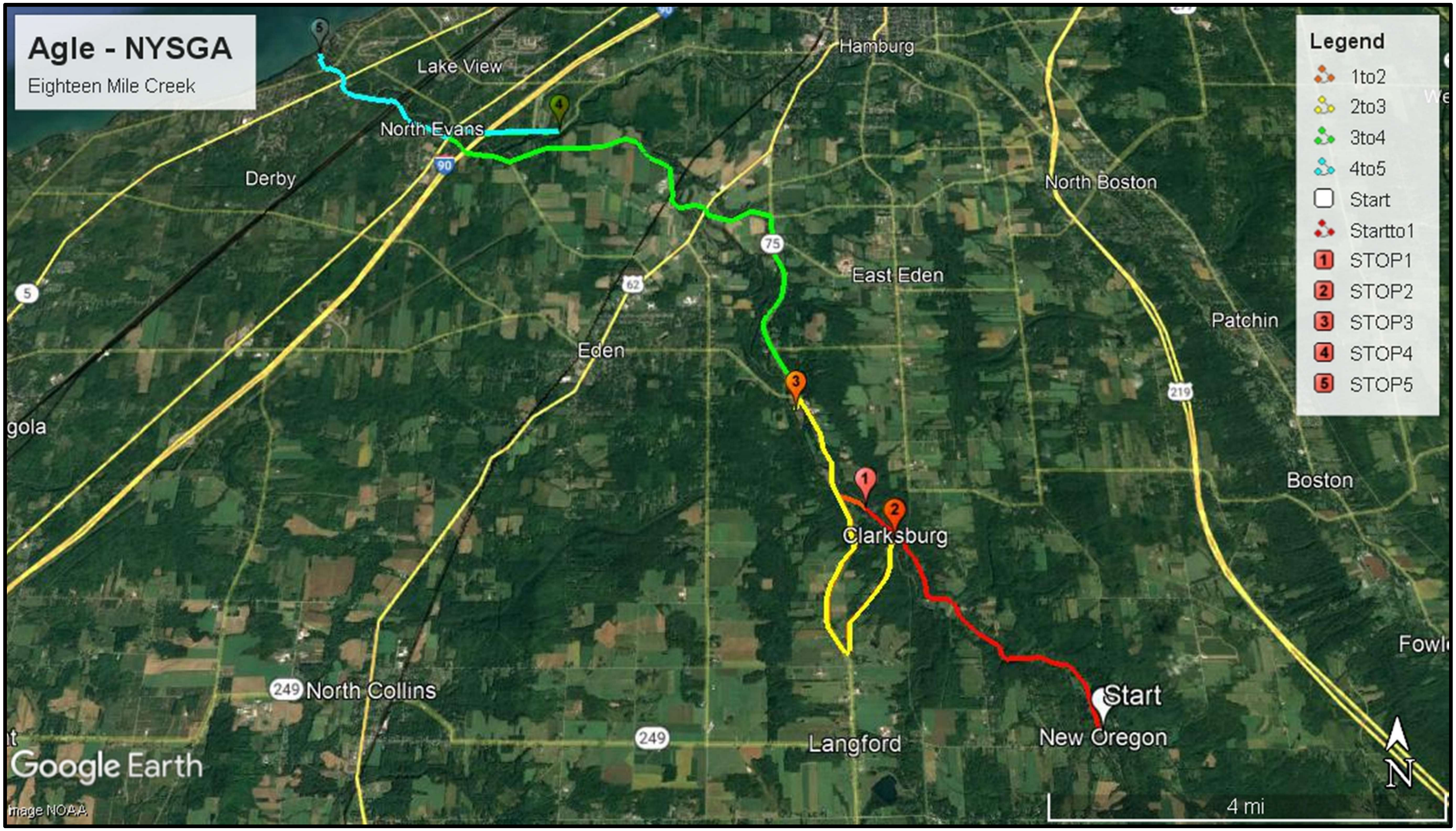

Figure 7. Location Map and Trip Route. Start – Frances Danter Memorial Park (corner of Langford and New Oregon Rds in New Oregon, NY); STOP1: New Oregon Road Bridge; STOP2: Clarksburg Country Club; STOP3: Kromer's Falls; STOP4: Eighteen Mile Creek Park; STOP5: The Bluffs; Please join us at Penn Dixie Fossil Park & Nature Reserve (4050 North St, Blasdell, NY 14219) for the afternoon trip. Basemap from Google Earth.

FIELD GUIDE AND ROAD LOG

If you are coming from the Buffalo area, you will want to take 219 South towards New Oregon, NY. Along the drive you will encounter a few Points of Interest (POI):

POI: Armor Duells Rd exit for Chestnut Ridge Park (do not Exit). This park is home to the Eternal Flame Falls, which is a dramatic illustration of the hydrocarbon potential of the Devonian shales that we will observe on this field trip.

POI: Continuing south on 219 near the 55.6 mile marker, you will cross a small tributary of the north branch of Eighteen Mile Creek. You will then start uphill onto the Appalachian Plateau

POI: At the Zimmerman Rd overpass, you will drive past some excellent outcrops of the Canadaway Group shales in a series of roadcuts.

Use the Rice Hill Road Exit, turn right on Rice Rd, then turn left onto Zimmerman Rd and head south, go straight through the intersection with Fennick Rd and Brown Hill Rd (Zimmerman will become Brown Hill Rd), turn right onto Langford Rd and head west into New Oregon.

POI: As you get nearer New Oregon, you will notice several large glacial erratics used as landscaping stones (such as at Boston Fire Co 2)

Meeting Point: Francis Danter Memorial Park in New Oregon (corner of Langford & New Oregon Rds)

Meet at or near the Pavilion *Chemical toilet on site*

Meeting Point Coordinates: 42.5886°N, -78.7913°W

Meeting Time: Saturday, September 25th @ 8:45am (We will leave at 9am. If you are running late and we've already departed, use the Road Log below to locate the group)

Distances in miles		
Cumulative	Point-to-Point	Route Description
0.0	0.0	Turn right out of Francis Danter Memorial Park, heading N on New Oregon Rd
1.1	1.1	POI: Kern 1 well (National Fuel Gas), penetrates to Queenston producing out of Medina SS. Outcrop of Canadaway Group in creek behind well with small waterfall
1.3	0.2	POI: Shear road cut exposure of Canadaway Group. Shoulder is often VERY soft
1.6	0.3	POI: Bridge over creek with some decent outcrop
2.1	0.5	POI: Cement driveway bridge across creek with easy access to pavement outcrops of Canadaway Group
3.4	1.3	POI: Intersection of New Oregon Rd and Clarksburg Rd (Clarksburg Bridge). Would be Stop1, but bridge is out. We will detour around and visit as Stop2
3.8	0.4	**STOP1:** Park on right shoulder in front of garden, just before the New Oregon Rd bridge over Eighteen Mile Creek South Branch

STOP1: New Oregon Rd Bridge, Upper Kellwasser Bed: Frasnian-Famennian boundary

Location Coordinates: 42.62488°N, -78.8380°W

At this location, we will examine the Upper Kellwasser horizon in outcrop. The interval is visible in the cutbank just to the NW of the New Oregon Rd bridge as the creek makes a turn towards the NE and heads towards Sisson Highway. This location is Stop7 of Baird and others (2006) and they provide an in-depth review of the interval, including a measured section, which is reproduced herein (Figure6B). The Upper Kellwasser Bed lies 2.4m below the abrupt contact between the Hanover Member and the overlying black shales of the Dunkirk Member (which we will also see at Stop2). This bed within the upper Hanover marks the Frasnian-Famennian boundary and a major faunal disturbance.

Distances in miles		
Cumulative	Point-to-Point	Route Description
4.2	0.4	Continue on New Oregon Rd to intersection with Route 75 (Sisson Highway). Turn left onto Rt75 heading south

4.2	0.0	POI: Immediately after turning onto Rt75, there is a small tributary on the right with a series of small waterfalls
6.1	1.9	Turn left onto Clarksburg Rd
7.5	1.4	**STOP2:** Clarksburg Country Club. Turn right into driveway (just before bridge) and park along the wood fence near the green bathroom buildings

STOP2: Clarksburg Country Club, Dunkirk Shale ****should* be STOP1, but bridge is out***

Rustic bathrooms available on-site

Location Coordinates: 42.6188°N, -78.8313°W

Unfortunately, the Clarksburg Rd Bridge is out which will force us to detour around. The locals informed me that a propane tanker truck lost its brakes coming down the hill and the driver was forced to bail out before the truck hit the bridge. Fortunately, there were no major injuries aside from the bridge as the engine block tore through the plate steel. The falls here mark the historic location of Simeon Clark's Mill (circa 1820) for whom the town is named. The face of the falls is composed entirely of highly fissile, organic-rich, black shales of the Dunkirk Member.

Distances in miles		
Cumulative	Point-to-Point	Route Description
7.5	0.0	Turn left out of Country Club driveway heading S on Clarksburg Rd
9.0	1.5	Turn right onto Rt75 (Sisson Highway) and head NE
11.3	2.3	POI: Bridge over creek with nice pavement outcrop and small waterfall in creek bed
12.1	0.8	Turn left onto East Church St
12.1	0.0	Almost immediately, turn left onto Old Mill Run Rd
12.4	0.3	**STOP3:** Kromer's Falls. Drive past the two private residences and park at the end of the driveway in front of the out-buildings near No Trespassing signs

STOP3: Kromer's Falls, Lower Kellwasser Bed (Pipe Creek Member)

Location Coordinates: 42.6415°N, -78.8519°W

The falls here also powered an historic mill and the area around it was referred to as Toad Hollow. Kromer's sawmill operated here into the 1980s when the roof collapsed during a winter storm. The foundations of the mill are still visible. This location is Stop6 of Baird and others (2006) and they provide an in-depth review and stratigraphic column which is reproduced herein (Figure6A). Most of the face of the falls is composed of gray silty, bioturbated shales of the Angola Member, but the falls is capped by the well-jointed black shale of the Pipe Creek Member. Bioturbation diminishes upward within the Pipe Creek and it is laminated at the top of the falls, indicative of the major faunal disturbance associated with the Lower Kellwasser Event.

Distances in miles		
Cumulative	Point-to-Point	Route Description
12.5	0.2	Go back out the driveway and turn right onto East Church St
12.5	0.0	Almost immediately, turn left onto Rt75 (Sisson Highway)
13.9	1.4	POI: Bridge over small tributary that exposes some section in a series of small waterfalls up Schintzius Hill
15.2	1.3	Turn left onto North Boston Rd @historic St. Paul's Lutheran Church
16.0	0.8	Turn right onto Eden Valley Rd

16.1	0.1	Go straight through the intersection with Rt62 (Gowanda State Rd) onto Bley Rd
16.4	0.3	POI: Where Mill Rd intersects with Bley Rd, there is an access road on the right that will bring you to the site of Eden Valley Mills (still standing), one of the historic mill locations along the creek. The creek bed is mostly covered through the relatively low-relief Eden Valley.
16.6	0.2	Turn right onto Belknap Rd. This road roughly parallels the creek (it is just beyond the treeline to the right). The gorge deepens dramatically in this area and access to creek level is not easy from above.
18.4	1.8	Gentle right onto Bauer Rd
19.3	0.9	Turn right onto Shadagee Rd
19.9	0.6	POI: We will pass under the Thruway bridge here. If you are ever travelling on the 90 (and not driving!), this bridge provides a good view of the gorge
20.2	0.3	Turn right onto Rt20 (Southwestern Blvd). Note: If you park at the North Evans / St. Vincent de Paul Cemetery, there is a fairly steep footpath (public fishing access) across Shadagee Rd (north side) that leads down to the flat areas at creek level, providing some views of the gorge although it is now mostly overgrown.
20.4	0.2	Cross the bridge and turn right to head E on North Creek Rd. The gorge is now on our right. This time we will go over the Thruway (20.8)
22.1	1.7	**STOP4:** The road winds along the rim of the gorge. After a sign that says "Slides" and a 20mph left bend sign, park facing the "wrong" way on the left shoulder (wider and paved) just as the road starts to bend towards the North. Across the road, there is a very small gap in the retaining barrier that leads to a footpath down to the main branch of Eighteen Mile Creek.

STOP4: The Forks @ Eighteen Mile Creek Park, Rhinestreet Member

Location Coordinates: 42.6969°N, -78.9088°W

The footpath provides a relatively easy descent into the gorge with outcrop in the path itself and along the right-hand side. This is the best opportunity to examine the Rhinestreet Member up close as the gorge walls here are quite steep and inaccessible. Near the bottom of the path, an ephemeral waterfall on a small tributary exposes some large septarian carbonate concretions within the black shales. Lash and Blood (2006) provide an in-depth analysis of this interval, the concretions, and a detailed stratigraphic section that is provided herein (Figure 5). Once you are at creek level, if you walk upstream you will come to The Forks (the confluence between the main branch and the South Branch of Eighteen Mile Creek.) In this area, there are several large concretions that have weathered out of the banks and now rest among the talus and alluvium. This land is currently being developed into Eighteen Mile Creek Park with footpaths providing easy access both upstream and downstream from this location.

****Alternate stops (upstream on main branch of Eighteen Mile Creek)****

If you'd prefer to work your way up the main branch of Eighteen Mile Creek rather than proceed downsection to the lake shore, there are a few interesting stops along the way. If you continue eastward on North Creek Rd you will continue to parallel the gorge, but access from above is very difficult. Make the first possible left (Lakeview Rd) and then turn right onto Lakeview Rd. The Town of Hamburg Recreation Center is near this intersection and there is a moth-balled Cold War Era Nike Missile Base (BU-52) on its grounds. According to the Hamburg Historical Society website, this missile base (and others) protected the Buffalo population center and the steel and iron industries of the area.

The intersection of Lakeview Rd and Old Lakeview Rd is directly across from the Recreation Center. If you take Old Lakeview Rd to the intersection with Smith Rd there is a sharp meander in the creek. Turn right onto the onto South Creek Rd bridge (42.7117°N, -78.8695°W) and to your right you can see a cable stretched across the creek with Posted signs strung across it. Beneath the cable on the SW wall of the gorge, there is a small tunnel (barely high enough to stand) that goes entirely through the meander (do not enter without permission!) at creek level. This tunnel is another callback to the industrial history of the area as it served as the upstream head race that powered a grist mill and the tail race comes out downstream of the falls on the other side. According to the nearby historical marker, this mill was erected in 1806 by John Cummings and was the first grist mill south of Buffalo.
Upstream from this location, towards the town of Hamburg, the gorge becomes shallower and much less steep. The creek is readily accessed from Woodview Park (Woodview Ave) and from Centennial Gardens Park (corner of Main St and Buffalo St). The creek continues behind Eighteen Mile Creek Golf Course and even across Rt219, but outcrop is limited.

Distances in miles		
Cumulative	Point-to-Point	Route Description
22.1	0.0	Make a safe U-turn (be careful of the blind corner). Head back to the W on North Creek Rd
23.8	1.7	Turn left onto Rt20 (Southwestern Blvd), go back across the bridge
24.0	0.2	Just across the bridge, turn right onto South Creek Rd. This road parallels the gorge with a few areas where access is possible (but mostly difficult)
24.3	0.3	POI: If you turn right on Versailles Rd, you will descend to creek level to Hobuck Flats parking area. There is a footbridge across the creek, easy access to creek level, and a short hike (stay on lower path) to a waterfall on a small tributary.
24.9	0.6	POI: We will go under a set of railroad tracks. The creek can be accessed from near the railroad bridge, but it is not recommended
25.8	0.9	Go straight through the intersection with Rt5. There is a way down to creek level at the Rt5 bridge, but this is also not recommended
26.1	0.3	Turn left onto Old Lakeshore Rd
26.2	0.1	Turn right into The Bluffs private community. This location should absolutely not be accessed without prior permission from the residents
26.4	0.2	Continue straight on Old Manor Rd, veer left toward Learmont Dr (follow sign), continue straight on Learmont Dr until the road dead-ends
26.4	0.0	**STOP5:** Park along the fence near the pool (with permission) and head through the unlocked gate (key from resident) and down the footpath to lake level

STOP5: Lake Erie Shoreline, Wanakah Member
Location Coordinates: 42.7143°N, -78.9722°W
At this location, we will be examining the Wanakah Member of the Ludlowville Formation (Hamilton Group). This interval is examined in detail by Over and others (1999) near the mouth of Eighteen Mile Creek, which lies just north of this location. Their map of the stratigraphic contacts along the creek and accompanying stratigraphic section are reproduced herein (Figure4). The lowest stratigraphic member exposed at this location (and often covered by shifting talus) is a light gray, shaly limestone of the Wanakah Member which contains abundant (and often complete) *Eldredgeops* trilobite fossils. This is likely the upper "Trilobite Bed" as described by Grabau (1898), which can also be observed at the mouth of the creek. Above that are more typical shales of the Wanakah and a thin, but laterally continuous band of chocolate brown shaly limestone that is largely unfossiliferous. Above that is more typical shale and then the upper contact with the Moscow Formation (Tichenor Member). The Tichenor forms a

prominent ledge in the cliff face and numerous large blocks of limestone with its characteristic pyrite and rust-staining are strewn across the beach. This rock is used as a building stone in many local buildings, including Frank Lloyd Wright's Graycliff which sits just above the cliff. The rusted steel remnant of the stair tower, which used to provide "easy" access down to lake level, is still standing along the shoreline.

It is difficult to examine the stratigraphically higher units (Windom Member) of the cliff at this location, but I encourage you to use the directions below and join us for Philip Stokes and Holly Schreiber's guided tour of the Penn Dixie Fossil Park & Nature Reserve in the afternoon where they will examine these units in detail.

"END" OF TRIP: Use directions below to Penn Dixie Fossil Park for the Afternoon trip led by Stokes & Schreiber

Distances in miles		
Cumulative	Point-to-Point	Route Description
26.4	0.0	Follow Learmont / Old Manor Dr back to the entrance to the private community
26.6	0.2	Turn left onto Old Lake Shore Rd which parallels the lakeshore
26.8	0.2	POI: Just before the bridge over Eighteen Mile Creek, there is a boat launch on the right with easy access to creek level
29.7	2.9	Make a gentle left onto Rt5 Lakeshore Rd and continue NE along the lake
33.5	3.8	Make a hard right onto Big Tree / New Big Tree Rd (at the Hamburg Clock Tower)
33.6	0.1	Continue straight through the intersection (stoplight) with Francis Dr
34.8	1.2	Continue straight on Big Tree Rd where Bayview Rd veers slightly to right
35.2	0.4	Turn left on Bristol Rd just before Fire Station (small blue Penn Dixie sign)
35.4	0.2	Right left onto North St and follow signs to Penn Dixie entrance
35.5	0.1	**AFTERNOON TRIP:** Penn Dixie Fossil Park & Nature Preserve

BEGIN AFTERNOON TRIP at Penn Dixie Fossil Park & Nature Reserve (led by Stokes & Schreiber)
4050 North St, Blasdell, NY 14219 https://penndixie.org #1 Fossil Park in the U.S.!
Location Coordinates: 42.7765°N, -78.8307°W

ACKNOWLEDGEMENTS

This trip has deep personal meaning to me and my family and I'd like to dedicate it to them. I'd like to thank my grandfather, Albert, who built a home along the Eighteen Mile Creek; my grandmother, Jayne, who nurtured my love of geology and gifted me a reprint of Amadeus Grabau's "Geology and Paleontology of Eighteen Mile Creek"; and my mother, Barbara, who has supported me in everything I've ever done.

I'd also like to thank my grandfather, Gerald, who farmed the banks of the Eighteen Mile Creek and even won Eden's prestigious "Corn King" on more than one occasion; my grandmother, Joyce, who raised 3 sons and worked in the greenhouses; and my father, Paul, who has guided me both in life and along the creek. This trip would not have been possible without the gracious hospitality of the landowners who shared their property, their time, and their stories and I can't thank them enough.

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PENN DIXIE FOSSIL PARK & NATURE RESERVE: A WINDOW INTO THE DEVONIAN PERIOD OF WESTERN NEW YORK

PHILIP J. STOKES and HOLLY A. SCHREIBER

Penn Dixie Fossil Park & Nature Reserve, 3556 Lakeshore Road, Suite 230, Blasdell, NY 14219
E-mail: phil@penndixie.org; holly@penndixie.org

AN IMAGE CAUGHT IN TIME

Earth was very different four hundred million years ago (Figure 1). Eons before the first dinosaurs scuttled through prehistoric forests, North America and Europe together formed a large landmass along the Tropic of Capricorn (Figure 2B). Western New York was submerged beneath tropical seas, which were teeming with both old and new forms of life.

The Devonian Period (419 to 359 million years ago) marked an important chapter in Earth history. Life on land was just starting to take root — literally, as Earth's first forests produced the oxygen needed by the earliest amphibians who hunted giant insects at the water's edge. Along the Eastern half of North America, mountains rose as the Acadian Orogeny (i.e., mountain building event) -- was underway. Like a highway pileup, the Proto-African Plate crashed into and onto the North American plate, causing earthquakes, intense volcanic activity, and crustal deformation (Figure 2C).

Beginning in the middle of the Devonian Period and lasting for 60 million years, The Acadian Orogeny set the stage for our unique assemblage of fossils. To the West of the coastal collision zone, the Earth's crust flexed and bent to accommodate the incoming mass of another continent. Like a compressed accordion, land was forced downwards. Over centuries, the ocean meandered its way along the widening and deepening depression towards New York. As the Eastern mountains grew, the Western basin was inundated with saltwater (Figure 2).

Like a beach slowly submerging, the flooded slopes of this ancient basin transformed dry land into the ideal environment for marine life. Shallow water reefs

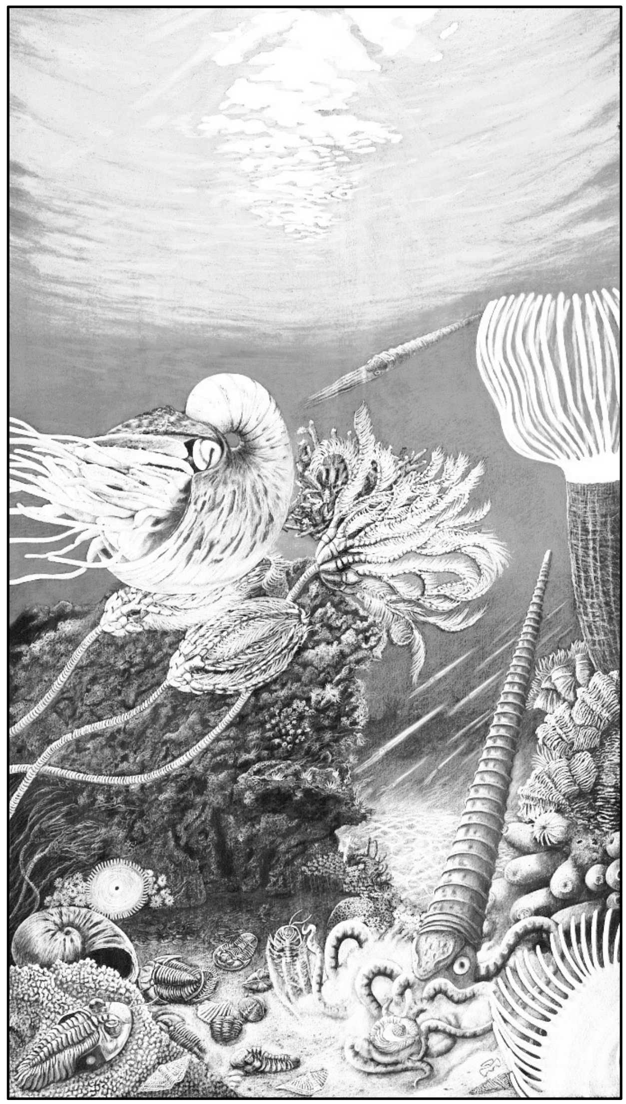

Figure 1 – *Into the Depths of the Devonian* © Mike Menasco. Reconstruction of a Devonian reef based on the fossils found along Eighteen Mile Creek near Penn Dixie.

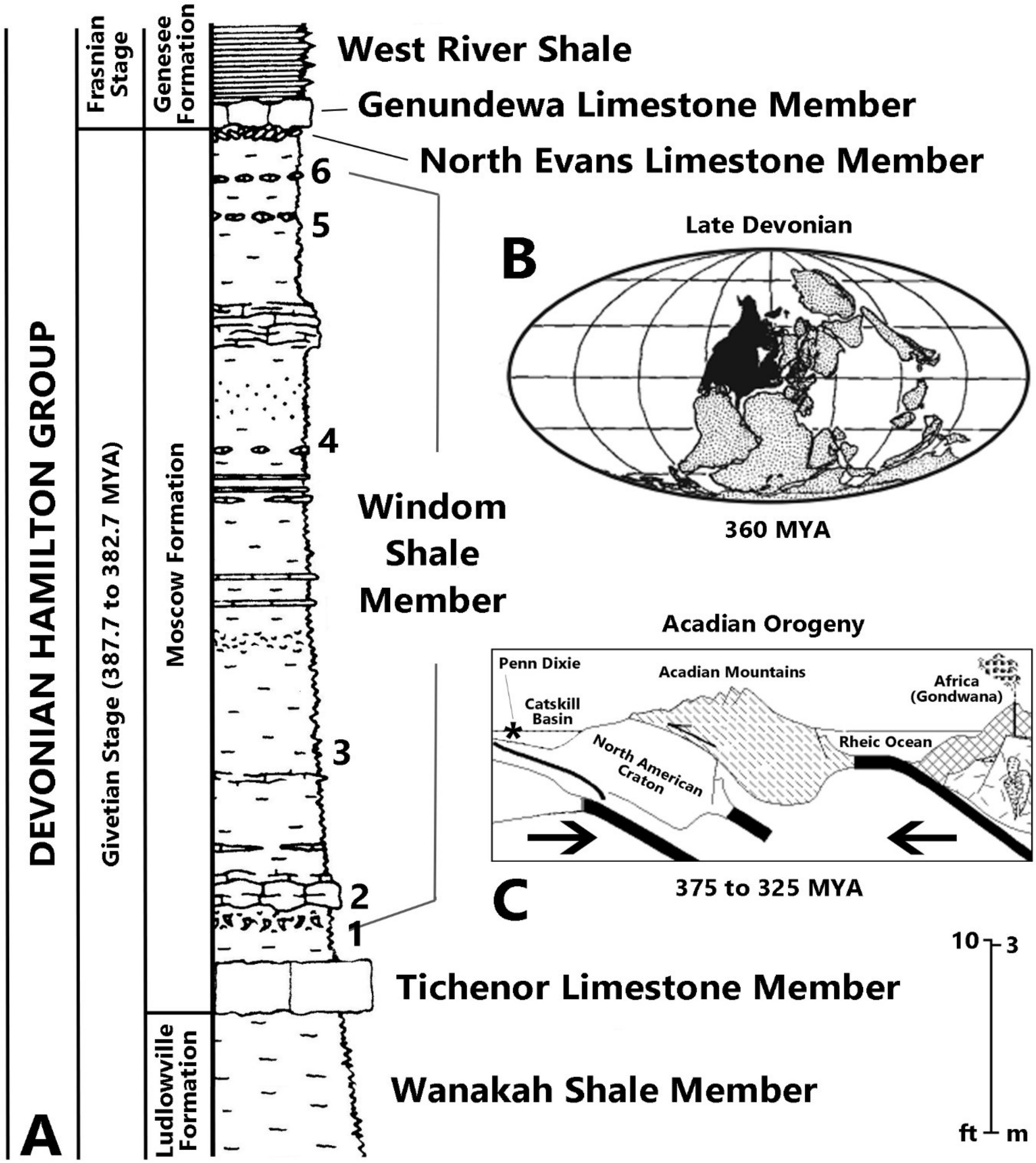

Figure 2 – A) Stratigraphic section of rocks present at Penn Dixie. Numbers 1-6 correspond to units of geological importance. See text for description of members. Modified from Brett and Baird, 1982. B) Paleomap showing locations of the continents during the Devonian Period. North American is in black. Modified from New York State Earth Science Reference Table. C) Cross-section of North America – Africa collision during the Devonian Acadian Orogeny. Location of Penn Dixie is denoted by asterisk. During the Devonian, Penn Dixie was located in the Catskill Basin. Modified from Fichter, 2014.

were dominated by brachiopods and corals. Trilobites — the potato bugs of the ocean — skittered and rolled along the murky bottom as they evaded predators and scavenged what they could. In open water, armored fish, sharks, and the first ammonites — predatory cephalopods in round, chambered shells — feasted on a diverse buffet of newly evolved fish. It was a snapshot of life on Earth -- a geological instant spanning roughly 1% of our planet's long history.

FOUNDATIONS OF STONE

Visitors often ask us why -- if the Earth is covered mostly in water -- do we only find fossils in particular places? Shouldn't fossils be more common? The short answer is that fossils are common, but only in certain places. The long answer is…more complicated.

In *Principles of Geology* (1830-1833), Scottish Geologist Charles Lyell discussed the idea of uniformitarianism -- that "The present is the key to the past." Lyell referenced the modern world as an analogue to the ancient world as he reasoned the stories of rocks and fossils.

Today, life is found globally. However, the conditions for fossilization to occur limit the types of places where the remains of life may fossilize. Four are needed: rapid burial, presence of body parts that will preserve, delayed/prevented decay by microbes, and a geologically stable environment. Penn Dixie met all of these conditions.

Our rocks -- which are geologically consistent across Western New York -- were created from sediments transported westward by ocean currents and waves. Layers of sand became sandstone; silt became siltstone; and clay particles became shale. Limestone, another sedimentary rock, formed from the secretions of skeletal fragments of tiny reef organisms such as corals, snails, crinoids, and plankton. Theses formed the local sequence of geological layers, referred to as stratigraphy, and preserved many types of fossils.

The layers represent varying sea levels in time. In general, sandstone formed in beaches; limestone in warm, shallow water; gray shale in deeper water; and black shale in the deepest, coldest water. Some environments had a much higher biodiversity than others. For example, warm, shallow water supports reef ecosystems with many species. Oppositely, cold, deep water -- with limited oxygen, very few food sources -- cannot support a diverse ecosystem and the rocks that form here are often devoid of fossils.

Fortunately, our local inhabitants possessed many body parts that fossilized, including exoskeletons, shells, and teeth. Though the soft body parts rarely become fossils, we can reconstruct many extinct creatures based on living relatives or internal physiology. For instance, we recognize that extinct horn corals had tentacles just like modern corals.

Preventing decay is pretty straightforward. An animal has to be buried, and thus sealed from outside contaminants, very quickly. The seal must be able to prevent the transfer of oxygen so that bacteria cannot do their work to break down the creature's cells. Charles Lyell might have asked: where do we find these conditions today? Beaches, lagoons, and shallow reef environments are often prime locations for the rapid influx of sediment caused by natural disasters. Storms (e.g., hurricanes) generating underwater avalanches are likely to blame for many of the fossil-rich rocks at Penn Dixie. These rocks -- where animals were brought together in mass mortality and buried -- are called death assemblages.

Finally, a geologically stable environment is needed to ensure that the deceased become fossils. What happens after a fossil is buried? First, in a process called diagenesis, transported sediments are heated and compacted into sedimentary rocks. Fluids flow through microscopic pores in the rock during diagenesis. Inorganic minerals replace organic molecules as water is ultimately driven out of solid rock. Too much heat and pressure results in metamorphism, which destroys fossils. The conditions must be just right during diagenesis.

There are many ways to destroy rocks after they form. Rocks can be uplifted and eroded, subducted beneath tectonic plates, fractured and faulted by earthquakes, deformed by metamorphism, or melted and recrystallized by magmatic fluids. With few geological threats, Devonian rocks survived to became the foundation of our region.

UNDER THE SEA

The sea that covered Erie County and Western New York during the Devonian was relatively shallow and probably had normal marine salinity as evidenced by the diverse and abundant fauna of stenotypic organisms (i.e., tolerant of a narrow range of environmental factors) (Brett & Baird, 1982).

One of the most common and diverse groups of animals during the Devonian period was brachiopods. Brachiopods are bivalved, benthic marine invertebrates. Most live attached to the ground by a fleshy stalk called a pedicle. They are still alive today, but are generally found in deeper, cold waters. Brachiopods eat by filtering organic particles from the water. Brachiopods are often confused with bivalves (members of the Phylum Mollusca) because they both have shells that are comprised of two valves. However, they are very different animals and not closely related at all. The easiest way to tell a brachiopod from a bivalve is the symmetry of their valves. Brachiopods (Figure 3) are bilaterally symmetrical through the two valves, while the two

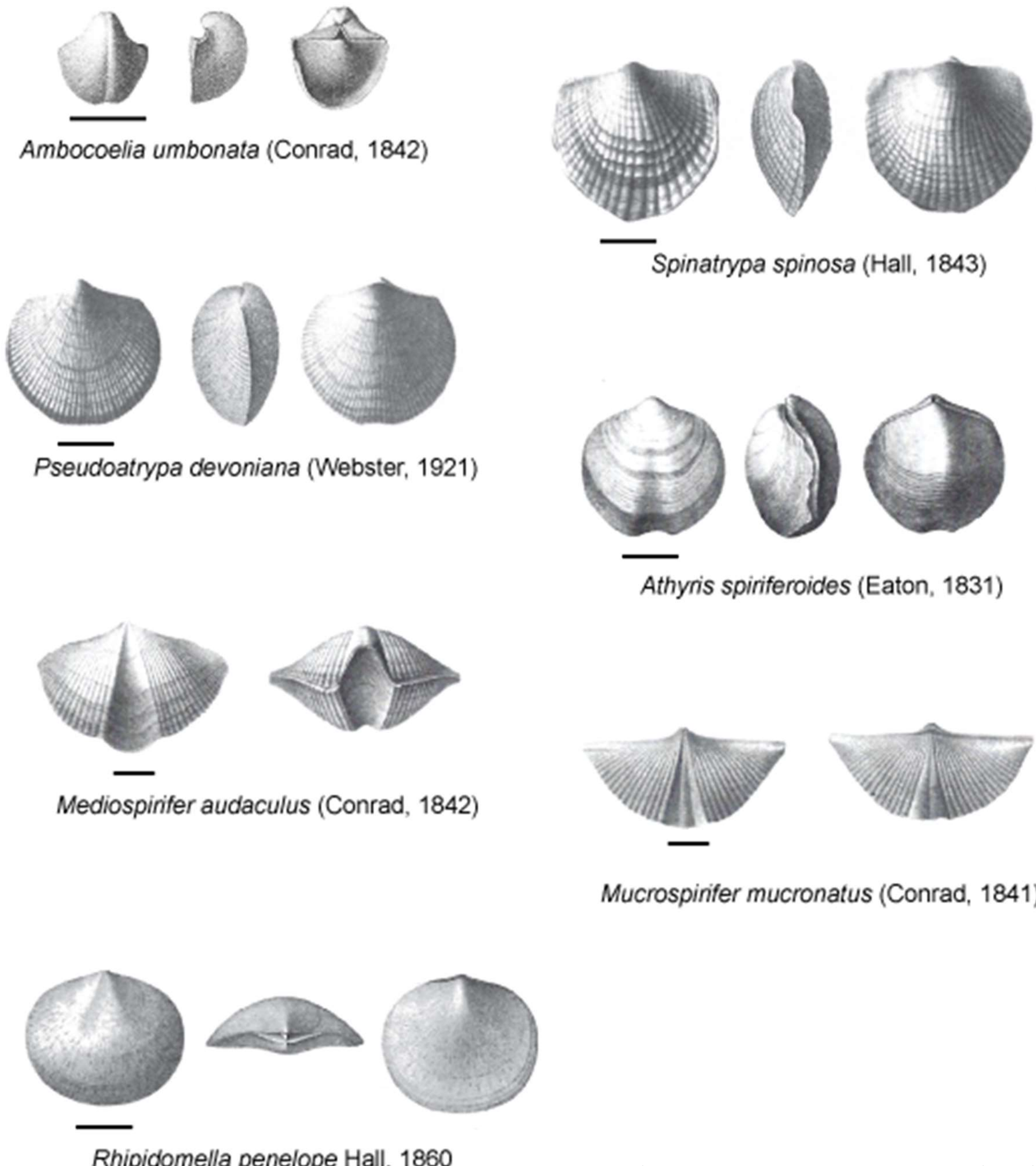

Figure 3 – Most common brachiopods found at Penn Dixie. All scale bars are 1 centimeter. Modified from Wilson, 2014.

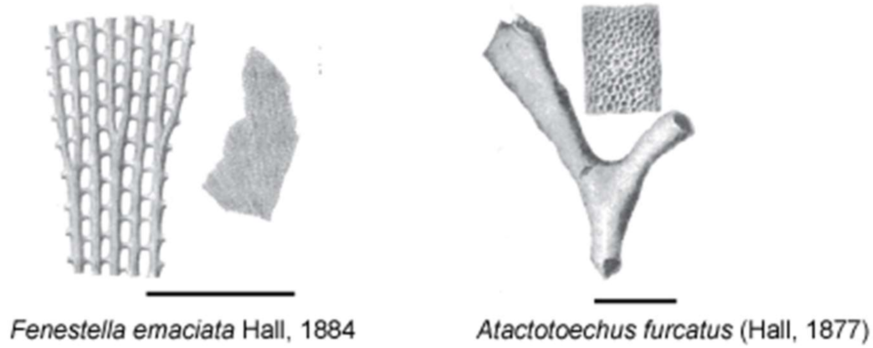

Figure 4 – Bryozoans found at Penn Dixie. Scale bars are 1 centimeter. Modified from Wilson, 2014.

valves of a bivalve are mirror images of each other (Figure 8).

Bryozoans were less common than corals during the Devonian, but can still be found at Penn Dixie (Figure 4). Bryozoans are commonly called moss animals and superficially resemble coral, but are more closely related to the brachiopods. Bryozoans and brachiopods both have lophophores, a fleshy organ used to feed and respire. They are colonial and often grow on rocks or other organisms. Bryozoans are still alive today and can be found in both fresh and salt water.

Many types of corals were found in Devonian seas, just like in the oceans today. These corals are the most common fossils found at Penn Dixie (Figure 5). Most fall into one of two categories: rugose (i.e., horn) corals and tabulate corals. Horn corals are named for their horn-shaped skeletons. Tabulate corals are named for the table-like horizontal structures of their skeletons called tabulae. Both rugose and tabulate corals were very prevalent during the Devonian, forming large, diverse reefs similar to the Great Barrier Reef of today. These two groups of corals are

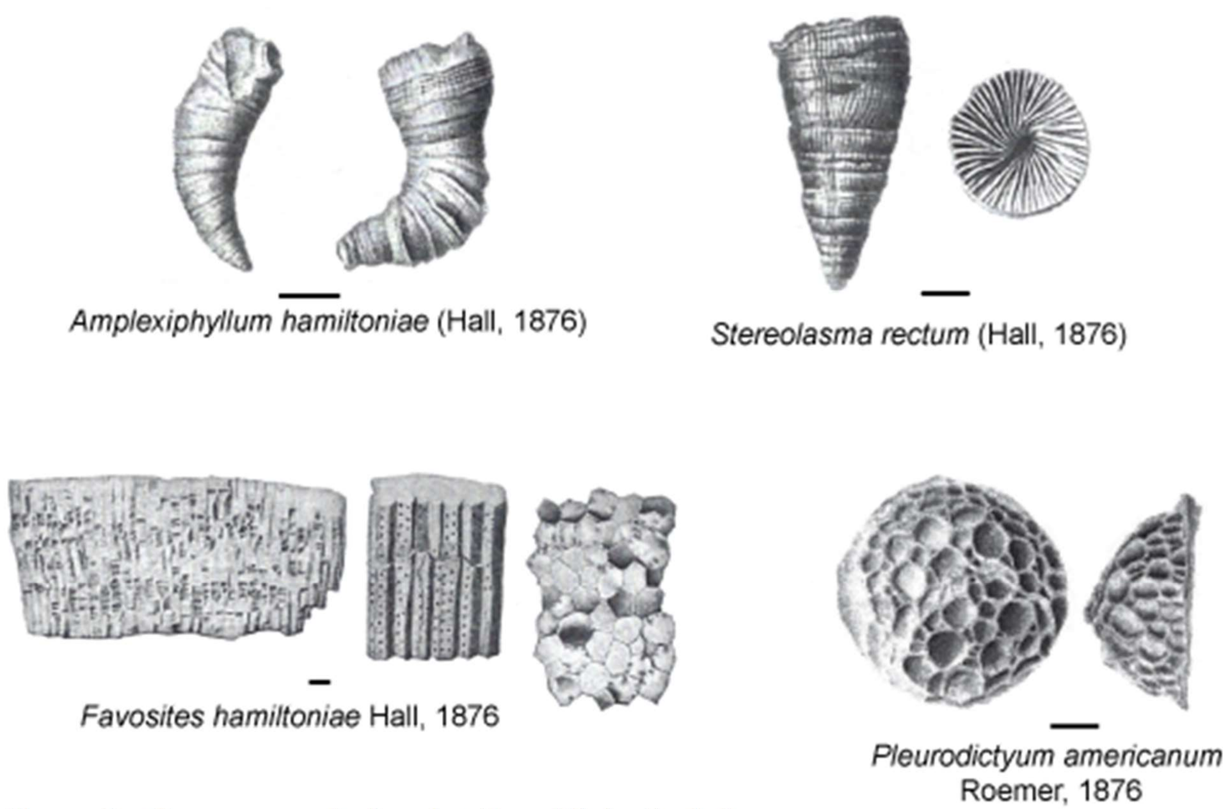

Figure 5 – Common corals found at Penn Dixie. Scale bars are 1 centimeter. Modified from Wilson, 2014.

distant cousins to the living members of Cnidaria. Corals are suspension feeders, grabbing food from the water as it floats past their tentacles. Both rugose and tabulate corals became extinct at the end of the Permian period about 250 million years ago.

Trilobites were also a major component of Devonian oceans (Figure 6). Trilobites are extinct, marine arthropods. They are called trilobites because of their three-lobed body and are sometimes called the potato bugs of the ocean. Trilobites were known for their excellent vision. They were the first animals to evolve true eyes. Earlier "eyes" consisted of photoreceptor cells that sensed light and dark, but could not distinguish shapes. Trilobites had compound eyes (made of many lenses – 17 columns of them!) like insects have today. Their lenses are calcite, and fossilized very well, so we know a lot of information about how trilobites saw their environment. Trilobites went extinct at the end of the Paleozoic Era, but their closest living relative is the horseshoe crab.

Crinoids are arguably one of the most beautiful components of the Devonian seas (Figure 7). Crinoids are members of the Phylum Echinodermata. Think of them as cousins to the starfish, sea urchins, and sand dollars, however they look very different from these other echinoderms. Crinoids are commonly called sea lilies due to their resemblance to flowers (Figure 1). Though they look like flowers, crinoids are animals and grab their food from the water with their feather-

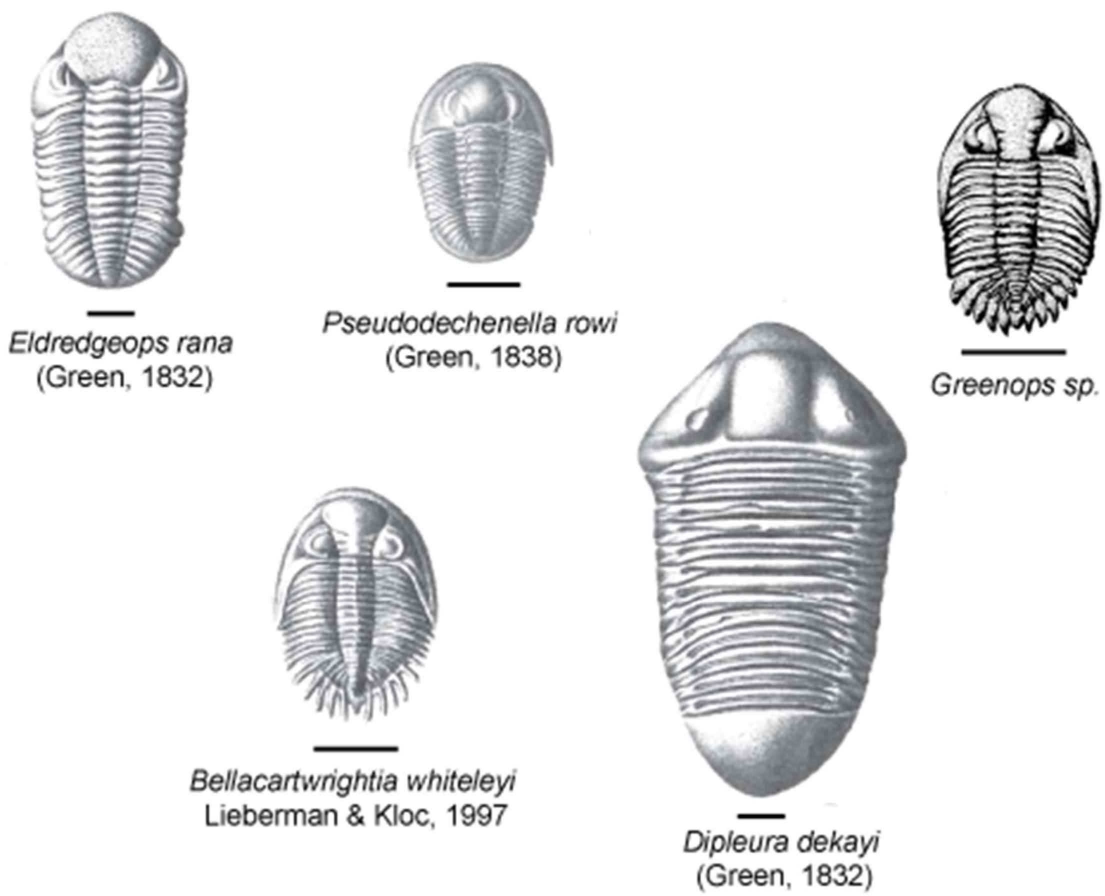

Figure 6 – Trilobites at Penn Dixie. *Eldredgeops* is the most abundant trilobite at Penn Dixie. *Greenops* and *Bellacartwrightia* are uncommon; *Dipleura* and *Pseudodechenella* are rare. Scale bars are 1 centimeter. Modified from Wilson, 2014 and http://www.isgs.illinois.edu/outreach/geology-resources/trilobites.

like fronds. Crinoids are still found in today's oceans, but usually at great depth.

The Devonian oceans was also home to a variety of molluscs, including cephalopods, bivalves (i.e., clams), and gastropods (i.e., snails; Figure 8). Bivalves (sometimes referred to by their former name, pelecypods) are relatively uncommon at Penn Dixie. Most fossils that have two valves found at Penn Dixie are brachiopods (see the discussion above). Bivalvia during the Devonian was far less diverse and abundant than brachiopods. It wasn't until after the late Devonian extinction that bivalves began to overtake brachiopods in diversity and abundance. Cephalopods were some of the top predators during the Devonian. Straight-shelled and coiled-shelled varieties of cephalopods were both common during this time and both can be found as fossils at Penn Dixie. Gastropods are also one of the rarer fossils at Penn Dixie. During the Devonian, marine gastropods tended to be small with variably coiled shells.

Occasionally Devonian plant fragments are found among the rocks of Penn Dixie. These are extremely rare, but should be mentioned because of the unique nature of plants during the Devonian. Plants as we know them today did not exist. Land was first colonized by plants during the preceding period, the Silurian. At the start of the Devonian, small plants (no more than a meter tall) with shallow root systems, dominated the land. By the end of the Devonian, ferns, horsetails and seed plants evolved (Figure 9). Most of these plants had true roots and leaves, forming the first trees and forests. Because large land herbivores had yet to evolve, plants expanded to new forms and colonized the land unchecked.

In Devonian seas, fish were also some of the top predators (along with cephalopods). But, the fish of the Devonian looked very different from the of fish today. Armored jawless fish (i.e., ostracoderms) were common. These fish generally lived along the ocean floor and had elaborate, armored exteriors, but lacked jaws.

By the middle Devonian, fish had evolved jaws. These first jawed fish were called placoderms, for their armored plates. The most famous placoderm was the ten meters long *Dunkleosteus terrelli* (Newberry, 1873; Figure 10). In Western New York, the Devonian oceans were ruled by its cousins, *Eastmanosteus magnificus* (Hussakof and Bryant, 1918) and *Dinichthys hertzeri* Newberry, 1873.

At Penn Dixie, isolated pieces of the armored plates and bone fragments of these fish can be found, most often in the limestone layers above the Windom Shale. It is nearly impossible to identify the specific species to which these pieces belong. Bony fish also evolved during the Devonian, one lineage of which evolved into land-living tetrapods from which humans descended.

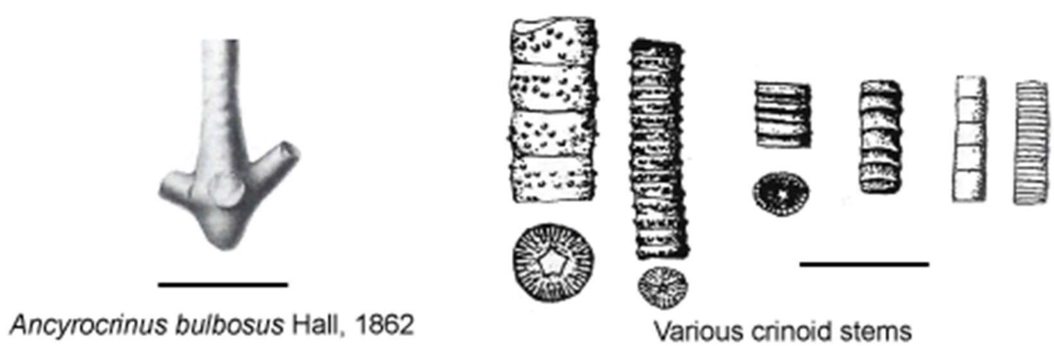

Figure 7 – Crinoids found at Penn Dixie. Scale bars are 1 centimeter. Modified from Wilson, 2014 and Grabau, 1898.

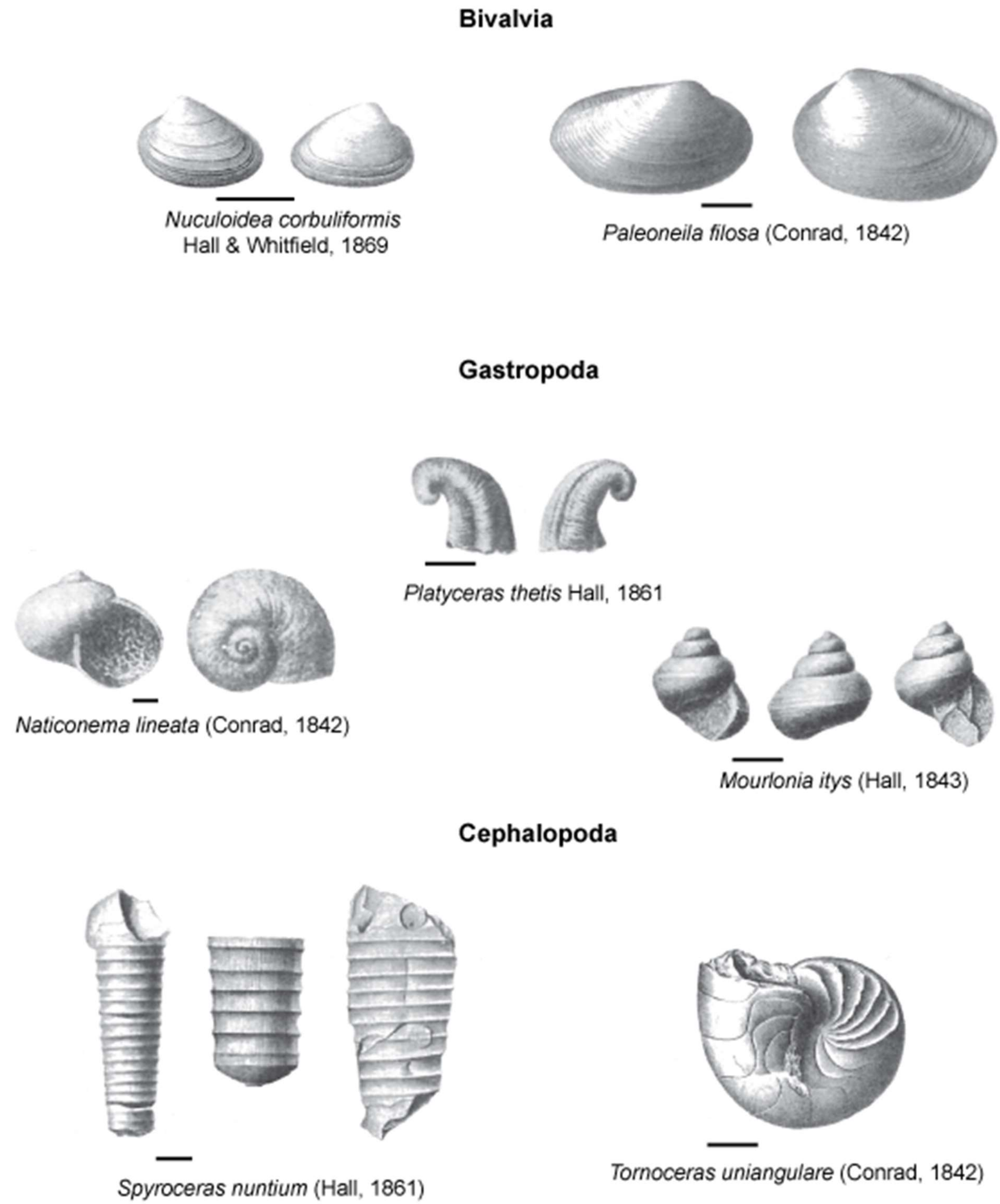

Figure 8 – Common molluscs at Penn Dixie. Scale bars are 1 centimeter. Modified from Wilson, 2014.

Figure 9 – *Vegetation of the Devonian period, restored* from Dawson (1888).

TURN TO STONE

Wanakah Shale – This is the uppermost unit of the Ludlowville Formation (Figure 2A) and is exposed in the northeast corner of Penn Dixie in the creek bed. The unnamed creek bed in the northeast section of the site is a tributary of Rush Creek. The uppermost meter of the Wanakah Shale is exposed at Penn Dixie. This thinly-bedded unit is gray in color and is often highly fractured where it is exposed in the tributary. Large concretions of muddy limestone within the unit are noticeable where erosion has removed the softer shale. These concretions often contain fossils and are thought to be formed by the rapid decay of organisms. The Wanakah Shale was deposited in moderately deep water.

The exposed portions of the unit are fossil rich. Bryozoans, trilobites, gastropods, bivalves, echinoderms, corals, sponges, and ostracods can be found. Brachiopods are especially abundant and the primary component of the fossils in the Wanakah. Common species include *Mediospirifer audaculus* Conrad 1842, *Mucrospirifer mucronatus* Conrad 1841, *Athyris spiriferoides* Eaton 1831, and *Ambocoelia umbonata* (Conrad, 1842). Water collects in this creek bed throughout the year, so fossils found here can be quite delicate and break apart easily. Flowing water and wet leaves can also make collecting in this area difficult.

Tichenor Limestone -- The Tichenor Limestone, the base of the Moscow Formation, overlies the Wanakah Shale and is approximately a meter thick at Penn Dixie (Figure 2A). The Tichenor is a prominent layer due to its off-white color and its resistance to weathering. Large blocks of the Tichenor are present where excavation has taken place. Erosion at the site has exposed bedded Tichenor Limestone along drainage channels. Exposed areas of the Tichenor darken in color over time due to chemical weathering. Portions of this unit are almost entirely fossiliferous and consist of microfossils and fragments of crinoids and other invertebrates.

The Tichenor represents a near-shore, shallow water environment with well-circulating waters. These currents allowed for rapid burial, but also mixed previously deposited remains. Oxidation of the upper surface of the Tichenor suggests that the unit was exposed to air prior to diagenesis. This exposure was likely the result of falling sea level and created an unconformity between the Tichenor and the overlying Windom Shale. The Tichenor Limestone, which rises and plunges within Penn Dixie, illustrates the folding of rocks by some yet unknown process.

The Tichenor Limestone contains diverse and abundant fossils, though they are difficult to remove from the hard limestone. Numerous horn corals (*Sterolasma rectum* Hall 1876; *Amplexiphyllum hamiltoniae* Hall 1876) and tabulate corals (*Favosites hamiltoniae* Hall 1876; *Pleurodictyum americanum* Roemer 1876) can be found throughout. Brachiopods, bivalves, and trilobites can also be found. Bryozoans (*Fenestella emaciata* Hall, 1884) are also plentiful in the Tichenor, one of the easiest places to find them at Penn Dixie. Pyrite nodules, often weathered to a reddish-brown color, can also be found throughout the unit.

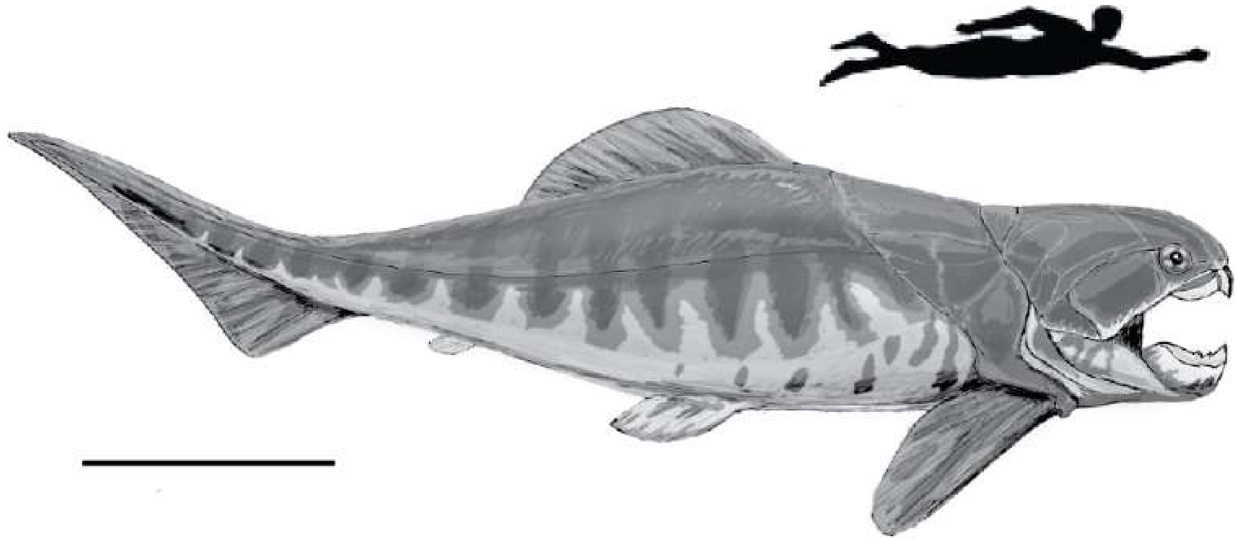

Figure 10 – Reconstruction of *Dunkleosteus terrelli* (Newberry, 1873) with human for scale. Scale bar is 1 meter.

Windom Shale -- The literature (e.g., Brett and Baird, 1982) describes many distinct beds within the Windom Shale; here, we discuss the most relevant beds to fossil collectors and educators (Figure 2A).

1) Bayview Coral Bed -- The Bayview Coral Bed is near the bottom of the Windom Shale. This layer is a soft, gray shale that is easily weathered. Excavation is required to reach this unit since surface exposures rapidly break down. Most likely a death assemblage, this unit lacks cohesive strength from smaller sediment grains (e.g., clay and silt). Our interpretation is that the smaller grains were washed away by currents -- disrupting the burial process, leaving behind only large and incomplete fossils.

The fauna of this bed is incredibly diverse with at least 50 species represented. Brachiopods are the most common and diverse fossils in this bed including *Rhipidomella penelope* Hall, 1860, *Mediospirifer audaculus* Conrad 1842, *Mucrospirifer mucronatus* Conrad 1841, *Athyris spiriferoides* Eaton 1831, *Spinatrypa spinosa* Hall 1843, *Pseudoatrypa devoniana* Webster 1921, and *Ambocoelia umbonata* (Conrad, 1842). Large (e.g., *Heliophyllum halli* (Edwards & Haime, 1850)) and small rugose corals are also very common throughout this bed. The exposure of this bed at Penn Dixie is one of the few places where the Bayview Coral bed can be easily accessed in Western New York.

2) Smoke Creek Trilobite Bed -- The Smoke Creek Trilobite Bed sits atop the Bayview Coral Bed. It is a distinctive marker bed within the Windom Shale. This layer is a dark gray, calcareous (i.e., rich in calcite) shale unit that is resistant to weathering. Large blocks of the Smoke Creek Trilobite Bed are challenging to split as the bedding is thick and massive. The unit has a distinctive oily odor due to the presence of hydrocarbons.

The Smoke Creek Trilobite is widespread in Erie and Genesee Counties and has a uniform thickness of 20-75 centimeters (Brett and Braid, 1982). This suggests a relatively stable geologic environment when compared to thinner beds within the Windom Shale.

As the name implies, it is rich in trilobites with at least five species represented, including *Eldredgeops rana* Green 1832 (often mistakenly called *Phacops rana*), *Greenops sp., Bellacartwrightia whiteleyi* Lieberman & Kloc 1997, *Pseudodechenella rowi* Green 1838, and *Dipleura dekayi* Green 1832. *Eldredgeops* is the most common trilobite found at Penn Dixie. It is commonly found either prone or enrolled. *Eldredgeops* individuals are also often found in clusters. In addition to trilobites, rugose corals, brachiopods, bivalves, and cephalopods can be found in the Smoke Creek Bed.

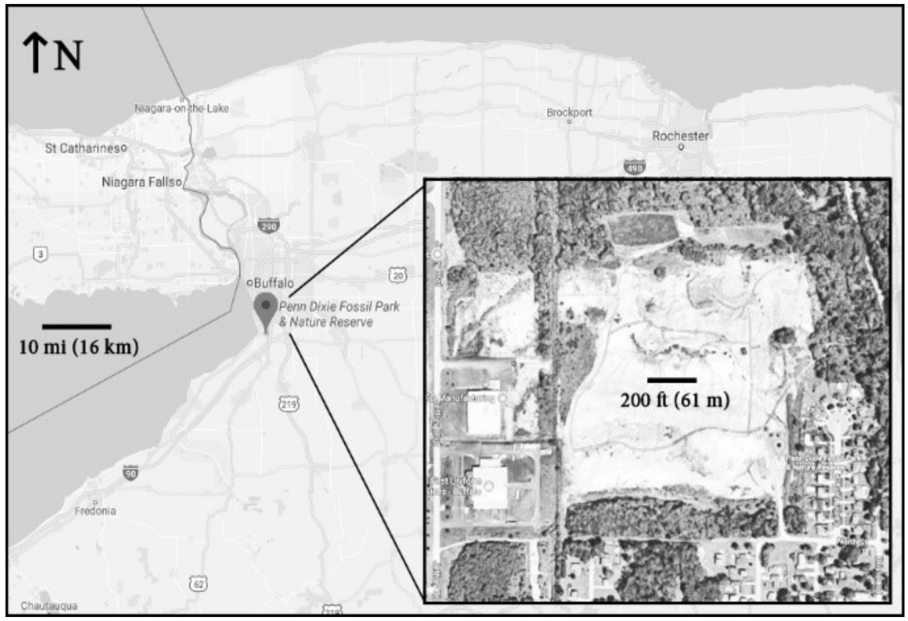

Figure 10 – Location of Penn Dixie, south of Buffalo, New York. Modified from Google Maps.

3) Barren Zone -- Shales that overlie the Smoke Creek Bed are lighter in color and are sparsely fossiliferous or completely devoid of fossils. This unit has been interpreted as a low oxygen environment, most likely deeper water.

4) Mid to Upper Windom Shale -- The Mid to Upper Windom Shale is sparsely fossiliferous, just like the earlier Barren Zone. This portion of the Windom Shale was exposed by original quarrying operations and is relatively unchanged from that time. This interval -- which contains many thin, interbedded layers of shale -- is known for a series of three regularly-spaced white limestone bands. Some portions are fossiliferous, but not to the degree of other beds in the Windom Shale. In the Upper portion, small brachiopods such as *Ambocoelia umbonata* (Conrad, 1842) can be found. As a whole, the mid to upper Windom Shale represents varying sea level within the local basin. Limestone bands likely represent shallow water; shales are deeper water deposits.

5) Penn Dixie Pyrite Beds -- Near the upper portion of the Windom Shale, a pronounced pyrite-rich gray shale layer is exposed. At Penn Dixie, fossils are commonly found in this interval pyritized and therefore are golden in color. Pyrite weathers to a brownish-red color when exposed at the surface, so many fossils have this weathered color as well. A variety of fossils can be found in this pyritized layer, including nuculoid bivalves, gastropods, ammonoids and enrolled trilobites (the golden trilobite!) Most fossils from this layer are diminutive in size, but well preserved. Pyritization of fossils usually occurs in deep, anoxic conditions.

6) *Ambocoelia* Bed -- The *Ambocoelia* bed is named for the abundance of the brachiopod *Ambocoelia umbonata* (Conrad, 1842). *Ambocoelia* is the most common fossil found in this gray shale layer. Other fossils include rugose corals and the occasional bivalve. Trace fossils that have been interpreted as worm coprolites are also common in this layer. The *Ambocoelia* bed tends to be much more fossiliferous that the Barren Zone or the Mid to Upper portion of the Windom, but the fauna is dominated by *Ambocoelia*.

North Evans Limestone -- This unit is a mottled beige to dark gray color with a granular texture despite the calcite composition (Figure 2A). The rock contains clasts of inorganic materials (e.g., inclusions of Windom Shale) and organic materials (e.g., microfossils) mixed together during deposition in shallow waters. The limestone is not well exposed at Penn Dixie due to ample vegetation, but the weathered surface can be accessed in the southern portion of the site. Wood, fish plates, teeth, bone, and mandibles can be found in the North Evans Limestone, but are extremely rare.


Figure 11 – Penn Dixie Cement Plant along Lake Erie, south of Buffalo, New York in 1959. Photo courtesy of the Hamburg Historical Society.

Genundewa Limestone -- This unit is thinly bedded muddy limestone and is dark gray in color due to clay impurities (Figure 2A). It is poorly exposed along the densely vegetated, southern ridge of Penn Dixie, but weathered blocks wash down slope and mix with North Evans pieces. The Genundewa fossils are similar to those found in the North Evans Limestone. The units are separated by an unconformity, indicating a pause in shallow water deposition.

West River Shale -- This unit is dark gray in color, weathers easily, and has essentially turned into the soil supporting the forested ridge along the southern boundary of the site (Figure 2A). No natural exposures exist at the surface but can found elsewhere in Western New York.

DIGGING IN THE DIRT

Though glaciers scoured the Western New York surface long after dinosaurs became extinct, human activities had the most noticeable effect on our geology. Beginning in the early 1950s, the rocks at Penn Dixie (Figure 10) were quarried for the making of Portland cement, which is the main ingredient in the concrete commonly used in construction (Streamer, 1988). Heavy machinery ripped off the surface layers -- which are rich in calcite and clay particles -- and crushed the rocks into smaller pieces. The rocks were transported approximately one mile away to a facility along the shore of Lake Erie for processing into cement. The proximity to railroads and water made our location ideal for this activity (Figure 11).

The Penn Dixie Cement Corporation was the third entity to operate a quarry at our location; Federal Cement and Bessemer Cement were the two prior. At one time, the combined plant and quarry were the third largest industrial employer in the Town of Hamburg! From the 1950s to the 1970s, Penn Dixie Cement, whose name was shortened from Pennsylvania-Dixie Cement, owned numerous plants and offices in the Eastern U.S. However, a number of factors led to the shutdown of active quarrying in our location by the early 1970s. The end came for the entire Penn Dixie Cement Corporation in 1980 when the company filed for bankruptcy (Streamer, 1988; NYT, 1981).

For 20 some years, the quarry changed hands several times through private ownership while remaining undeveloped. During this time, trespassers used the open land for the illegal dumping of household appliances and vehicles, hunting and shooting, and revelry. The former quarry also became known in the fossil collecting and educational communities due to its profusion of well-preserved fossils. When development threatened to turn the quarry into a waste transfer station, a group of local activists coalesced around the goal of preserving the land for science education (Bastedo, 1999).

Formed in 1993, The Hamburg Natural History Society, a 501(c)3 nonprofit, successfully lobbied the Town of Hamburg to purchase the land and donate it to the society. In turn, The Society agreed to remove accumulated waste and to develop the property so that it could become a unique destination for science education and tourism. Since its formation, The Society has grown from an all-volunteer organization to a cultural institution embedded in the community fabric of Western New York. Annual attendance averages 18,000 with one in eight visitors traveling from out-of-state -- or outside the U.S. -- to collect fossils (Figure 12). The Society delivers on its mission of hands-on science education by offering engaging, inquiry-based field trips for K-16 students and summer programming for youth and families.

CLOSER TO THE HEART

Fossil collecting -- once a wealthy gentleman's hobby in Georgian England -- is now enjoyed by many.

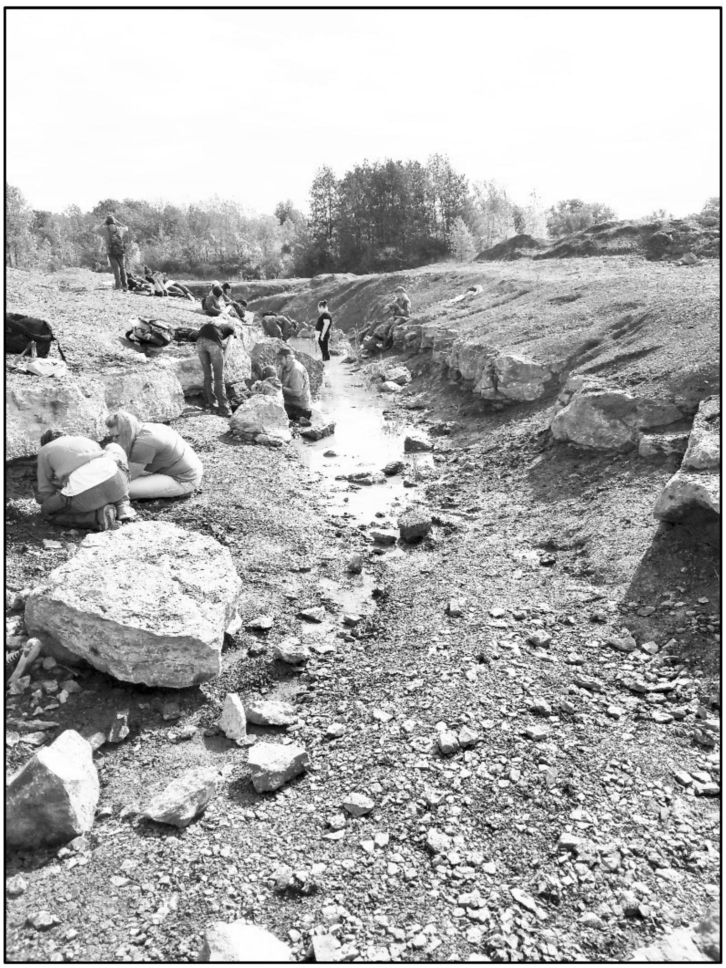

Figure 12 – Visitors digging at Penn Dixie during the 2017 season.

Happily, our mission extends beyond this one activity; it serves to educate the public about the past, to engage the imagination while looking back in time, to improve understanding of scientific and environmental issues affecting society, to promote tourism, and to inspire youth to science, technology, engineering, and math (STEM) careers.

Informal science education, which includes trips to science and nature centers, is an important complement to formal science education in a K-12 setting. These sorts of experiences educate and motivate youth to pursue more STEM knowledge and build confidence and self-efficacy. For example, outdoor experiences -- and particularly those shared with family members -- are frequently cited by geoscience majors as critical factors in their choice of major (Stokes et al., 2015). Can you imagine what problems society might solve if students had more opportunities for hands-on learning?

We are very proud of the positive impact that we have on our young visitors. Penn Dixie is a unique treasure. In 2011, we were honored to be recognized as the top fossil park in the U.S. in a scientific study (Clary and Wandersee, 2011). Of the parks studied over seven years, the authors wrote that Penn Dixie "provided the best visitor experience and the greatest opportunities to learn geobiological concepts in an informal fossil park environment (p. 131)." We are honored by this recognition and continue to develop and refine our goals to maintain this ranking.

In their concluding remarks, Clary and Wandersee wrote that "there is no equivalent virtual substitute for direct interaction with Earth, and fossil ownership sparks thought about deep time, evolution of life forms, and environmental change over geologic time (2011, p. 132)." We wholeheartedly agree. Albert Einstein, perhaps the most influential scientist of the past 100 years, propelled innovation and discovery across many fields. His observations led to a radical new awareness of our universe, yet the simplicity of his theoretical writings is often overlooked. “Look deep into nature,” Einstein wrote, “and then you will understand everything better.” We hope you’ll look with us.

ACKNOWLEDGEMENTS

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Finally, we thank the citizens and governmental leaders of Erie County for having the vision and wisdom to turn an old, abandoned quarry into a global geological treasure.

For the latest updates, visit www.penndixie.org.

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APPENDIX 1 – Abbreviated and in-progress list of species found at Penn Dixie. Modified from Brett (1974).

SUMMARY CHART
OF THE DISTRIBUTION
OF FOSSILS IN ROCK
LAYERS PRESENT AT
PENN DIXIE

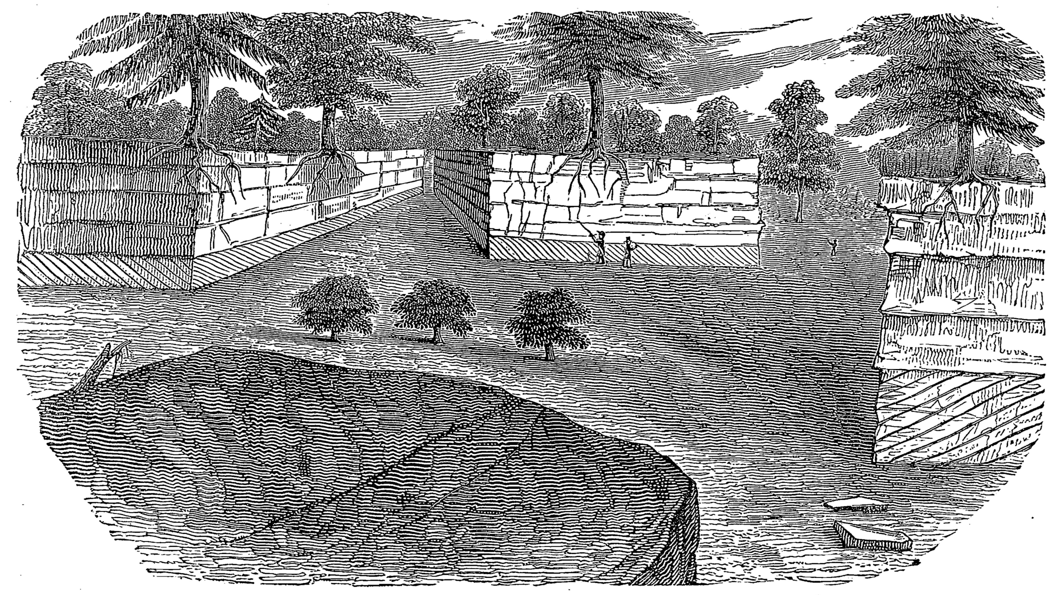
C = Common
R = Rare
L = Locally abundant,
but rare elsewhere

	Wanakah Shale (A)	Tichenor Limestone (B)	Windom Shale: (1) Bayview Coral bed (C)	Windom Shale: (2) Smoke Creek Trilbite bed (D)	Windom Shale: (4) Mid to Upper Windom (E)	Windom Shale: (5) Pyrite bed (F)	Windom Shale: (6) *Ambocoelia* bed (G)	North Evans Limestone (H)
Cnidaria:								
Amplexiphylloides hamiltonae (Hall, 1874)	C			C				
Aulocystis dichotoma (Grabau, 1899)			C					
Aulocystis jacksoni (Grabau, 1899)			C	C	C			
Cystiphylloides americanum (Edwards & Haime, 1851)			L					
Cystiphylloides conifollis (Hall, 1876)			L					
Hadrophyllum woodi Grabau, 1899			C					
Heliophyllum halli (Edwards & Haime, 1850)			L					
Heterophrentis simplex (Hall, 1843)			L					
Pleurodictyum americanum Roemer, 1876		L	L	C		R		
Stereolasma rectum (Hall, 1876)	C		C	C		R		
Streptalasma ungula Hall, 1876	C			C				
Favosites hamiltonae Hall, 1876		L						
Bryozoa:								
Fenestella emaciata Hall, 1884		L						
Leptotrypella ssp.			C					
Reptaria stolonifera Rolle, 1851						R		
Sulcoretipora incisurata (Hall, 1881)								
Atactoechus furcatus (Hall, 1877)								
Echinodermata:								
Arthrocantha sp.				R				
Deltacrinus clarus (Hall, 1862)		C	R	R				
Dolatocrinus liratus Hall, 1862			R					

	(A)	(B)	(C)	(D)	(E)	(F)	(G)	(H)
Brachiopoda:								
Amocoelia umbonata (Conrad, 1842)	C		C	L		C	C	
Athyris spiriferoides (Eaton, 1831)	C		C					
Atrypa reticularis (Linneaeus, 1758)	C		C					
Emanuella praeumbonata (Hall, 1857)						R	C	
Longispina mucronatus (Hall, 1843)			R					
Mediospirifer audaculus (Conrad, 1842)	C		C					
Megastrophia concava (Hall, 1857)			R					
Mucrospirifer consobrinus (D'Orbrigny, 1850)			C					
Mucrospirifer mucronatus (Conrad, 1841)	C		C					
Protoleptostrophia perplana (Conrad, 1842)			R					
Pseudoatrypa devoniana (Webster, 1921)	C		C					
Rhipidomella penelope Hall, 1860			C	C				
Rhipidomella vanuxemi (Hall, 1857)			C	C				
Spinatrypa spinosa (Hall, 1843)	C		C					
Spinocyrtia granulosa (Conrad, 1839)			C					
Tropidoleptus carinatus (Conrad, 1839)			R					
Mollusca:								
Gastropoda:	R		R					
Mourlonia itys (Hall, 1843)			R	R				
Naticonema lineata (Conrad, 1842)			R	R				
Platyceras thetis Hall, 1861								
Bivalvia:								
Cypricardinia indenta Conrad, 1842				R				
Nuculoidea corbuliformis Hall & Whitfield, 1869	R			R				
Nuculites triqueter Conrad, 1841				R				
Palaeoneilo filosa (Conrad, 1842)			R					
Palaeoneilo sp.			R					
Pterinopecten sp.				R				
Cephalopoda:								
Michelenoceras ssp.	R		R	R				
Spyroceras nuntium (Hall, 1861)		R				R	R	
Tornoceras uniangulare (Conrad, 1842)						R	R	
Trilobita:								
Bellacartwrightia whiteleyi Lieberman & Kloc, 1997				R				
Dipleura dekayi (Green, 1832)		R		R				
Greenops sp.	R		R	C				
Eldredgeops rana (Green, 1832)	C		R	C				
Pseudodechenella rowi (Green, 1838)		R		R				

SALAMANCA CONGLOMERATE

A RECORD OF STORMS, FLOODS, AND TIDES IN DELTAIC-COASTAL DEPOSITS OF THE UPPER DEVONIAN CATTARAUGUS FORMATION



View in Rock city, seven miles south of Ellicottville. From a sketch by Mr. E. N. HORSFORD.

(Geology of New York, 1843)

JAMES H. CRAFT

Engineering Geologist (retired), New York State Dept. of Environmental Conservation

PREFACE

A version of the following article was prepared after a brief field season for the 2017 New York State Geological Association fall meeting (http://www.nysga-online.net/). Ongoing fieldwork has continued over the last four years and the use of DEMs/GIS has located and correlated Salamanca (and other conglomerates) outcrops over an area of 2000 km^2. The overall interpretation (prograding delta - flood dominated with wave & tide-influence) is the same with some refinements for the type section ("Little Rock City") and near-field areas (Rock City State Forest), chiefly:

- recognition of major channel/bar complexes (up to 5 m thick) interpreted as delta distributaries and tidal channels) overlying marine strata on the western outcrop flank. Also, a large mouth bar was recognized at the base of outcrop #3 as the Rim Trail turns south);

- recognition of pebbly hummocky cross-stratification (HCS coarser than fine sand is uncommon in the rock record). Paleohydraulic estimates of large-scale forms (largest hummock wavelengths > 7 meters; largest pebbles > 5 cm at outcrops # 4 & 5) suggest large waves, combined flows, and energetic storms/hurricanes, in particular, at the top of the sequence.

- recognition of twin conglomerate layers in Allegany State Park separated by 10-15 m of apparent shallow marine strata. The lower unit correlates with the Salamanca and the upper unit is very similar in scale and depositional motif (bi-directional X-strata and HCS) suggesting a re-advance of the paleoshoreline (two Regression-Transgression cycles over ~ 25m-35m formation thickness). Other than the basal Wolf Creek conglomerate, the other historical/locally-named Devonian conglomerates (e.g., Pope Hollow, Tuna Creek, Irish Hollow))appeared to correlate with this pair of Salamanca conglomerates. Other shoaling-upward units are being examined in the Park (largely sandstone, ~ 1-2 m; locations include Sugarbush Cabin loop, Camp Allegany, and Angel Falls).

- recognition of a similar conglomerate of presumed Mississippian age (correlates with Mississippian formations to the east and west and is stratigraphically ~ 100m above the Salamanca in Allegany State Park). Two or three tabular beds of ~ one meter thickness contain abundant coarse-grained HCS with flat pebbles. GIS projection of its caprock plane suggests a much more extensive outcrop of Mississippian rocks in Allegany State Park than currently mapped.

INTRODUCTION

On a hilltop in Rock City State Forest, three miles north of Salamanca, New York, the Salamanca Conglomerate outcrops in spectacular fashion. Part of the Upper Devonian (late Fammenian) Cattaraugus formation, the quartz-pebble conglomerate forms a five to ten-meter high escarpment and topographic bench at ~ 2200 feet elevation amid a mature cherry-maple-oak forest. In places, house-sized blocks have separated from the escarpment along orthogonal joint sets and variably "crept" downhill. Where concentrated, a maze of blocks and passageways may form so-called "rock cities", an impressive example of which is Little Rock City. The well-cemented blocks permit extraordinary 3-D views of diverse and ubiquitous sedimentary structures and features.

Seven outcrop areas with the most significant exposures were logged over a four-kilometer north-south traverse. The traverse largely follows the east-facing hillside which roughly parallels the presumed paleo-shore of the Devonian Catskill Sea. Extensive "bookend" outcrops at the north face (off the Rim Trail) and at the southeast perimeter ("Little Rock City" along the North Country-Finger Lakes Trail) and vertical (caprock) control allow a nearly continuous look at spatial and temporal changes in sedimentary deposits along a four-kilometer stretch of inferred late Devonian seacoast.

Summary of Findings

Three major depositional environments, reflecting a high-energy and varied coastline, are interpreted from north to south:

Shoreface to foreshore (beach) to channel deposits coarsening-upward sequence - ("north face" **Outcrops #2** from base)

- ~ 2 m of thin-bedded (5-10 cm) wave cross-laminated and small-scale (< 0.5 m) HCS strata; a coset of cross-strata mostly buff, medium sand with some coarse sand, granules, and a few fine pebbles.
- ~ 3 m of amalgamated coarse-grained (1 mm -10 mm) , large (10-20 cm x 50-100 cm) hummocky cross-strata (HCS) interbedded in places w/thin fine-grained (rolling-grain) wave ripples; some trough/planar cross-beds near top.
- ~ 2-m of parallel/low-angle strata of gray interbedded coarse sand and some pebbles.
- ~ 3 m of point bar/lateral accretion/channel deposits.
- pebbly caprock with some hummock forms.

Prograding flood-dominated delta (w/coarse-grained distributaries, mouth bars, tidal channels, bars, and shoals) **(Outcrops # 2, 3, 4, 5, 6 & 7)**

- abundant channels (two main distributaries ≈ 5 m deep, others, ~1-2 m deep) and channel point bars (coarse sand to pebble lateral-accretion deposits of tidal, delta distributary/fluvial channels). Two channel complexes directly overlie fine-grained wave ripple-laminated (marine) sandstones.
- large coarse sand/pebbly hummocky cross-stratification (HCS coarser than fine sand is uncommon in the rock record). Paleohydraulic estimates of large-scale forms (largest hummock wavelengths > 7 meters; largest pebbles > 5 cm at outcrops # 4 & 5) suggest large waves, combined flows, and energetic storms/hurricanes, in particular, at the top of the sequence.
- cross-bedded strata of various dimensions (~ 0.05 m to +1 m), some bidirectional.

- current indicators mainly directed shoreward (E-SE) and alongshore (S) , bi-directional cross-beds common in places; some truncation surfaces show wave influence.
- pebbly caprock with some hummock forms.

<u>Sub-aqueous tidal dune field</u> – **Outcrop area # 7** (“Little Rock City”)

- very large scale (up to 5+ m thick) planar 2-D cross-beds with fine-to-coarse sand and abundant granule/fine pebble concentrations and occasional larger pebbles; dune foresets mostly inclined 20-30^o and ~ 5-10 cm thick; granule layers usually thicker; some dunes are traceable up to 150 m across several blocks.
- foreset azimuths (50^o to 150^o) show dunes migrated parallel with and toward the paleoshore with no major reactivation surfaces; most toesets are tangential; planar truncation surface at the top of the dunes shows wave influence.
- a complete 2-3 m dune bedform (“form-set”: foreset, topset, stoss preserved); core shows directionally-opposed cross-strata which aggraded vertically until one flow direction (100^o – apparent flood tides) prevailed and the ~ 3 m dune began to migrate by periodic foreset deposition.
- $\sim$ 2 – 3 m of point bar/lateral accretion/channel deposits.
- pebbly caprock with hummock forms.

The uppermost sequence of ~ 2-3 m thick channel/lateral accretion deposits with some reddish, well-oxidized strata, plant remains and displays a HCS/wave layer in most places. The caprock varies spatially and is generally similar to the underlying deposits with some reworking evident. Within the deltaic sequence (outcrop #5), the uppermost caprock contains large (average 2-4 cm; up to 7 cm) densely/randomly-packed flat-lying vein-quartz pebbles (and some red and brown sandstone, and red mudstone rip-up clasts not seen elsewhere) with abundant aligned plant remains. The caprock shows a wave ravinement origin with an apparent transgression that brought shallow marine conditions: wave-ripple laminated buff-colored sandstones and an abundant marine fauna not seen elsewhere in the sequence.

The orthoquartzitic Salamanca conglomerate evidently records a high-energy Upper Devonian seacoast, with at least meso-tidal range, as indicated by a pebbly beach, a flood-dominated delta prograding over marine wave-rippled fine sands, very large-scale and very coarse-grained HCS, and a sub-aqueous large-scale dune field formed by strong flood tides. Most of the sequence records delta progradation and sediment transport/redistribution along shore to dunes and beaches by tides and waves and storm circulation of longshore/rip currents in the breaker/surf zone to the shoreface. Well-exposed channel deposits at the top (which overlie wave-truncated dunes and beach deposits at a similar elevation) suggest either expansion of the delta/delta plain or a transition to a coastal plain terrestrial environment (perhaps including a major flood event as suggested by localized large clasts of quartz, sandstone, mud rip-up clasts, and abundant plant fossils) followed by an apparent abrupt rise in relative sea level and a transgression as indicated by subsequent fine-grained wave-formed strata with an abundant marine fossil fauna.

Location and Physiographic Setting

The conglomerate beds of southwestern New York have long been a source of wonder. Appearing in widely-scattered and limited outcrops and more often, as isolated “float” blocks, these beds may more rarely form accumulations of large joint-separated blocks (“buildings”) and passages (“streets”) dubbed “rock cities”. Examples include “Rock City Park” south of Olean (Pennsylvanian age), “Thunder Rocks” (Mississippian? age) atop Allegany State Park, “Panama Rocks” (Upper Devonian age) and “Little Rock City” (the Upper Devonian Salamanca Conglomerate), the subject of this study and perhaps the finest example of a rock city in an unrivaled and freely-accessible setting.

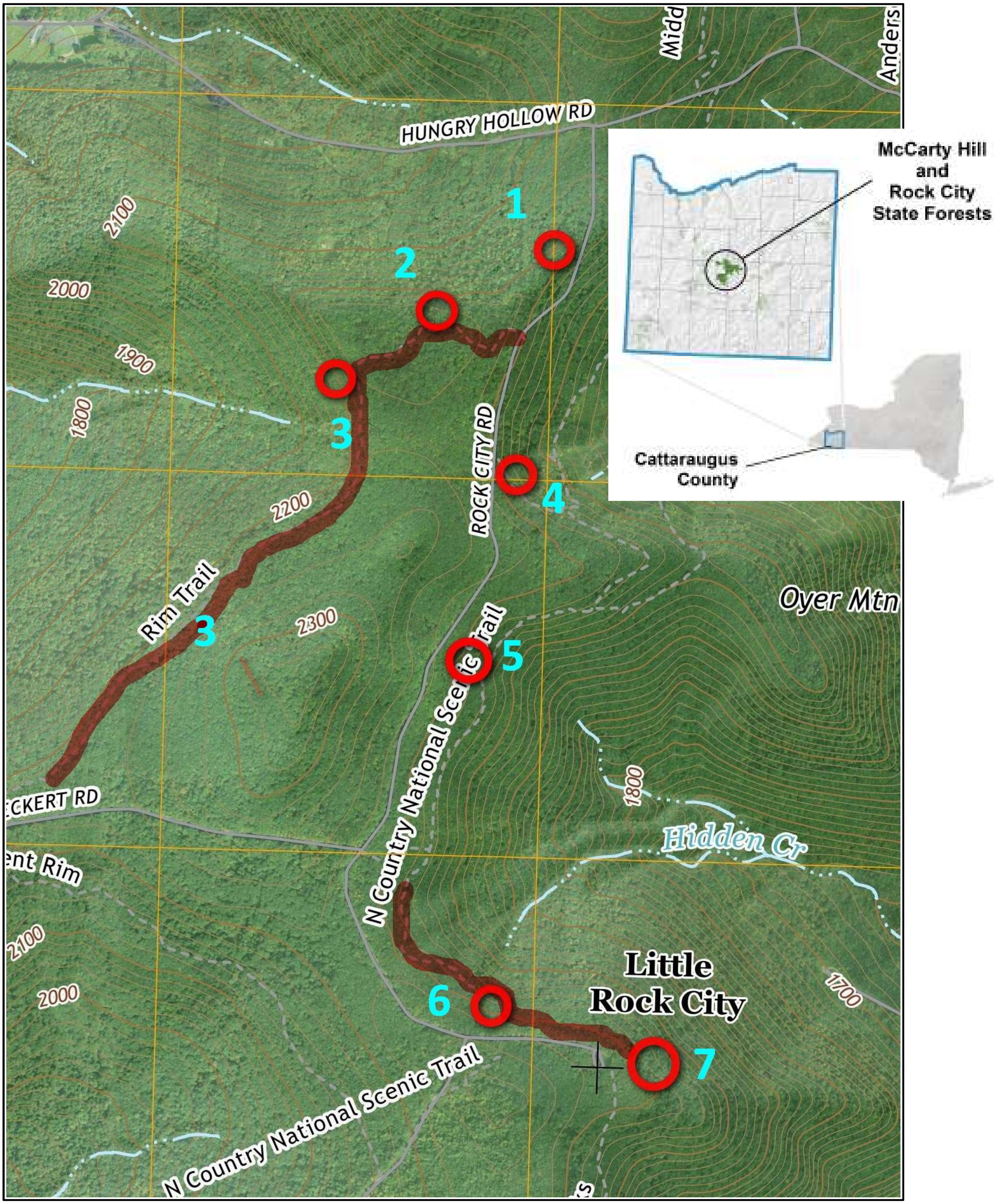


Figure 1. Location Map – Outcrops in Red (#1 - #7)

Scale: 1 cm = 200 m Source: USGS – Salamanca Quadrangle (2016)

The Salamanca Conglomerate outcrops prominently (up to a 10m escarpment) and forms a locally-widespread plateau (~ 2200 feet elevation) in Rock City State Forest and adjacent McCarty Hill State Forest (www.dec.ny.gov/lands/77184.html). This mature forest of cherry, maple, and oak blankets nearly ten square miles of Appalachian Plateau uplands between Ellicottville and Salamanca, NY. "Little Rock City" (LRC - outcrop #7) at the southeast outcrop perimeter is the type locality (Tesmer, 1975). With perhaps the most exquisite exposures, LRC has been an attraction since the early 1800s (e.g., Hall, 1843). Much of the outcrop belt is partially obscured by vegetation, rubble, and in places, glacially-deposited debris but it is readily traceable around the entire hill perimeter as facilitated by a network of hiking trails such as the North Country National Scenic Trail, the Finger Lakes Trail, and the Rim Trail. The outcrop and separated "blocks" are also readily visible with online orthoimagery (https://orthos.dhses.ny.gov/) and with one-meter DEMs (http://gis.ny.gov/elevation/lidar-coverage.htm); the outcrop belt and outcrops of interest can be traced to the Pennsylvania border.

Glaciation – Evidence and Effects

The study area is mapped within the Salamanca Re-entrant, which is part of the unglaciated Appalachian Plateau and northernmost unglaciated area in the eastern United States. Muller (1977) placed an "uncertain" glacier margin at about 1800 feet elevation at roughly one to five kilometers north of the outcrop belt. However, evidence of glaciation in the study area includes: (1) "drab" glacial till layer (chaotically-oriented thin-bedded sandstone in a gray clay matrix) exposed in a small ephemeral stream east of Eckert Road at 2200 feet AMSL; (2) a stretch of outcrops disrupted and largely buried (from outcrop #5 south to Salamanca Road that includes a topographic col/saddle which may have focused ice movement albeit in east-west directions); (3) upside-down garage-sized blocks atop the caprock at outcrop #4 as reported by Smith and Jacobi (2006); and (4) a stretch of tilted strata (point bars at the escarpment appear "pushed" with increased dip in places) at the top of the sequence at outcrop #3 (west Rim trail).

Other areas appear largely unaffected such as the isolated and well-weathered "sentinel" blocks (outcrop #1) and isolated erosional remnants (~ 3m "cubes") perched on pedestals at outcrop #4. It appears then that direct glaciation affected this area variably but periglacial effects such as permafrost, prolonged freeze-thaw cycles, and ice wedging were likely intense. Such conditions likely enhanced block separation, undermining/slump, and downslope movement due to solifluction ("soil flow"/creep due to saturated conditions) and genifluction, (creep in contact with ice/permafrost; e.g., Millar & Nelson, 2001). And the general process of soil creep continues, typically the slowest (~ mm/year on average) but geologically the most significant mass movement process (Allen, 1982).

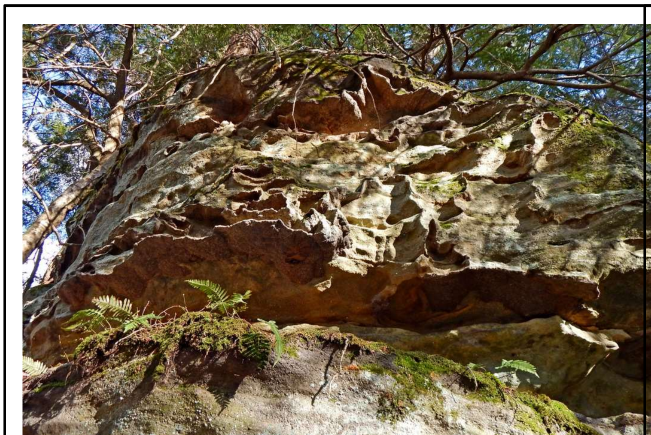
Structural Features

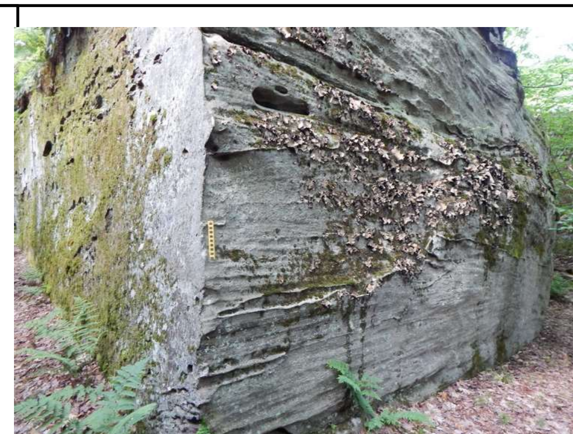
The regional dip is gently southward (about 30 feet/mile – S/SW; Glenn, 1902 and 20-50 feet/mile – South; Tesmer, 1963). No surface expression of folds or faults were observed but Glenn (1902) reported small folds in Cattaraugus County and the Clarendon-Linden fault complex is nearby in Allegany County (Smith and Jacobi, 2006). Jointing is the most obvious structural feature as it controls the similar block dimensions and the extraordinary rock exposures on the sides of the blocks. The vertical joint sets are generally orthogonal, spaced ~ 10-20 meters apart, and trend NE-SW (30^o-45^o) and NW-SE (125^o-140^o). Per Engelder (1986), the NW-oriented "cross-fold" joints are extension fractures formed by abnormal pore pressures in response to NW-directed tectonic compression during the Alleghanian Orogeny. The orthogonal strike ("release") joints are thought to develop later during regional uplift aligned with NE-oriented residual compressive fabric. The NE strike-joint set may not be as well developed and may waver more in direction and linearity as seen in the gentle sinuous patterns at outcrop #5 and along the "streets" of Little Rock City. Joints can also be affected by changes in lithology and bedding as suggested by the frequent overhangs of the upper channel deposits at the top of the blocks. Apparently these joints either did not readily propagate through the more varied (more permeable?) channel bedding in places or did so at a different spacing and/or direction. Similar effects are can be seen in shale/siltstone/sandstone sequences elsewhere (Engelder, 1986).

Iron Seams

The "iron ore" (hematite) seams of Hall (1843) are red to black in color, 1-3 cm thick, usually sub-horizontal but in places, smoothly contorted and commonly crosscut bedding. The seams appear most common higher in the sequence and in close association with fluvial/deltaic channels/redbeds and plant remains (Fig. 2 - outcrop #1). In the dune field (outcrop #7), iron seams cover several vertical joint surfaces (Fig. 3). And rare cylindrical shapes (10-20 cm in diameter) are suggestive of hollow "logs".

The joint-plane occurrences must have formed during or after joint formation. Jointing involves extension fracturing via abnormal pore pressure generated by tectonic compression (Engelder, 1986). Iron-rich porewater seems possible but the source of large amounts of reduced iron is unknown. However, the high porosity and permeability of this conglomerate could have facilitated later fluid migration. Together with hematite replacement of quartz cement in places, iron seam formation at depth after lithification, and during or after joint formation is indicated.





Iron Seams - (Figure 2 - many contorted seams in redbeds. Figure 3 - vertical iron seam on joint plane obscuring large-scale cross-bedding).

In somewhat similar but more varied occurrences in Jurassic sandstones, Chan et al. (2000) reviewed iron mobility and reactions (including Fe-reduction reactions with hydrocarbons) and proposed that mixing of fault-related saline brines with shallow, oxygenated groundwater accounted for the precipitation of iron and manganese. Given the proximity to the world's first oil fields and structural "complications", the mixing of saline brines and hydrocarbons with oxygenated groundwater as well as red bed sources may help explain the Salamanca iron seams.

<u>Previous Work and Stratigraphy</u>

James Hall provided the first scientific descriptions of these rocks ("the conglomerate") as part of the multi-year Geologic Survey of New York (1839-1843). Working in western and central NY (the 4th district), Hall's descriptions and interpretations of some sedimentary structures (e.g., "diagonal lamination" and "ripple marks") and depositional environments (e.g., Medina Sandstone beach) were among the earliest recorded in scientific literature.

Hall's (1843; p. 285-290) conglomerate description (which is difficult to improve upon other than adding "well-rounded" to pebbles) of what at the time was apparently the premier rock city (and perhaps still is) follows below:

> *"The conglomerate consists of a mixture of coarse sand and white quartz pebbles, varying from the size of a pin's head to the diameter of two inches. They are generally oblong, or a flattened egg shape. Some of these are of a rose tint when broken, but white upon the exposed surface. Pebbles of other kinds are very rare in the mass, though red and dark colored jasper are sometimes found.*
>
> *This rock in the Fourth District occurs in outliers of limited extent, capping the summits of the high hills toward the southern margin of the State…From its position, it has been much undermined; and separating into huge blocks, by vertical joints, which are often many feet apart, the places have received the name of ruined cities, Rock city, etc.*
>
> *There are several points in Cattaraugus County where the conglomerate is very well exposed upon the tops of the hills. The best known of these is the "Rock City," about seven miles south of Ellicottville* (present-day Rock City State Forest)*….The sketch* (shown above on the title page) *represents a few of the immense blocks at this place, with the passages between them. The large trees which stand upon the top, have often sent their roots down the sides, where they are sustained in the deep soil, supporting the huge growth above upon an almost barren rock.*
>
> *The masses present the same features as before described, and offer fine exhibitions of the diagonal lamination and contorted seams of iron ore. The rectangular blocks are from thirty to thirty-five feet in thickness, and standing regularly arranged along the line of outcrop, present an imposing appearance, and justify the application of the name it has received."*

The Salamanca Conglomerate is a member of the Cattaraugus Formation of the Upper Devonian (late Fammenian) Conewango Group (Tesmer, 1963, 1975). First described by Hall (1843) as a single widespread unit, "the conglomerate", Carll (1880) named the Salamanca conglomerate and proposed correlation of several similar beds. Glenn (1902) likewise correlated several conglomerate beds and traced the Wolf Creek conglomerate (a very similar cross-bedded unit of sand and discoidal pebbles overlying “Chemung” beds) and the Salamanca conglomerate from the Portville/Olean area into the Salamanca quadrangle. Clarke (in Glenn, 1902) in a very prescient interpretation, cautioned Glenn about unconformities that rings true today: *"...these sand reefs constantly display indications of deep decapitation due to shifting of bars and change of directions of currents, or a modification by heavy tidal flow on a shelving coast."* Other stratigraphic work (e.g., Caster, 1934) was summarized comprehensively by Tesmer (1975) who concluded that conglomerate correlation is difficult and uncertain due to limited and separated outcrops, glacially-derived cover, probable facies changes, and possible structural complications (e.g., slight dip changes/folding/faulting). Tesmer (1975) tentatively placed the Salamanca member in the middle of the Cattaraugus formation, following Glenn (1902) who had mapped the Salamanca member well above (~ 60-70 m) the basal Wolf Creek member in the Olean area.

Baird and Lash (1990) noted some progress with correlation of the Panama Conglomerate member with the LeBeouf Sandstone in Chatauqua County and also noted the need to locate and observe the upper and lower contacts of these conglomerate units in order to place them in geological context (this study offers glimpses). Smith and Jacobi (2006) placed the Salamanca conglomerate at the base of the Conewango Group which appears to place it between the Wolf Creek conglomerate (type section near Olean, NY) and the westernmost Panama conglomerate (type section at Panama, NY).

PALEOGEOGRAPHY - GEOLOGIC SETTING

A paleolatitude of 25-30 degrees south (Fig.4) and a warm, seasonally wet-dry climate has been posited for the Upper Devonian of New York (e.g., Woodrow et al., 1973; Scotese, 2000). Southeast trade winds likely prevailed but the Acadian highlands presented a rain shadow (Woodrow, 1985). However, abundant rainfall would be expected from postulated monsoonal circulation (Witzke, 1990, Streel et al., 2000, Smith and Jacobi, 2006), perhaps similar to the present-day Indian Ocean/Indian subcontinent or the northern coast of Australia. The Arafura Sea and northern Australia coast, with frequent tropical cyclones and a wet-dry monsoonal climate, is considered a modern analog for the epeiric Catskill sea and coast (Dott and Batten, 1980; Woodrow 1985). Climate is a primary control on source-to-basin sediment flux and in warm climates, siliciclastic flux is greatest under highly seasonal rainfall (Cecil, 1990).

Given the likelihood of monsoonal rainfall, frequent floods, episodic hurricanes (Duke, 1985; Craft and Bridge, 1987; Baird and Lash, 1990; Smith and Jacobi, 2006) with possible storm-flood (Collins et al., 2016) and storm-tide coupling, and evolving plants which paradoxically may have increased weathering rates in places (Berner, 1997), significant weathering and transport of sediment to the Catskill Sea would be expected. In addition, pulsed orogenesis in the source area (the third collisional tectophase of the Acadian Orogeny; Ettensohn, 1985) would likely have acted to increase stream/erosional gradients significant fluxes of sediment into the basin, and basin subsidence in response to tectonic/sediment loading on the crust.

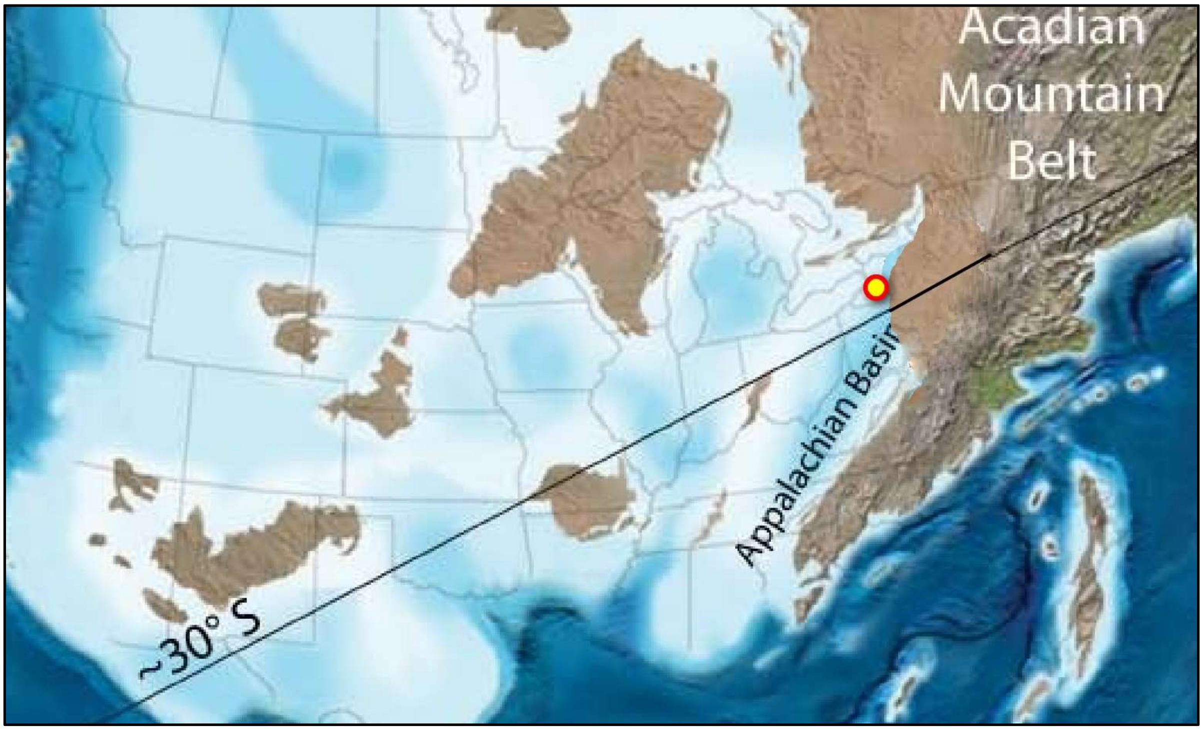


Figure 4. Late Devonian paleogeographic map – study area highlighted (modified from Blakey, 2017

and Zambito, 2011; shoreline extended into western NY; all boundaries approximate)

Basin Subsidence/Deposition Rates

The average subsidence rate in the Catskill foreland basin was nearly an order of magnitude higher in the upper Devonian than the middle Devonian (Faill, 1985) with deposits as thick as ~ 2000-3000 m over the ~ 14 million year duration of the Upper Devonian. In western New York, Faill (1985; Fig. 7) showed an estimated subsidence/deposition rate of ~ 100 m/million years (or roughly 2 million years to deposit the ~200 m Cattaraugus Formation). Coupling that rate with an average shoreline advance of ~ 30 km/million years (Dennison, 1985; Fig. 4), gives a rough volumetric deposition rate of three million cubic meters per million years per linear meter of prograding shoreline.

Shoreline

The coastal zone of the Devonian Catskill Sea varied in space and time as shown by the varied interpretations of shoreline deposits. Coastal paleoenvironments included deltas, distributary channels/mouth bars, tidal channels/flats, mud flats, and beaches (e.g., compilation in Sevon, 1985). Despite a number of early tidal interpretations, an assumption persisted that the probable tidal range/energy was low. But seminal work such as Johnson and Friedman (1969; tidal channels/flats) and Rahmanian (1979; tide-dominated delta) and tidal modeling by Slingerland (1986) and Ericksen et al. (1990) which suggested at least mesotidal (2 m - 4 m) range, recognition of tidal coastal deposits increased over time (e.g., Bridge and Droser, 1985; Bridge and Willis, 1988; Bishuk et al., 1991, 2003; Duke at al., 1991; Willis & Bridge, 1994; Prave et al., 1996).

Slingerland and Loule (1988) documented a tide-dominated shoreline (tidal channels / flats / shoals / estuaries) with a wave-dominated (sand ridges) offshore in a shore-parallel, time-equivalent (mid-Frasnian) transect through central Pennsylvania. They posited that nearshore circulation was to the SW (clockwise), estimated tidal range was high meso-tidal, and that three major clastic dispersal systems (drainage basins) existed across Pennsylvania. They also noted that meandering fluvial deposits capped all sections studied and that a lack of mouth bars and levees was attributed to strong tidal currents.

In a review of Devonian Catskill alluvial and coastal deposits, Bridge (2000) noted several coastal features in common: *"(1) sandy, tide-influenced channels; (2) shallow bays and tidal flats where mud and sand were deposited; (3) rarity of beaches; (4) storm-wave domination of the marine shelf. Much of the variability in the deposits across the area could be explained within the context of a wave- and tide-influenced deltaic coastline with a tidal range that varied in time and space."*

Regarding variations in tidal ranges, Reynaud and Dalrymple (2012) noted that since tides interact strongly with shelf and coastline morphology, changes in relative sea level can have a profound effect on tidal currents and deposits. Tidal resonance (amplitude strength) varies with shelf width (i.e., highest at increments of one-quarter of the tidal wavelength) and is directly affected by changing sea levels. They stated that: *"The increase in tidal influence can be geologically instantaneous in situations where the geomorphology changes rapidly. This was the case in the Gulf of Maine-Bay of Fundy system, which changed from microtidal to extreme macrotidal over a period on only a few thousand years."* Short-term changes then (e.g., tectonic or climate-driven sea-level variations) can bring about rapid change, *"potentially causing an alternation between tidal and non-tidal deposits"* and *"different parts of the transgressing sea can become resonant at different times"*. Also, once tidal resonance has been reached, further increases in sea level often result in a decrease in tidal influence. They suggested, as possible examples, abandoned tidal dune fields preserved beneath North Sea muds and tidal sandbodies in the Devonian Castkill Sea. They cited Ericksen et al. (1990) for the latter, who did not provide specific examples but the Salamanca tidal dune field at LRC is a possible example of decreasing tidal influence with wave-truncated dune tops overlain by channel deposits.

<u>Sediment Sources & Dispersal Systems</u>

Based on the inferred position of the Acadian orogen (Faill, 1985), source areas were likely located about 400 km to the southeast (cf. Pelletier, 1958) during Fammenian time. Weathering and erosion of actively-rising mountains produced detritus (including tabular vein quartz gravel) that was conveyed by streams to the foreland basin. As the shoreline advanced, drainage networks continually expanded and likely interacted to varying degrees. Sevon (1985) depicted up to six "sediment dispersal systems" which could have affected western NY, Slingerland and Loule (1988) noted three major drainage systems, and Boswell and Donaldson (1988) posited five stable drainage systems with large trunk streams for the Fammenian of West Virginia. The size of these drainage basins and streams are difficult to gauge but given an alluvial plain of at most 400 km, these were not the large continental rivers and deltas of today. Bridge (2000) noted that Catskill river channels were smaller near the coast (i.e., sinuous, single-channel rivers, tens of meters wide, maximum depths of 4 - 5 m, sinuosity of 1.1-1.3, mean bank-full flow velocity of 0.4 - 0.7 m/s) and perhaps distributive (delta-related). With increasing distance from the coast, slopes increased, rivers became wider (up to hundreds of meters), deeper (up to 15 m), coarser grained, and possibly braided. In this study, two large channels were identified along outcrop #3, one associated with a mouth bar complex and the other about 600 m south on the Rim Trail. Based on point bar deposits, a channel depth of 5 meters was inferred for both paleo-streams. And given their proximity and progradation over marine deposits, these paleo-streams can be considered distributaries.

Sediment - Sand & Pebbles

Other than localized rip-up clasts, no mud-sized sediment was observed. Quartz sand ranges in size from fine to very coarse, is sub-rounded to sub-angular, and composed largely of clear monocrystalline quartz. Clear quartz is mainly derived from intrusive plutonic rocks such as granite; such crystals are generally < 1 mm and are the source of most quartz sand. Cloudy polycrystalline quartz (the stuff of pebbles) predominates in coarser (1-2 mm) grains. Sand lithology is +95% quartz with occasional opaque grains including magnetite. Fine-grained magnetite comprises a very minor overall component (<<1%) of sand but it may concentrate locally along laminations in places. Bagged samples of disaggregated sand obtained from nearshore marine, beach transition, and channel deposits were magnetically-separated; all showed trace amounts of magnetite with channel deposits containing somewhat higher amounts. Since the specific gravity of magnetite (5.18 g/cm^3) is nearly double that of quartz, fine (0.125-0.25 mm) grained magnetite sand is roughly the hydraulic equivalent of medium (0.25-0.5 mm) quartz sand. At the shoreface/foreshore transition and in foresets of some dunes, dark-colored laminations and streaks occur. However, where samples could be obtained (e.g., moss-weathered outcrops), magnetite was rare. Magnetic separation showed partial black coatings on quartz grains and separated black flakes (magnetic attraction varied but mostly slight; possible hematite?).

Medium to very coarse sand dominates much of the sequence with variable percentages of pebbles; where interbedded with single or multiple layers of discoidal pebbles, the pebbles usually conform to bedding and accentuate sedimentary structures.

PEBBLES

Perhaps the most interesting geological feature of the Salamanca conglomerate (in addition to cross-bedded monoliths the size of houses) are the ubiquitous, well-rounded, discoidal, vein-quartz pebbles. The Salamanca is classified as an orthoquartitic conglomerate since the pebbles are lithologically and texturally mature. The milky polycrystalline quartz pebbles range from ~ 2 mm to 60+ mm (very fine to very coarse pebbles), average ~ 8-10 mm in size and are oblate ("flattened") ellipsoids in shape. Pebble lithologies are +98% quartz with minor amounts of red jasper and rock fragments. Shallow pits and fracture traces are evident on the surface of many, mainly larger, pebbles. Most pitting is likely related to point-contact pressure solution upon burial which probably provided much dissolved silica for this well-cemented unit. Some surface ornamentation may be impact-related such as possible percussion marks on beach clasts (Allen, 1970) and V-shaped pits. The milky/cloudy nature of the polycrystalline quartz pebbles derives from microscopic fluid inclusions which disperse light. Fluid inclusions are consistent with a hydrothermal origin where silica-rich fluids were likely emplaced under pressure and crystallized rapidly in fractures of an active orogen source zone. And sedimentary deposits of vein quartz, such as channel bars/floodplains and bedrock in the drainage basin, may also provide source material.

Pebbles often conform with and accentuate sandy stratification and hence are very helpful in defining sedimentary structures and paleoflow directions, and assessing paleohydraulics. However, in beds where pebbles dominate (e.g., minor sand matrix, clast-supported,"open framework gravels" such as present in channel fills, bars, and storm beds), stratification may be crudely developed and difficult to interpret. Pebble imbrication can be helpful such as the common orientation of oblong pebbles transverse to flow but pebble inclination may be ambiguous. Jumbled/chaotic/unstable pebble orientations suggests disequilibrium in rapidly depositing or changing flows (e.g., an "unsteady" combined flow of tides and waves).

Pebble Shape/Origin

The origin of the distinctive discoidal pebble shape has been ascribed to beach abrasion since the Salamanca was named (Carll, 1880) and accepted over time by Glenn, 1902, Tesmer, 1975, and Miller, 1974, as cited by Baird and Lash 1990). While appealing, it is not clear how an entire population (trillions?) of extremely durable quartz pebbles could be systematically and symmetrically abraded/flattened to yield co-planar sides and with a probable concomitant mass loss of 70 to 80% (assuming a quasi-spherical start and an ellipsoid finish with a "c-axis" shortened by 75%). Prolonged abrasion experiments show little mass loss for quartz pebbles after initial edge rounding (e.g., Krumbein, 1941; Keunen, 1956; Attal and Lave, 2009; Domokos, 2012). Also, at this locality the majority of Salamanca pebbles are found in channel deposits; beach deposits are not common. It seems clear then that the cloudy pebbles of vein-quartz derived their tabular shape from their origin in tabular quartz-filled fractures (veins) in the source area and rounding/smoothing during stream transport. As Pettijohn (1975) noted, the end-shape of sedimentary quartz is an expression of its initial shape.

Pebble Dimensions - Fractures/Fragmentation

Clast thickness is largely determined by the dimensions of tabular quartz veins in the source area. Caliper triaxial measurements of ~ 100 pebbles spanning the available size range yielded a C-axis (the short ellipsoidal axis) range of 1.5 mm to 16 mm which suggests veins of that size range in the source area. And that rather restricted thickness range suggests a rather consistent source area of narrow tabular veins of quartz (few "rogue" spheroidal clasts; other lithologies are less durable). And hydraulic (size) sorting and fracturing during extended fluvial transport would tend to limit clast size (the majority of pebbles are < 20 mm). Larger clasts then are sequentially sorted out; Pelletier (1958) showed an exponential decline in pebble sizes with distance from the source area in the Pocono Group. The sudden appearance of some very coarse pebbles (up to 60 mm and 25% different lithologies; sandstone, metamorphic, and mudstone clasts) at the top of the caprock suggests an unusual event or process.

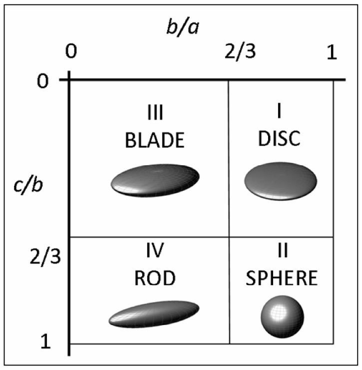


Figure 5 – Zingg diagram Figure 6 -Well-rounded quartz pebbles (~ 2mm to 60 mm)

However, smaller size ranges (< 8 mm) trend toward, and in particular, many 2-4 mm ("granules") pebbles, plot within the Sphere zone. The C-axes, while thin (1.5 mm – 3 mm), are still recognizable as parallel which suggests the same vein origin. These more equant shapes result when the A - B axes approach the C-axis in dimension.

So the C-axis is essentially fixed (vein-pebble thickness = lowest common diameter); the A and B axes can get smaller due to breakage normal to C. The suggested mechanism is the greater susceptibility of thinner veins and clasts to weathering and fragmentation at the outcrop and in early transport in high-gradient streams. Fracture traces, often outlined by iron-oxide staining, are common and are generally normal to the two co-planar sides (A & B axes). Some pebbles show smoothing/rounding of fracture-parallel edges (missing chunks) which suggests fragmentation/smoothing occurred during transport (Fig. 6). Other pebbles show sharp-edged breaks which, if natural, suggest little transport after fragmentation. So planes of weakness would tend to focus breakage along the short "C" axis and the thinner the veins/pebbles, the higher the expected rate of disintegration (lots of thin veins/pebbles = lots of milky granules and odd shapes, e.g, irregular or roughly triangular, appear more common in small pebbles). The highest fragmentation rates during transport would be expected in near-source high-gradient streams where strong flows, a wide size range of particles in motion, and high impact velocities which, along with existing planes of weakness, would promote fragmentation (e.g., Attal and Lave, 2009).

<u>Rounding</u>

Experiments have shown that lithology controls abrasion rates (e.g., Keunen, 1956, Domokos et al., 2014). Quartz pebbles are extremely durable with fairly rapid rounding (Attal and Lave, 2009; Domokos et al, 2012) but little overall change in clast diameter ("virtually indestructible", Pettijohn, 1975; Southard, 2006). In an elegant series of experiments, modeling, and field studies (Domokos et al,. 2012; Miller et al. 2014) demonstrated that *"abrasion occurs in two well-separated phases: first, pebble edges rapidly round without any change in axis dimensions* *until the shape becomes entirely convex; and second, axis dimensions are then slowly reduced while the particle remains convex." The first phase occurs mainly in high-gradient, source-proximal streams, the second, in lower-gradient alluvial plains where size sorting due to stream hydraulics and lithologically-controlled abrasion prevails."* Coupled with the fragmentation process noted above, most sizing and shaping (fragmentation) and rounding (convex shaping) of pebbles likely occurs in near-source high-gradient streams whereas most hydraulic (size) sorting and abrasion (slight for quartz) occurs in lower gradient alluvial plain streams. And considering "chipping", which contributes smaller particles to the bed-load, clast attrition by bed-load transport can occur by abrasion, chipping, and fragmentation. But clasts < 10 mm generally only abrade; chipping and fragmentation require sufficient clast mass and momentum to collide effectively. (Novák-Szabó et al., 2018; Miller and Jerolmack, 2021).

The Granule "Problem"

Pettijohn (1975) and Southard (2006) noted a general scarcity of very coarse quartz sand and granules (1 mm - 4 mm) in the rock record. The cause appears related to the fact that the most common sizes of quartz crystals in plutonic rocks, the source of most quartz sand, are mostly less than one-millimeter. The Salamanca and other Upper Devonian conglomerates have an abundance of coarse sand, granules, and fine pebbles likely produced by fragmentation and chipping of vein quartz particles during fluvial transport. These size ranges are more spherical than larger pebbles and readily transported by nearshore currents and are abundant in dunes and nearshore marine deposits. The general source of quartz grains can be roughly distinguished in field: Clear = monocrystalline plutonic sources vs. Milky = polycrystalline vein sources. The tannish granule/pebble layers ("grain striping") within the light gray sands of large dune foresets is a macro-example.

Possible Vein-Quartz Source-Area Analogues

Pettijohn (1975) and Baird and Lash (1990) noted that large vein quartz accumulations imply the destruction of large volumes of source rocks since quartz veins make up a just small percentage of normal lithosphere. However, vast amounts of vein-quartz pebbles transported within fairly limited drainage basins for tens of thousands of years suggest an unusual lithosphere (an abundance of quartz veins) within areally-limited drainage basins/source areas. Hack's (1957) law indicates that a stream about 400 km long (approximate distance from late Devonian shoreline to the Acadian orogen) would have a drainage basin of roughly 20,000 km^2. The near-source catchment width is uncertain but probably on the order of 100 km. Whereas clasts of various sizes would be expected to "banked" within an immature drainage network (e.g., at near-source alluvial fans, "wedge-top depozones", and with sorting downstream, in fluvial bars and channel deposits, and possibly older formations), a relatively small catchment area suggests a large concentration of quartz veins in a limited source area. And a much more extensive sedimentation pattern of conglomeratic, largely fluvial, deposition continued in the Mississippian/Pennsylvanian Periods. Deposited largely in Pennsylvania, the larger, more equant quartz pebbles of the Olean/Pocono/Pottsville formations suggest unroofing of vast quantities of thicker quartz veins.

Possible analogues for a vein-quartz source terrain include the Ouachita Mountains where more than 8000 meters of Paleozoic strata were folded during the Mid-Pennsylvanian Ouachita

Orogeny. Innumerable steeply-dipping fractures, related to the major folds and faults of the region, controlled the emplacement of hydrothermal quartz (Miser, 1943; Engel, 1951). And the famous Witwatersrand gold deposit in South Africa (source of 50% of the world's gold for over a century) is a Precambrian fluvial conglomerate with discoidal vein quartz (~ 30 mm) pebbles that has received much study. Another example of an abundant source area of vein-quartz as well as long distance transport of discoidal pebbles is the Miocene uplift in the southern Appalachians. As reported by Missiner and Maliva (2017), pulsed tectonism resulted in a surge in coarse siliciclastic sediment (including abundant discoidal vein-quartz pebbles of up to 40 mm in diameter) and long distance (up to 1000 km) fluvial transport.

Fossils

Plant remains are common in channel and associated deposits particularly within the deltaic sequence. With the exception of possible escape burrows in a channel base, trace and body fossils were not observed within the Salamanca conglomerate. However, abruptly overlying the caprock are finer-grained buff sandstones that contain an abundant brachiopod fauna and rich marine faunas are common in shallow marine deposits nearby. For example, an intact *Productella sp*. was found lying directly on the caprock seemingly in life position (Fig. 24).

Hall (1843) noted that fossils are extremely rare within the "conglomerate" citing 3 brachipod species in a sandy correlative of the Panama member. Tesmer (1975), citing the work of Butts (in Glenn, 1902) in the nearby Olean quadrangle, noted two brachiopod species in the Salamanca, Camarotoechia contracta and Crytosprifer sp? along with 13 pelecypod species, an ammoniod, and a gastropod.

DESCRIPTION OF LITHOFACIES

Three major categories of sedimentary structures observed in the Salamanca Conglomerate are:

- Structures formed mainly by wave (oscillatory) currents - Coarse-grained hummocky cross-stratification (HCS) and wave ripple cross-stratification, and low-angle planar beds (foreshore);
- Structures formed by unidirectional currents - 2-D and 3-D cross-stratification (2-D planar [mostly] and trough cross-beds of various scales (0.05 to 5 m) and low-angle to planar strata (bedload sheets);
- Large-scale structures (macroforms) formed by fluvial and tidal processes - Low-angle stratasets ("point" bars), channel-forms/fills, mid-channel bars, stacked/downlapped sigmoidal strata sets (components of mouth bars).

Given the diversity and overlap of these structures and processes, each will be described and interpreted separately followed by a synthesis of the depositional environments.

The overall formation geometry is tabular, ranging up to 10 m thick with major bedding planes of one to two meters thick. With a pebbly caprock of 10-50 cm in thickness, the well-cemented conglomerate of coarse sand and pebbles forms a prominent topographic bench (i.e, the Appalachian Plateau) locally and the base shows little relief where rarely exposed. Channel-forms, associated low-angle stratasets, and mouth bars outcrop prominently along the well-exposed northern and western rims (outcrops #2 - 3). comprise a thin (2-5 m) channel belt.

Wave Ripples - Rolling Grain to Vortex to Hummocks

Wave ripples may be symmetrical as formed by orbital wave oscillations or asymmetrical to varying degrees as formed by combined (oscillatory and unidirectional) flows. Larger forms in particular may require a net influx of sediment via transport by unidirectional currents. A continuum of ripple wavelengths was observed ranging from 0.05 m to > 10 m. 2-D vortex ripples are uncommon whereas 3-D forms are abundant and may be considered small-scale hummock/HCS (range up to ~ 0.5 m). Larger 3-D hummocky ripples form HCS and range from ~ 0.25 m to > 10 m. Both 2-D vortex and smaller-scale 3-D HCS transition laterally and vertically into larger HCS.

Rolling-grain ripples may form as sediment begins to move under waves. Near-bed wave oscillations set up low-profile, symmetric, sediment furrows spaced by a relation of the orbital diameter. As the orbital velocity increases, the ripples may grow vertically and develop flow separations at the crest, thus becoming vortex ripples. Rolling-grain ripples were observed at the top of some gravelly amalgamated HCS beds at outcrop #2. The very thin (~ 0.005 m), buff-colored layers that may weather in relief and help distinguish the HCS beds. The base of outcrop #6 also includes some examples.

Vortex wave ripples range from 0.05 m to 0.5 m in length and average 0.05 m to 0.1 m in height. With linear, sharp to rounded crestlines and symmetric to asymmetric trochcoidal profiles, wave-ripple cross-laminations may display discordant and/or concordant laminae with common truncations and a wavy base. Usually a thin-bedded medium sandstone with a sugary somewhat friable texture and gray-green color, amalgamation is common as are lateral transitions to HCS (e.g., outcrop #5 and the base of outcrop #6 (Fig). With increasing orbital velocities, 2-D ripples become less steep, rounded over and transform into 3-D ripples (Southard et al., 1990) and increase in size with increasing orbital velocity (Pedocchi and Garcia, 2009). With a combined flow, even a small component of unidirectional current can lead to rounded profiles, ripple asymmetry and migration (Dumas et al., 2005).

Hummocky cross-stratification (HCS) was first described by Gilbert (1899; "giant wave ripples", see highlighted section below), was "re-discovered" by Campbell (1966; "truncated wave-ripple laminae") and then formally described and named "hummocky cross-stratification" by Harms et al. (1975). The typical criteria are:

- low-profile 3-D domal bedforms ("hummocks" with wavelengths of one to a few meters by ~ 0.5 m high separated by "swales"; usually within a tabular bed);
- internally stratified by gently-dipping (<15°) convex laminae with no preferred orientation which typically thicken toward the swales and thin toward the hummock;
- composed of coarse silt to fine sand which may be form-concordant (isotropic form) and/or exhibit low-angle truncations (anisotropic form); and

- the bases may be erosional or conform to the strata below. Sole marks are common where interbedded with shale and planar laminations may occur above the base followed by HCS.

  A related type of cross-strata consisting predominately of <u>concave-upward</u> laminations ("swales") about 0.5 m - 2 m wide and a few decimeters deep is termed "swaley cross-stratification" (SCS; Leckie and Walker, 1982). It occurs mainly above HCS in coarsening-upward sequences.

Hummocky cross-strata observed within the Salamanca Conglomerate display typical hummocky forms and strata but with significant differences:
<u>Larger grain sizes</u> - Mean grain sizes of 1 mm - 2 mm (very coarse sand) are an order of magnitude greater than typical fine-grained HCS and "flat" pebbles (2 mm to > 20 mm) commonly follow stratification. Some very coarse-grained, large dimension HCS display some inverse grading with large pebbles at or near the top of the bedforms (see Fig.). One notable example is nearly all clast-supported pebbles and crudely stratified on the flanks but with chaotic fabric (even vertical clasts) at the center (Fig.)..

<u>Note:</u> Reports of coarse-grained HCS are uncommon in the literature and conglomeratic HCS, quite rare. Brenchley and Newall (1982) described HCS in coarse-grained sandstones. DeCelles and Cavazza (1992) recorded HCS of up to coarse sand size that was deposited in shallow (2-5m) depths. McClung et al. (2016) reported conglomeratic HCS in Fammenian-aged strata in West Virginia but the gravels appear in lag deposits and cross-bedding beneath the finer-grained HCS. And Jelby (2020) noted a variety of HCS types including "complex" gravelly forms from the Cretaceous of Svalbard.

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Figure 7. Typical HCS on left; Conglomeratic HCS on right

Left: A fine example of fine-grained HCS "float"ing on a hillside in Allegany State Park. Note the domal 1.5 m bedform with concordant laminations that arch gently in all directions, thin upward, and are capped by small (~10 cm; likely anorbital) wave ripples that bend/refract around the hummock (likely as the storm waned). This location is about 50 m below the Salamanca outcrop in the park.
<u>Right:</u> Very large/coarse HCS at outcrop #5 in RCSF (tape ~ 50 cm). Note low-angle strata defined by discoidal pebbles, some in excess of 2 cm in diameter. About half of the hummock is

shown extending 3.5 m in one direction. This and some other large examples exhibit poor sorting with general coarsening upward/inverse grading with some of the largest pebbles at or near the top of the bed. Bedload and suspended transport by a combined flow is indicated: intense oscillatory currents (to mobilize the sediment and mold the bedforms) and unidirectional currents to transport it.

Below: Full length view of hummock above; crest is part of the caprock (total L = 7 m; tape = 1

m)

Above: Outcrop #5 – Two views (1.5 m vertical exposure) / same channel & x-beds - Large (rip?) channel(s)/gutter(s) on left; large-scale (0.5 m) cross-beds oriented south (longshore direction) and hummocks in pebbly caprock.

Below: Very large amalgamated HCS at the top of the sequence, SW of LRC

Amalgamated HCS (outcrop #5) - Largest and coarsest at top (notebook = 20 cm); bases show both scour and drape; small-scale HCS present at base of upper HCS on left. On right, subjacent to caprock, note 3-D view of hummock at tape (= 1 m) and similar hummock to the left in same bed. Finely-laminated wave ripples (med. sandstone) just below major beddding plane.

Above: Outcrop #4 – very coarse clast-supported HCS   Outcrop #2- Granule HCS ; ruler = 18 cm.

Hummock wavelengths – Mean wavelength ≈ 2 m (see Fig. 11); based on 53 "apparent " hummock measurements in outcrop cross-sections and block corners (hence likely biased low) and commonly exceed 5 m. The uppermost stratum just below the caprock appears to contain the largest examples and are best exposed at outcrops #4 and #5 where the coarsest (up to 6 cm) caprock clasts were observed.   (Figs. 7 , 8). And at the lower range, smaller-scale forms of about => 0.25 m and includes some pebbly (clast-supported) examples (e.g., small "domes" ~ 0.30 m) beneath larger forms at outcrop #2. Also, the caprock at outcrops #2 and 7 show domal bedforms at this scale. Campbell (1966), Dott and Bourgeois (1982), and Craft and Bridge (1987) reported similar low-end ranges.

Bedding dip angles – As shown in Fig. 8 , dip angles calculated from hummock height and ½ length (arcsin H/(L/2)) for 53 hummock measurements yielded a range of 43^{o} to 7^{o} degrees with a distinct inflection point at about 20 degrees. Of the 15 hummocks with laminae dips > 20^{o}, 11 hummocks are < 0.5 m in length.   Kriesa (1981) and Craft and Bridge (1987) also reported higher dips (~ 25^{o}) for smaller-scale HCS.

Figure 8. Outcrop #5 - Big HCS - (L > 10 m; H ≈ 1 m)

Big pebbles – caprock; 18 cm ruler

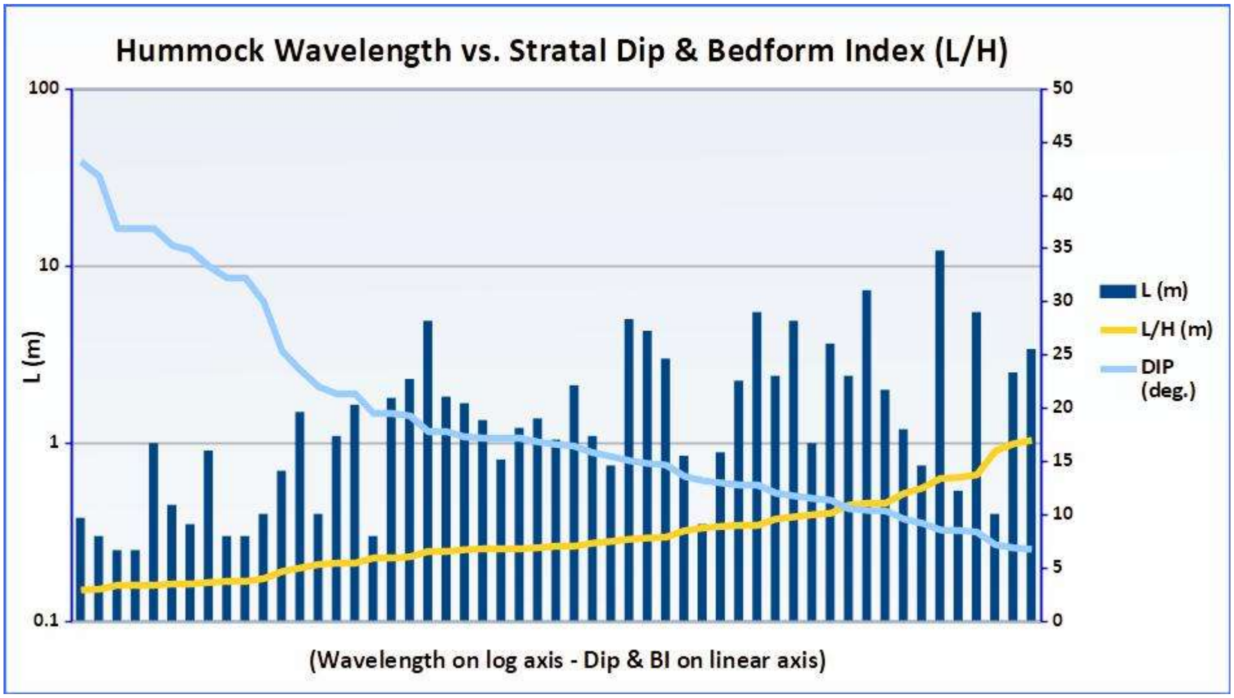


Figure 9. Smaller hummocks tend to show higher dips perhaps the consequence of a 2D-3D bedform transition and a non-equilibrium bed state. Similarly, these forms show a low bedform index (BI) due to a "taller" profile. The overall BI is < 15 and mostly < 10 as is characteristic of wave ripples (unidirectional dunes show a BI range of about 15 - 40; Leeder, 2009). Trends in these parameters (and a symmetry index) may help identify distinct combined-flow bedforms.

Unidirectional cross-beds of variable foreset inclinations (range of about 20 - 30^{o}) are interspersed in places with HCS. Most are less than 0.3 m in thickness, up to 10 m long, and directed generally southward (roughly shore-parallel). At outcrop #5, two cross-beds of about 5 to 10 meters in extent appear to grade into HCS although the exposure is poor. Amalgamation is common and on display at outcrops #2 and #5 (Fig. 7.).

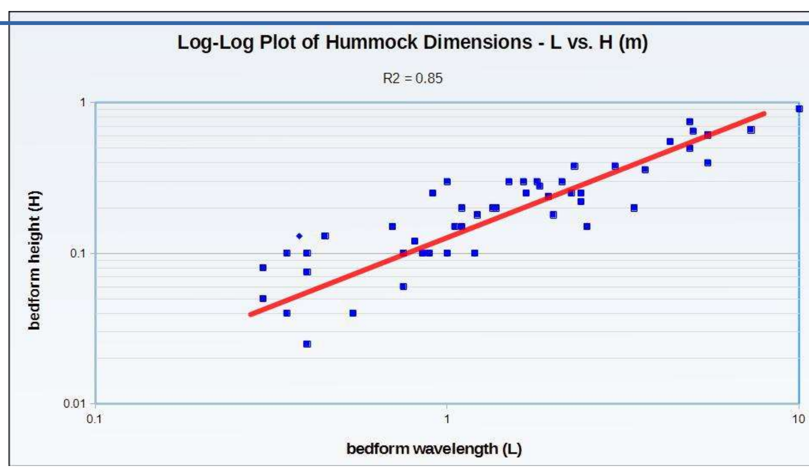


Figure 10. Plot of measured hummock dimensions shows a strong linear trend indicating similar form with size and a continuum of orbital 3-D wave ripples (hummocks). Also, pebble size, where present, typically scales with hummock size corroborating larger flow velocity/wave-forcing.

Considered a signature of storm wave deposition, HCS has been reported in numerous studies of shallow marine and a few lacustrine deposits, ancient and modern, and in laboratory flume experiments. Experiments have established the overall hydraulic conditions and sediment properties necessary to form wave ripples of various scales, in particular, HCS (a.k.a. large wave ripples). Key lab studies include Southard et al. (1990), Arnott and Southard (1990), Dumas and Arnott (2006), Cummings et al. (2009), Pedocchi and Garcia (2009), Perillo et al., (2014, and Ruessink et al. (2015).

Perillo et al. (2014) and others have demonstrated a continuum of wave ripples of increasing size under oscillatory (wave only) and combined flows with wave dominance. Given these and other results, direct field observations (Greenwood and Sherman, 1986 ; Amos et al., 1996; Keen et al., 2012), and compelling evidence of sediment transport from the rock record, a wave-dominated combined-flow origin is evident for most HCS.

Most lab work focused on fine-grained sediment since the vast majority of ancient wave deposits are fine-grained and coarser sand can present equipment limitations. In a wave tunnel study with fine and coarse sands, Cummings et al. (2009) reported large wave ripples (LWRs; wavelengths 50–350 cm, heights 10–25 cm) that varied with grain size. Fine-grained LWRs were subdued, 2-D (sharp-crested) or 3-D (smooth-crested) that resembled HCS. Coarse-grained LWRs were of 2-D form with linear crests and steep ripple faces as reported in some ancient deposits (e.g., Leckie, 1988, who posited that such 2-D ripples are the coarse-grained equivalent of HCS). In addition, they noted that all LWRs were orbital and that the wavelengths of fine-grained LWRs scaled on average to $0.6d_o$, which is close to $0.65d_o$, the most commonly reported scaling ratio for orbital ripples starting with Komar (1974). By contrast, coarse-grained LWRs scaled on average to $\mathbf{0.4d_o}$. ; similar to $0.45d_o$ reported for medium sand (Williams et al. 2004). They concluded:that their experiments did not rule out large 3-D hummocky ripples in coarse sediment since 125 cm/s, was the highest oscillatory velocity tested and a significant amount of phase space exists up to the plane bed estimate of 200 cm/s, (Clifton, 1976).

This phase-space gap was addressed by Ruessink et al.(2015) who used a large-scale (15m x 70m) wave-flume with medium to coarse sand (D_{50} = 430 µm) to examine the cross-section and planform geometry of wave-formed ripples under high-energy shoaling and plunging random waves. They determined that the ripple planform changes with the wave Reynolds number (a measure of wave forcing) from quasi <u>two-dimensional vortex ripples</u>, through <u>oval mounds with variably-oriented ripples attached</u>, to <u>strongly subdued 3-D hummocky-type features</u>. Also the ripples remained orbital for the full range of encountered conditions. By combining their data with existing coarse-grain ripple data, they developed new equilibrium predictors for ripple length, height, and steepness suitable for a wide range of wave conditions and a D_{50} larger than about 0.3 mm. Their proportionality between (L) and (*d)* is not constant, but ranges from about 0.55 for $d/D_{50} \approx 1400$ (mild waves) to about 0.27 for $d/D_{50} \approx 11{,}500$ (strong wave forcing).

<u>Paleohydraulic Reconstructions</u>

The collection of wave ripple data (e.g., wavelength, height, grain size, and crest azimuth, if present) can be useful to estimate the general wave climate, shoreline orientation, relative distance to shore, and the number and intensity of storms over a time interval. For example, Banjeree (1996) used HCS wavelength data to assess changes in wave regime and cyclicity at outcrop scale, Ito et al. (2001) used it as a climate proxy over swaths of geologic time, Yang et al. (2006) used changes in HCS wavelengths to demonstrate changes in water depth and distance from shore on a modern open-coast tidal flat, and Keen et al. (2012) used modern hurricane deposition/preservation ratios to estimate storm frequency in Cretaceous HCS strata.

And the general wave climate can be estimated; if the wavelengths of symmetric wave ripples exceeds about 75 cm, then an unrestricted water body such as an open sea can be inferred (Cummings et al., 2009).

More specifically, if limits can be placed on some hydraulic and/or sediment parameters, more precise estimates of paleo-conditions can result. In the present case with a wide range of particle sizes, possible bounds include:

- wave period (T ≈ 8 - 15 seconds) - inferring ancient hurricanes, based on paleo-latitude and given modern data and the moderate fetch of hurricanes, gives an upper bound of T ≈ 15 seconds);

- water depth (maximum ~ 10 meters) - since significant offshore transport of very coarse sediment is estimated to be limited and the shelf gradient is inferred to be low;

- wave height (from wave charts and orbital diameter estimates);

- orbital diameter (d = L/0.4); L = ripple wavelengths, field measured. For sinusoidal waves, d ≈ .wave height

- orbital velocity (derived from d ≈ L/.4 then U ≈ d (3.14)/T) - can estimate particle entrainment limits and transport potential (Fig.), and sediment transport rate (≈ cube of U), and bed shear stress (≈ square of U x drag forces).

Since hummocky bedforms are large 3-D wave ripples, ripple wavelength (L) relates directly to the near-bed orbital diameter (d) and thence to other properties such as the orbital velocity. The L/d relationship is affected by grain size and the ripples must be orbital (i.e., L scales with d; some smaller fine-grained ripples are anorbital). For coarse to medium sands, orbital diameter predictors (L/d=) include Ruessink et al. (0.27 to 0.55), Cummings et al. (0.4), Williams et al. (0.45), and Gilbert (0.5); a rough average gives **0.4 = L/d**.

Rearranging, d = L/0.4 and using HCS wavelengths (L) averaging about 1 m and ranging up to ~ 2 m for most of the sequence gives an orbital diameter (d) = 2.5 m and 5 m, respectively. Orbital velocity, $U_o = \pi d/T$; where π = 3.14 and T = 10 s (a common wave period for hurricanes) then U_o = 0.8 m/s to 1.6 m/s. These energetic wave excursions are capable of entraining pebbles of about 0.7 mm to 30 mm, respectively (Fig 10; per Komar's chart), 1974. Similar but somewhat higher values were obtained from a similar Mississippian "flat pebble" conglomerate situated about 100 m above the Salamanca conglomerate in the highlands of Allegany State Park (see); all are indicative of large waves/powerful storms.

Significantly larger hummocks (L ≈ 5 to 10+ m) are located in upper portions of outcrops #4 and #5. Maximum pebble size appears to trend with hummock size. The largest and coarsest (some with pebble-bedding and one clast-supported bed w/chaotic fabric in the domal area) are located immediately subjacent to the caprock at outcrop #5 which displays the largest clasts (up to 60+ mm) observed in this study (Fig. 7, 8).

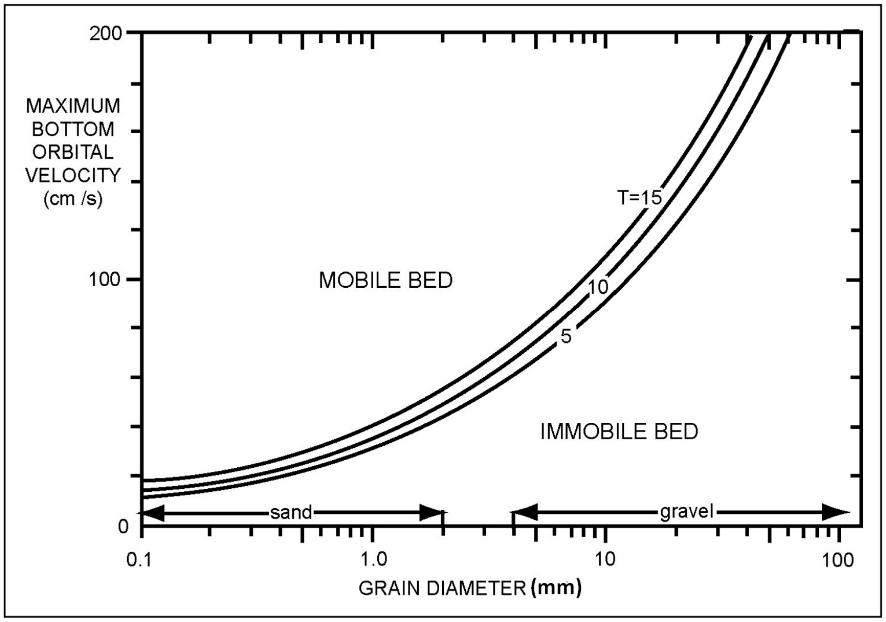
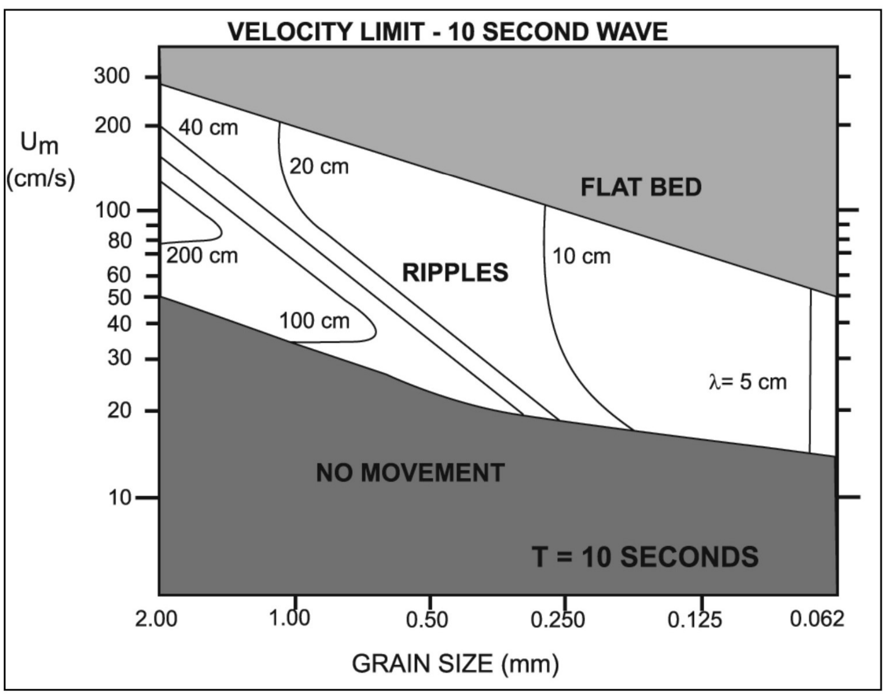


Figure 11. Wave entrainment and bedform stability diagrams

Threshold velocities for sand and gravel under oscillatory flow produced by waves with periods of 5, 10, and 15 seconds. Adapted from Komar and Miller (1973).

To assess variability in time and space and look for anomalies, measurements of 53 hummocks from across the outcrop belt were plotted on Fig. below along with orbital velocity values calculated with T= 10 and T = 15 and limits for maximum velocity and plane bed transition.



Influence of grain size on bedform under symmetrical oscillatory currents generated by a 10 second wave. Adapted from Clifton (1976).

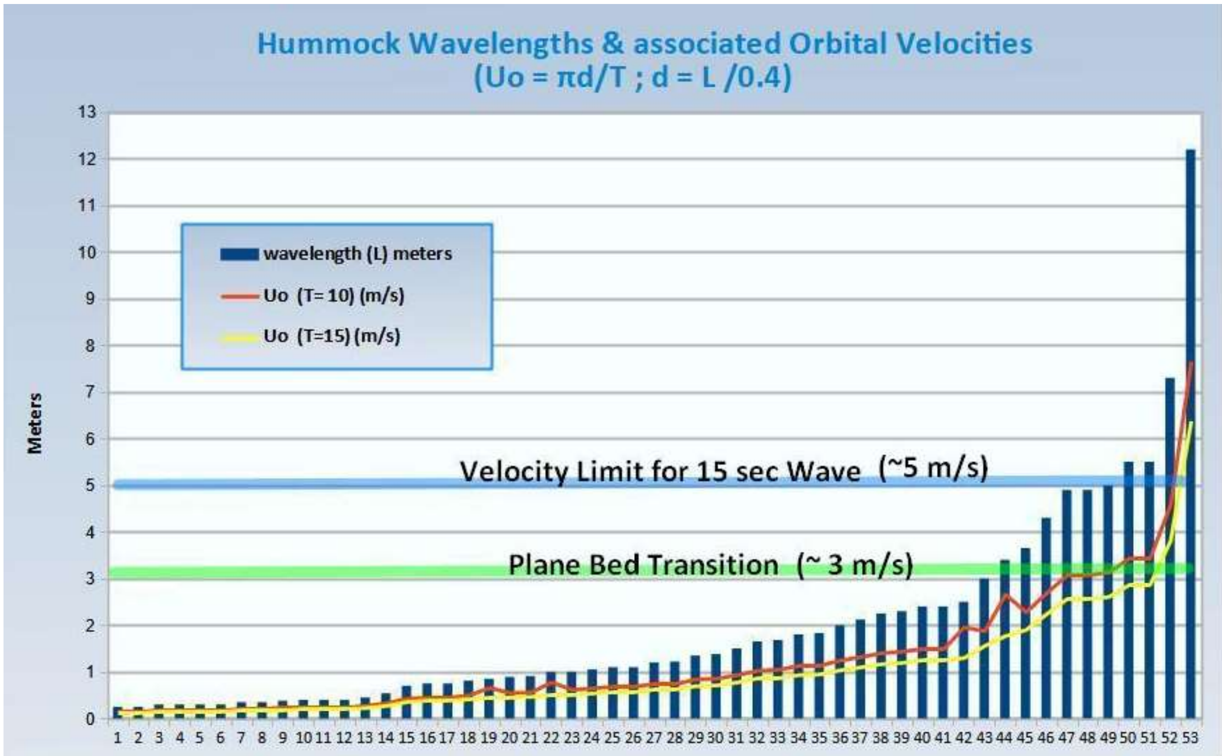


Figure 12. Measured hummock wavelengths with calculated orbital velocities

Velocity estimates for the seven largest hummocks are at or above the expected transition to sheetflow/plane beds (about U_o = 3 m/s for 2 mm sand). Since plane beds were not observed and since pebbly sediment would likely push this transition line higher (especially for more difficult to entrain "flat" pebbles), plane beds are unlikely from oscilatory currents under these conditions. As Clifton (1976) noted, sheetflow in coarse sand is unexpected except in shallow water under extreme conditions. The two largest hummocks exceed the maximum velocity "fence" (about U_o = 5 m/s and 4 m/s for waves of T = 15 and T = 10 seconds, respectively; Clifton (1976; 2003; Fig.10). Unnoticed amalgamation could be a factor or Airy wave theory may not be strictly applicable. And considering the inferred transport of large volumes of coarse sand and pebbles for most HCS and in particular, the coarsest and largest hummocks, combined flows are necessary. With higher potential velocities and bed shear stresses under waves, sediment may be entrained/mobilized and transport may then be effected primarily by unidirectional currents. Several offshore sediment transport scenarios include longshore and tidal currents directed parallel to shore and rip current cells directed offshore and downwelling from storm setup. The effect of combined flow is multiplicative rather than simply additive (Bridge & Demicco, 2008). Mike Leeder (2009; p. 421) summed it up nicely: "Storms generate rip and gradient currents that transport sediment offshore, giving rise to sharp-based, sheet-like, poorly sorted gravels and sands exhibiting hummocky cross-stratification."

Based on the paleo-hydraulic analysis of hummock formation above, plane bed/sheetflow is unlikely under oscillatory flow but antidunes are possible under unidirectional flow. Moreover,

the uniform planar bedding, subparallel with erosional set boundaries cross cutting sets at different inclinations caused by deposition under frequent changes in elevations and slope from waves and tides.
A thickness of 1 to 3 meters suggests a uniform steady process and its stratigraphic position indicates a foreshore/beach origin. Landward of the breaker zone, waves run-up the beach, the swash zone and the return flow is called backwash. Low-angle stratification dipping seaward a few degrees, is produced and generally increasing with grain size. Maximum longshore currents occur in the breaker and surf zones (meters/second). During storms, large volumes of sediment are suspended and entrained by wave/surf action. Longshore currents may readily transport it and with offshore-directed rip currents deliver coarse sediment offshore.

<u>Note:</u> Perillo et al. (2014) provided a predictor of L/d = 0.82 for <u>combined flows</u> for 0.25 mm grain size and Dumas et al. (2005) provided 0.5 for combined flows in fine sands. Since the ripples in their studies remained orbital and the unidirectional component was << than the oscillatory component, the more common oscillatory predictors, with a greater range of grain sizes (the key variable), are preferable.

<u>Equilibrium Times</u>

A surprising result of the unified model of bedform development and equilibrium (Perillo et al., 2014) is that large bedforms reach equlibrium much faster than smaller bedforms apparently due to higher sediment transport flux (and higher associated current velocities). As their flume study showed, large bedforms, such as hummocks required much less time (only a few tens of minutes) to reach equilibrium conditions than small wave ripples which generally required several hours. And they presented four stages of bedform genesis and growth; the second stage (“growing bedforms”) showed exponential growth by sediment capture and amalgamation. Large hummocks then may develop quite rapidly and the time represented by individual HCS beds may be measured in hours or less.

The unified model of bedform development and equilibrium under oscillatory, unidirectional, and combined flows (Perillo et al., 2014) that showed bedform equilibium times for all flow types were inversely proportional to the sediment transport flux. Since the flow conditions required for large wave ripples mobilizes significantly more sediment than small ripples, large ripples reached equilibrium faster. This result illustrates that the sediment transport dynamics play a key role in controlling the equilibrium time of bedforms. The experimental results show that the processes of bedform genesis and growth are common to all types of flows independent of bedform size, bedform shape, bedform planform geometry, flow velocities, and sediment grain size. Four stages of bedform genesis and growth: (i) incipient bedforms (e.g., rolling grain ripples); (ii) growing bedforms (e.g., hummocks; exponential growth by sediment capture and amalgamation); (iii) stabilizing bedforms; and (iv) fully developed bedforms (dynamic equlibrium; wavelength and height fluctuate bedforms may merge, split, and produce a relatively large range of active sizes). Based on these experimental results, it was observed that the genesis and growth processes are common for all types of flow.
This model may help explain the presence and growth of small-scale HCS at bases of some larger hummocks (such as outcrop #2 and #5; (Fig. 7). The small forms may be considered “incipent bedforms” which then grew exponentially into large hummocks as flow conditions

allowed, stabilized and in dynamic equilibrium, may form an array of sizes. Arnott & Southard (1990) noted a similar effect where large 3D wave ripples, which grew in size to 1-3 m at the highest attainable U_m of 1 m/s, shifted in position and changed in size with time, seemingly at random, causing substantial local and temporary erosion and deposition at a given point on the bed.

Effects of Pebble Shape

By some measures, such as a simple fluid/particle force balance and frictional considerations, low-profile discoidal pebbles should be more difficult to mobilize Experiments by Komar and Li (1986) showed that pivoting angles are key variables and listed spheres as most readily entrained followed by smooth ellipsoids; angular clasts were last. With smooth discoidal clasts, "rolling" like a sphere is impeded but the typical biconvex oblate ellipsoid shape may be more readily "lifted" and entrained in a current. Once entrained ("mobilized") in a current, the discs would likely settle slower as indicated by calculation of the Maximum Projection Sphericity (Sneed and Folk, 1958). Also known as Maximum Settling Sphericity, a range of pebble sizes yielded about 0.5 (equivalent to about twice the cross-sectional area of a sphere of equivalent volume) which suggests slower settling of discoidal pebbles.

Theory and experiments on the effects of grain shape on settling rates confirm such reductions in settling velocities for discoidal vs. spherical pebbles. For example, Komar and Riemers (1978), Dietrich (1982) and Wu and Wang (2006) showed about a 50 - 60% reduction in settling velocities for one-centimeter discoidal pebbles (~ 30 cm/sec) vs. spherical pebbles (~ 60 cm/sec to 70 cm/sec for) due to larger form/shape drag for discs. Somewhat smaller reductions were noted for smaller particles (e.g, ~ 40% for very fine sand; 0.51 cm/sec vs. 0.82 cm/sec). Overall, reductions in settling velocities would be expected to increase pebble suspension and transport with the potential for shape sorting and association with smaller but hydrodynamically similar particles. With observations of large isotropic HCS with pebble-bedding (Fig. 7; end of spectrum) and high U_o predicted for large hummocks (Fig.12), the suspended load of sediment may include pebbles (as well as saltation). With flow deceleration, settling from suspension may be a major part of the formative process as with fine-grained HCS. And once deposited, the low-profile "flat" pebbles, being more difficult to entrain, may "armor" the hummock and with highly variable storm currents, larger pebbles may then be deposited (localized inverse sorting).

So pebble shape and mass (about ¼ the mass of a spherical clast of the same diameter) may be a primary determinant of pebbly HCS and perhaps help explain its apparent rarity in the geologic record. Given that typical shallow-marine strata with typical fine-grained HCS sandwiches the Salamanca, a rapid change in wave climate is unlikely to account for conglomeratic HCS is unlikely. Rather, under intense storm waves, fine sand may stay suspended or develop sheetflow until conditions allow deposition as HCS. With the availability of coarse sand and discoidal pebbles (which would not likely develop sheetflow but orbital velocity limits may be approached as noted above) may allow formation of HCS in the phase space of near sheetflow conditions under full-storm conditions. Sediment shape, size, and mass may help explain large-scale coarse HCS, the largest of which is at the top of formation where inferred shallower water depths may also have been a factor.

Whereas the hummocks appear fairly symmetrical, the evidence of unidirectional currents is abundant (angle of repose cross-beds, high sediment transport and depositional volumes, large clast transport (ideally one would expect some asymmetry but outcrop orientations are variable).

In any case, given the overall range of coarse hummocky forms, significant sediment transport is evident in these HCS layers (which varies with the cube of velocity) along with deposition (by combined flows decelerating spatially and temporally) in definitive patterns (variably-sized hummocks with similar bedform indices) in a relatively brief periods of time (probably hours to days). And, this exercise highlighted some unusual forms that suggest unusual depositional conditions in this area just below the caprock (a probable "maximum regressive surface" or "transgressive surface" depending on one's view).

Giant Ripples of the Medina

Gilbert (1899) was apparently the first to describe the type of cross-bedding now known as hummocky cross-stratification (HCS) and moreover, the first to deduce and reconstruct possible wave conditions for ancient deposits. As he observed in Medina sandstone quarries and outcrops near Lockport, NY, the cross-bedding is "*a peculiarly intricate type, exhibiting dips toward all points of the compass in the same quarry, and associated with it are many unconformities*" (random cross-bedding/truncations which made these beds unsuitable for building stone). In a quarry, the strike and dip of a layer *"can be traced through an elliptic arc, like the end of a spoon for 150 degrees"*. *"In width* (wavelength), *these giant ripples range from 10 to 30 feet; in height, from 6 inches to 3 feet. Their material is a sand of medium grain."*

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FIGURE 1.—GIANT SAND-RIPPLE

Occurring in upper sandstone lens of the Medina formation at Lockport, New York, exposed in old quarry west of the Indurated Fiber Works

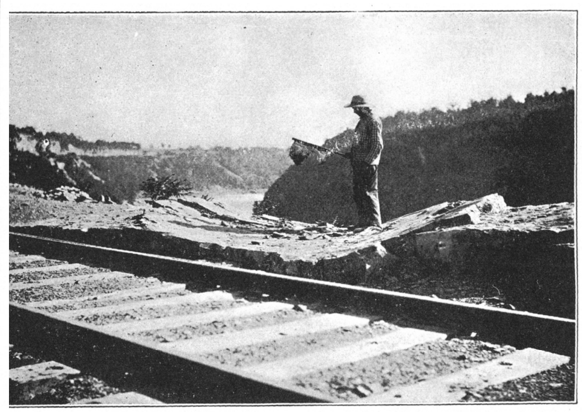

FIGURE 2.—GIANT SAND-RIPPLE

curring in "quartzose sandrock" of the Medina formation, exposed in the Niagara gorge, one mile above its mouth. New York Central railroad track in the foreground

GIANT SAND-RIPPLES

(<u>Historical fotonote:</u> The 1888 invention of roll film and affordable/portable cameras in Gilbert's hometown must have transformed fieldwork just a few decades after the classic 1843 Geologic Survey of New York relied on exquisite but tedious hand-drawn illustrations. A century later, Kodak invented digital photography, again transforming the landscape and driving innovation (and personally, filling hard drives, bereft of Kodachrome). And this centennial tribute to Gilbert's myriad geologic contributions is illuminating: https://eos.org/features/reflections-on-the-legacy-of-grove-karl-gilbert-1843-1918).

Bedding plane exposures of the giant ripple bedforms were/are uncommon as with most bedrock outcrops. But building stone quarries were common before the advent of concrete and the quarry floors presented early geologists with the rare plan view/third dimension as might be otherwise seen only in shallow rock-floored streambeds. The example depicted in the left photo above includes two crests and an intervening trough; the trough is 23 feet across and 29 inches deep. A partial exposure shown in the right photo shows a trough fragment 15 feet across and 16 inches in depth. And another quarry exposure reportedly showed a bedform with a convex crest 15 feet across. The associated cross-bedding (n.k.a., HCS) however, is common in vertical outcrop exposures of Medina sandstone between Lockport and the Niagara gorge and elsewhere (from which measurements of "apparent" wavelength can also be attained).

In a series of deductions based on nascent wave-ripple experiments (e.g., Darwin, 1883/84; son of Charles; "ink mushrooms/trees" discussed at length in Allen, 1982) of the day and his own observations, Gilbert surmised that these bedforms and cross-bedding formed by *"sand rippling, differing in no respect except size from the familiar ripple-mark of the bathing beach…the cross-bedding is a result of deposition during the maintenance of a rippled surface on the ocean bed, and the unconformities record the readjustment of the sand ripple pattern when the controlling water movement assumed a new direction."* He further noted that the orbital motion induced by waves becomes elliptical near bed and *"the frequency of the natural oscillation equals the frequency of the wind waves, and its amplitude is a function of the size of the waves and the depth of the water...so that a relation will ultimately be established between wave-size, wave-period, and water depth as conditions and ripple-size as a result."* He presciently summed up the crux of the wave/ripple-mark conundrum. Gilbert concluded that at the most, ripple-mark wavelengths are about half of the wave height. And he humbly noted that if this relationship is substantiated by future research, "*the geologist may infer from the structure of the Medina sandstones that the Medina ocean was agitated by storm waves sixty feet high. As great waves require broad and deep bodies of water for their generation, such a result would demonstrate the association of the Medina formation with a large ocean.*"

Some six score years later and a tsunami of wave research, unique solutions for ripple size from wave size, period, and water depth are still lacking but useful estimates can be had with paleo-depth inferences and wave and sediment constraints. Most modern predictors of ripple wavelength (L) vs. wave orbital diameter (d) ≈ sinusoidal wave height) use a relation which ranges from about **L/d ≈ 0.3** to **0.8**. Gilbert's prescient predictor of **L/d ≈ 0.5** appears "substantiated" and we may infer that the sedimentary structures he recorded (ripple wavelengths of 10 to 30 feet) were formed by storm waves on the order of 20 to 60 feet high in an unrestricted ocean of some depth. Such wave heights are common in large storms according to hurricane data collected over the last few decades (an eye blink, geologically). For example, significant (highest 1/3) wave heights range roughly from 10 to 20 meters for Category 3 to 5 hurricanes. Ocean buoy data collected during Hurricane Katrina showed significant wave heights of ~ 17 m (55 feet); statistically, the highest waves could have been as high as ~ 32 m (105 feet). Similarly, wave periods are elevated for large hurricanes averaging at least T = 10 seconds within a 100 km radius of the eye and T ≈ 15 seconds near the eye (https://sos.noaa.gov/catalog/datasets/wave-heights-hurricane-katrina-2005/).

Description of Unidirectional Current Structures - Subaqueous Dunes and Bedload Sheets

Unidirectional currents (e.g., streams, tides, longshore/rip cuurents) are ubiquitous in nearshore environments. With bedform formation and migration, cross-stratification of various styles and sizes result. Dunes scale with flow depth (pint to building size here) and bedload sheets are low profile (few grains high) bedforms which transport mainly pebbles. Other common bedforms such as ripples and upper-stage plane beds occur in finer-grained sediment (~ < 0.6 mm) that is uncommon here. Likewise, upper-flow stage antidunes are possible (with high velocities and shallow depths) but with pebbles, transverse ribs would be expected and were not observed.

Large-scale Cross-stratification

Cross-stratification is present in most and dominates some outcrops. Individual set dimensions range over two orders of magnitude in scale (~ 0.10 m to +5 m). Most cross-strata are planar (i.e., straight- to slightly- crested = 2D type) with some trough (i.e., sinuously crested = 3D type = higher flow stage) evident in channel deposits and upper shoreface/lower foreshore deposits.

Foresets are largely composed of grayish medium to coarse sand with variable interbeds of milky granules and pebbles ("grain striping" within large-scale foresets). Larger sets are generally coarser. Some channel bars and fills are composed in large part with pebbles (open framework gravel) that are crudely stratified or imbricated.

The vast majority of paleocurrent data are cross-bed inclinations and range from 90^o– 150^o (mostly) with minor clusters at 40^o – 60^o ; 170^o – 190^o (coast parallel); 220^o – 240^o.

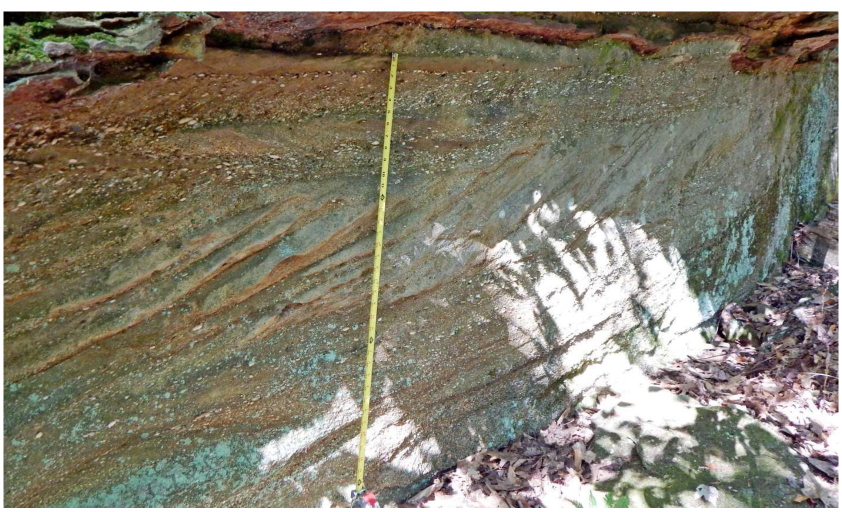

Figure 13. Bi-directional cross-bedding at top; tidal bar ; tape = 1 m – outcrop #7.

The largest cross-strata (0.50 – 5 m) increase in size and abundance from north to south across the outcrop belt. At the southernmost outcrops at "Little Rock City", large foresets may comprise ~ 75% of the outcrop exposures with dips of 20^o – 30^o and no obvious or major reactivation surfaces. Most toesets are tangential. Some foresets are traceable for +150 m across several blocks and a planar truncation surface at the top of the dunes shows wave influence (e.g., wave ripples with crests parallel to the paleo-shore and HCS beds above).

Figure 14 . Largest dune (+ 5 m) observed at LRC - outcrop # 7.

Interpretation of Cross-Strata

Hall's (1843) explanation of "diagonal lamination": *"...where the sand is carried on and spread over the surface, sloping off towards one side farthest from its origin. The next deposition covers this sloping side necessarily in the same manner, producing the oblique lines..."* was the first documented account of cross-stratification (Allen, 1982); it describes the essential process of sand movement and deposition on an inclined surface. To embellish slightly, currents transport sediment along a gentle stoss slope to the bedform crest where repeated sediment avalanches down the steeper lee slope form cross-strata at or near the angle of repose. The resulting cross-strata are the depositional units formed by the migration of bedforms, dunes of various scales in this case.

Based on mostly shoreward- and some bi-directional-oriented cross-strata, the dominant currents were tidal and predominantly flood tides. Most sediment transport likely occurred during high spring tides of the bi-monthly spring-neap tidal cycle. Bedload transport rates scale roughly with the cube of the current velocity; if the flow rate doubles, bedload transport increases by a factor of roughly eight (Wang, 2012). A mesotidal range of 4 m appears to be a reasonable estimate; similar modern deposits/bedforms are produced in that range such as in the North Sea.

Paleohydraulic Estimates for Unidirectional Currents

As discussed above, discoidal pebbles are likely more difficult to entrain but also settle about 50% slower. And their mass is about 60% -75% less than a sphere of the same diameter. Overall, reductions in settling velocities would be expected to increase pebble suspension and transport with the potential for shape sorting and association with smaller but hydrodynamically similar particles.

Bradley et al. (1972) studied the effect of shape both in the field (Knik River, Alaska; high-gradient glacial-meltwater stream) and in the laboratory. They detected downstream sorting of shapes, with platy pebbles being the most easily transported, then elongate pebbles (rollers), and more equant pebbles being the least easily transported. The different shape-sorting effects were attributed to particles moving by traction and by suspension and hence closely related to flow strength and particle size.

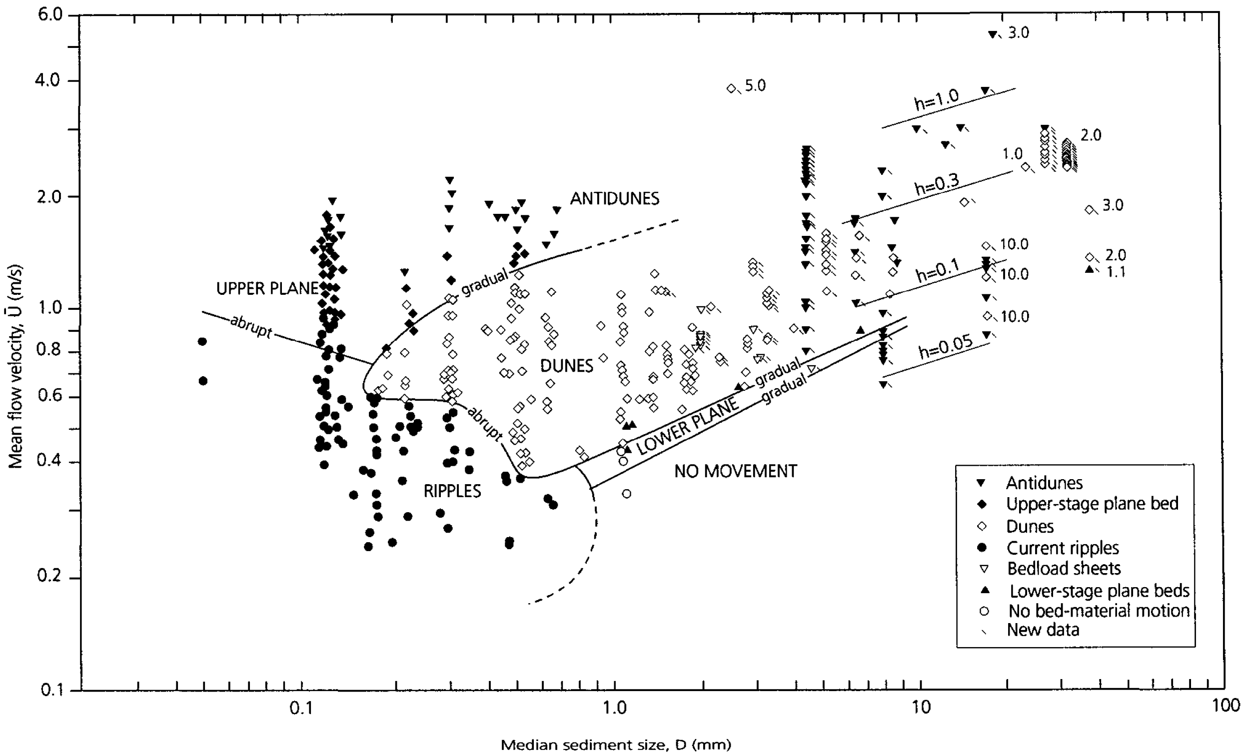

Figure 15 - Bedform Existence Fields for Unidirectional Flows (Carling, 1999; redrawn after Southard and Boguchwal, 1990 and extended to gravel sizes, D ~ 33 mm). For general background, see Middleton (1977), Allen (1982), and/or Harms et al. (1982).

The bedforms most applicable to conditions at LRC are <u>Dunes</u> and <u>Bedload Sheets</u>. Dunes of various dimensions produce cross-strata which incline downcurrent and scale with depth, flow strength, and grain size. Ripples, smaller-scale (<= 4 cm) dune-like bedforms, require sand of < ~0.6 mm which is uncommon at LRC; ripples have not been observed. For sand sizes, a current of roughly 0.4 m/s to 1.0 m/s would be required to form dunes. For dunes composed of granules and fine pebbles (2 mm to 8 mm), a current of 0.6 to 1.5 m/s is indicated. Flows required for coarser pebbles (8 to 32 mm; limit of graph) are less clear given prominent disc shapes and data scarcity, but currents of 1 m/s to +2 m/s appear likely. Bedload sheets, low-amplitude bedforms which can transport a range of sediment are likely common at LRC but difficult to definitively identify; their existence field appears to coincide with dunes.

Based on paleohydaulic estimates noted above, the currents required to form dunes ranged from about 0.5 to 1.5 m/s. A similar velocity range (0.5 - 1.0 m/s) has been reported in the Dutch North Sea where very large simple dunes (like those at LRC) are actively migrating decimeters to a few meters per year (e.g., Tonnon et al., 2007; Passchier and Kleinhans, 2005; Stride, 1982). Allen (1982) reported that on the European continental shelf, sandwaves (large dunes) are found where tidal currents associated with spring tides range between 0.65 and 1.30 m/s. LRC dunes are somewhat coarser than modern examples and perhaps formed in somewhat shallower depths. A shoreface depth of ~ 10 m would conform with the dune height/depth ratio of 0.5 (Allen, 1982) for the largest (~ 5m) LRC dune (note that lower *h/d* ratios are the norm; Reynaud and Dalrymple suggest ~ 0.2). But it is unclear how smaller superimposed dunes (sediment "caravans" that migrate up the stoss slope of the large dunes delivering sediment to the avalanche face) affect this estimate; these would scale to a much shallower depth.

Tidal transport of coarse sand and pebbles at much greater depths may have been limited by the "littoral energy fence" whereby coarse particles are sequestered nearshore (Allen, 1970; Thorne and Swift, 1989). Even sand is rarely transported offshore by fair weather processes but evidence for Devonian hurricanes in the Catskill basin is strong and modern studies of sediment transport inform the past (e.g., Keen et al., 2012).

Figure 16. Large dunes at Little Rock City, outcrop #7. This series of blocks form one dune > 125 m long with foresets ~ 4 m high. Considering the very consistent flow direction and steady sediment transport rate (reflected in fairly uniform foresets) and formation over decades to centuries implicate tidal currents.

The large foresets on large "simple" dunes suggest strong very asymmetric tides. Allen (1980, 1982) depicted four general variants of "sandwaves" (what geologists now call "dunes" per Ashley, 1990) based on tidal current symmetry. Allen's conceptual "sandwave"/dune generated by the most asymmetric tides (Fig. 16; large simple foresets in bottom frame; note the velocity asymmetry of U* critical, the threshold velocity to move sediment) conforms with the large dunes at LRC. Allen (1982) also depicted superimposed smaller dunes supplying sediment to the large foresets of a larger host dune.

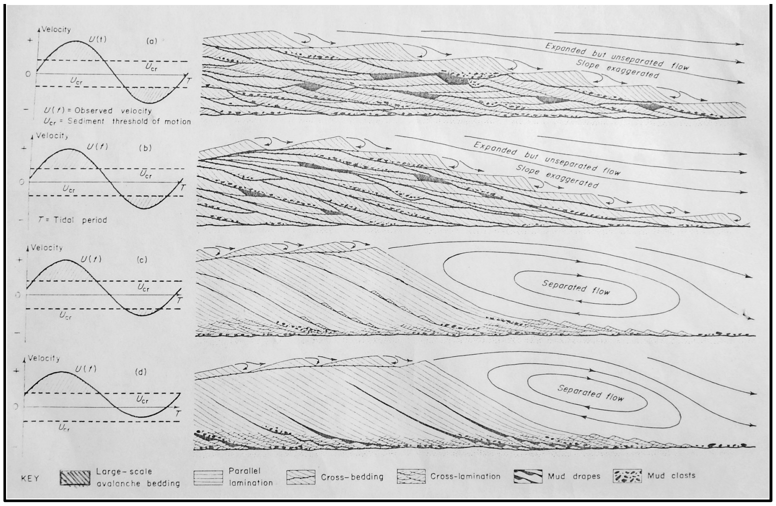

Figure 17. Conceptual Model for generation of large dunes (Allen, 1982)

The formset dune at the LRC dune field contains a small dune that appears to have "stalled" at the crest and reformed/"sharpened" it (center of Fig. ; above 18 cm ruler); deposition continued along the aligned stoss and lee of both dunes (cf. Bridge and Demicco, 2008). Superimposed dunes pre-sort and transport sediment to and over large host dunes often in concert with bedload sheets. Pre-sorted wedges of sediment, as formed by smaller dunes ("trains") advancing over the crest ("cliff") of large host dunes, move down the flow-separated lee slopes often haltingly, by "grainflow" and finer sediment is distributed by "grainfall" from suspension (Reesink and Bridge, 2007; 2009).

One consequence is the scattered formation of lobate tongues of coarse sediment that may show sorting and inverse grading due to kinetic sieving. Otherwise known as "grain striping" (Reynaud and Dalrymple, 2012), a likely result of this process is shown in the dune cross-section along depositional strike (dune "backside") with scattered and variable convex piles of granules and pebbles along the foresets.

Harms et al. (1982) described the process of foreset avalanches at high sediment concentrations as oversteepened areas which slump in places and slide down the lee slope as long "tongues". A slight scour or channelized grainflow may form which then "debouches" with a slight positive lobe at the basal portion of the foreset. Some foresets and "grain stripes" at LRC nicely display these subtle structures in dip cross-section (Fig.10) and the small granule "piles" on a dune backside noted above are interpreted as "grainflows" along the strike of dune foresets (Fig.18, below).

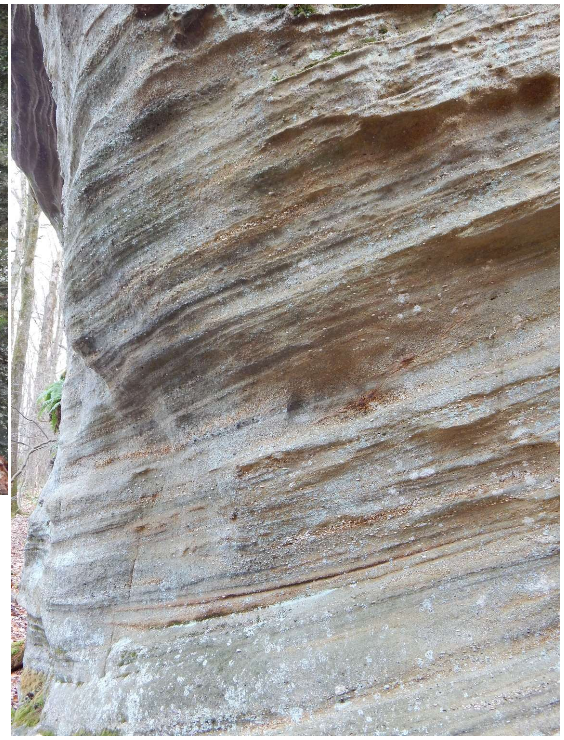

Figure 18. Above: Humps of coarse sediment (grainflows from subordinate dunes) along depositional strike of the dune. Some humps show inverse grading likely from kinetic sieving during foreset avalanching. Right, large dune displays “grain striping” produced by grainflow of coarser sediment (tan granules in this case) down the foresets.

The largest cross-strata (0.50 – 5 m) increase in size and abundance from north to south across the outcrop belt which may indicate increasing water depth since dunes scale with flow depth. And the LRC dune field is downcurrent (longshore, tidal, and possibly storm setup) of the inferred delta complex which provided an abundant sediment supply and may partially explain the location of the dune field.

Figure 19. Form-set Dune

The form-set dune at the LRC dune field contains a small dune that appears to have "stalled" at the crest and reformed/"sharpened" above 18 cm ruler; deposition continued along the aligned stoss and lee of both dunes (cf. Bridge and Demicco, 2008).

Superimposed dunes pre-sort and transport sediment to and over large host dunes often in concert with bedload sheets. Pre-sorted wedges of sediment, as formed by smaller dunes ("trains") advancing over the crest ("cliff") of large host dunes, move down the flow-separated lee slopes often haltingly, by "grainflow" and finer sediment is distributed by "grainfall" from suspension (Reesink and Bridge, 2007; 2009).

In this case, it appears that a medium ebb-dune formed at the base and small competing dunes aggraded vertically (or slightly in the ebb direction) until the flood (shoreward) tides began to dominate about where the small dune is perched in the middle of the bed. With an abundant up-current sediment supply, the flood tidal current began to dominate and the ~ 2–3 m dune began to migrate. In effect, the simple large dune has compound small dunes at its core/start and other superimposed dunes supplying and presorting sediment probably along with bedload sheets. In addition to a small dune in the topset bed, at least 2 topset locations show continuous strata between inferred bedload sheets and foreset beds. In a review of dune preservation, Reesink et al., (2015) noted that dune sets may climb due to local dominance of deposition over dune migration which generally fits this situation. But more specifically in this case, it appears that the localized balance between ebb and flood dune deposition aggraded a vertical core until the more dominant flood current and sediment supply tipped the balance toward large dune migration.

Wave-truncated Dunes

All of the largest dune (> 3 m) foresets at LRC appear to have been truncated ("beheaded" dunes) horizontally at similar elevations (~ 3 – 4 m from the top of the sequence). Evidence of HCS (see below) and wave ripples at this interface is common which suggests storm wave action, as well-documented in the North Sea (e.g.,Terwindt,1971; Reynaud and Dalrymple, 2012).

Figure 20. About one meter of uppermost dune is exposed and appears truncated, followed by 0.75 m of HCS and (on the left) about 1 m of low-angle strata which is truncated and followed by HCS red beds and capped by a 0.5 m pebble-rich stratum (caprock). Two storm wave truncation events are interpreted.

Channel Deposits

Figure 21. Large channel complex at outcrop #6. Marine strata w/HCS at base, large channel-form above. Multiple channelss, some incision (tape = 1 m); deltaic depositional environment inferred.

Channel deposits comprise upper portions (uppermost ~ 1 - 3 m and up to ~ 5 m at outcrops #2, 3 & 6) Low-angle stratasets and associated channel-forms/fills dominate these deposits.

Low-angle stratasets (a.k.a. “lateral accretion deposits”; LADs):

- dips ≈ 10 -20^{o} (up to 25^{o});
- are typically 2- 4 m thick, measured normal to bedding;
- extend laterally tens or up to hundreds of meters; and
- may be organized/stacked vertically into several “storey” units.

Individual strata/beds of low-angle stratasets:

- are centimeters (cm) to decimeters (dm) thick;
- vary in cross-sectional shape from co-planar (mostly) to gently concave-upward (as beds may thicken along dip) and rarely convex-upward; and
- display planar (normally-graded) stratification and cross-stratification with some alternating bi-directional foresets in places) and cryptic or massive bedding where pebbles dominate.

Figure 22. Three low-angle stratasets (identifible by color and bedding style) stacked laterally over /
horizo.ows two small wave ripples at top; the center set dips ~0° (local flow ≈ E-W and contains pebbly cross-strata directed southward; on the left, gray thinner-bedded set incised the center set and dips ~45° (local flow ≈ NW-SE) and the bedding shows arched weathering (possible wave influence); and chilling human standing on mounded HCS caprock.

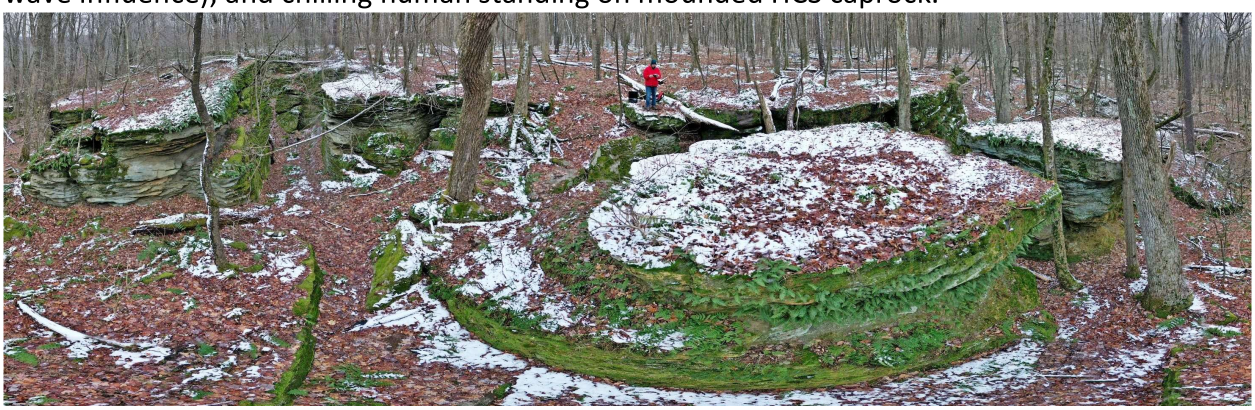

Figure 23. . Outcrop #3 – Apparent preservation of a curved channel-form and adjacent curved low-angle strataset (an intact point bar/channel deposit).

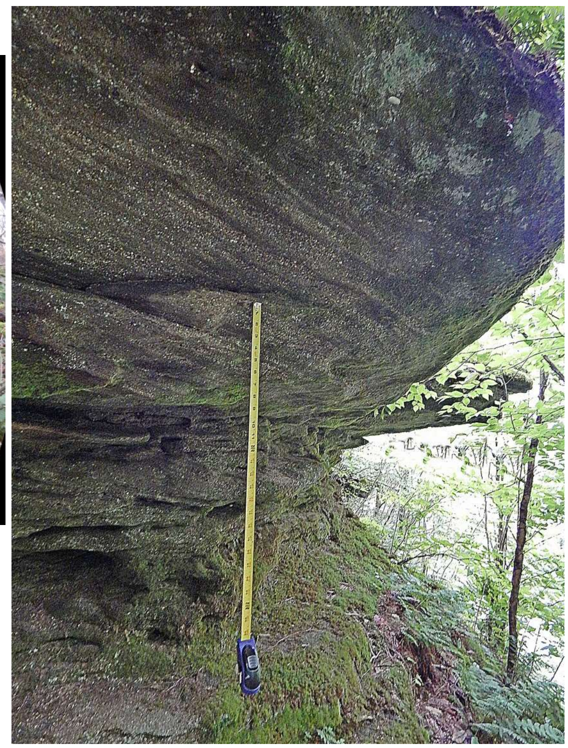

Figure 24. Top – Low-angle strataset (point bar) of large channel complex ; 5 m deep channel inferred, outcrop #3. Above – Head on view of same,. Right – Smooth channel-form, outcrop #2 (tape = 0.7 m).

Channel-forms are very common with basal surfaces which curve gently downward from truncation/pinch points (apparent channel tops/sides) into sub-horizontal planes (channel bottoms). Some incision and scouring (lag deposits) along channel bases are evident especially where channels cross-cut and stack vertically. Rarely, terraced or two-stepped channel sides were observed at outcrop #6. The contacts of basal channel deposits with wave-ripple and hummocky cross-stratified deposits (observed at outcrops #2, 3, and 6) show little relief and are essentially horizontal at outcrop scale. The symmetrical (most common) to asymmetrical wave ripples and hummocks (heights ~ 10-20 cm; wavelengths ~ 20-80 cm) are typically finer-grained (dominantly fine to coarse sand) than overlying channel deposits but pebbles are found throughout.

Interpretation of Channel Deposits:

• The low-angle stratasets are interpreted as "point" bars which migrated by erosion of cut banks and deposition on the point bars (i.e., lateral accretion deposits; LADs). The rate of movement/degree of meandering depends a number of factors such as stream gradient, bank stability, flood frequency, sediment bedload and suspended load.

• Channels and point bar deposits are well exposed in places. The sense of flow direction is less clear but bi-directional cross-beds are present. At outcrop #2 and #6, some channels weather out and overhang dramatically with steep (> 45^o) sides; some joint surface exposures are much less obvious.

• The general E-W strike of strataset beds indicates overall E-W orientation of meandering channels. An exception is several (likely-related) outcrops near the NW corner of outcrop #3 which dip generally eastward indicating a general N-S flow direction.

• One such outcrop, a large strataset of 5 m vertical thickness which overlies the mouth bar (Fig. 25), showed a sequential 90^o variation in stratal dips (120, 90, 60, 30^o upward) This swing equates to NE-SW, N-S, NW-SE, NW-SE change in flow directions for a meandering stream reach over time. Four LADs are apparent and distinctive and may represent stream and delta plain incremented response to relative sea level change. Alternatively, the relatively small direction change may indicate low sinuosity for a relatively large ($\sim$ 5 m deep) channel (probable distributary) which likely fed the mouth bar deposits beneath it. The individual strata are very coarse but display little structure, mostly flat pebbly beds of possible bedload sheet origin.

• Cross-strata dipping largely E-SE indicates a dominant current flow and associated dune migration toward the SE (shoreward) within channels.

• Bi-directional and shoreward-oriented cross-strata within channel bars/deposits is indicative of tidal currents with a dominant flood component.

• Stratification variably composed of ubiquitous pebbles (2-20mm) suggests transport by strong tidal currents with a significant tidal setup (probably meso-tidal range [2m-4m] or more as corroborated by thick foreshore deposits and very large-scale dunes elsewhere).

• Abundant low-angle stratasets and intermingled channel-forms suggest significant lateral channel migration and frequent channel avulsions as might be expected with non-cohesive channel banks and strong flood tides (directionally-opposed to ebb/fluvial currents in sediment-choked channels).

• The apparent dominance of flood-tide currents in channel and other deposits (e.g., tidal flats/delta plain and large dunes) suggests these outcrops expose a roughly shore-parallel swath through a deltaic complex

• The coarse-grained wave-ripple laminated and hummocky cross-stratified deposits observed in at the base of some outcrops formed by wave/combined-flow currents under open marine conditions as suggested by larger-scale bedforms and inferred formative waves with large orbital diameters and periods.

• The facies relationships and relatively smooth/low relief contacts between the basal channel deposits and the wave-formed marine deposits suggest a low gradient/gradual progradation of a complex of channels and associated deposits onto a low gradient marine shelf.

• Thick lateral accretion deposits exposed over a 200 m stretch of outcrop #3 and ending in a massive channel complex (~ 5 m deep) are suggestive of a large distributary oriented roughly W-NW.

• Bridge (2000) noted that Devonian Catskill river channels were smaller near the coast (i.e., sinuous, single-channel rivers, tens of meters wide, maximum depths of 4 - 5 m, sinuosity of 1.1-1.3, mean bank-full flow velocity of 0.4 - 0.7 m/s) and perhaps distributive (delta-related). With increasing distance from the coast, slopes increased, rivers became wider (up to hundreds of meters), deeper (up to 15 m), coarser grained, and possibly braided. In this study, two large channels were identified along outcrop #3, one associated with a mouth bar complex and the other about 600 m south on the Rim Trail. Based on point bar deposits, a depth of about 5 meters was inferred for both channels. Given their proximity and progradation over marine deposits, these paleo-streams are interpreted as distributaries.

• The paleo-sea level established by the beach deposits at outcrop #2 correlates with that for the interpreted mouth bar at outcrop #3 (about 5 m below the vertical control provided by the caprock). Given that sea level must have risen at least 5 m (likely on the order of 10 m) in order to form very large-scale HCS that is present at and just below the caprock with thin-bedded wave cross-laminated sandstones present on top (outcrop #5). An estimate of time for this relative sea level rise could be bracketed by the lateral accretion deposits (LAD) within the point bar above the mouth bar. Assuming that one LAD represents one flood deposit (Bridge and Demicco, 2008) or (count each distinctive stratum) and that hurricanes are the causal agent and recur on the order of 2 years (Fielding et al., 2005; for major flooding) or 10 to 50 years (Keen et al. 2006, 2012; for major storms yielding storm beds), counting the LADs (and/or each sedimentation unit) could provide a time estimate.

• Tidal channels and inter-channel tidal flats are inferred but few examples have been identified. The overlap in properties and often cryptic coarse strata present difficulties but more uniform LADs and a flared channel openings , if visible, might be helpful. Subtidal areas are most likely to be preserved and most tidal-flat deposition results from lateral accretion in association with progradation of the flat and the point bar associated with meandering tidal channels. Therefore, much of the sedimentary record for tidal-flat successions is comprised of features associated with channel fills and tidal point bars.

Mouth Bars

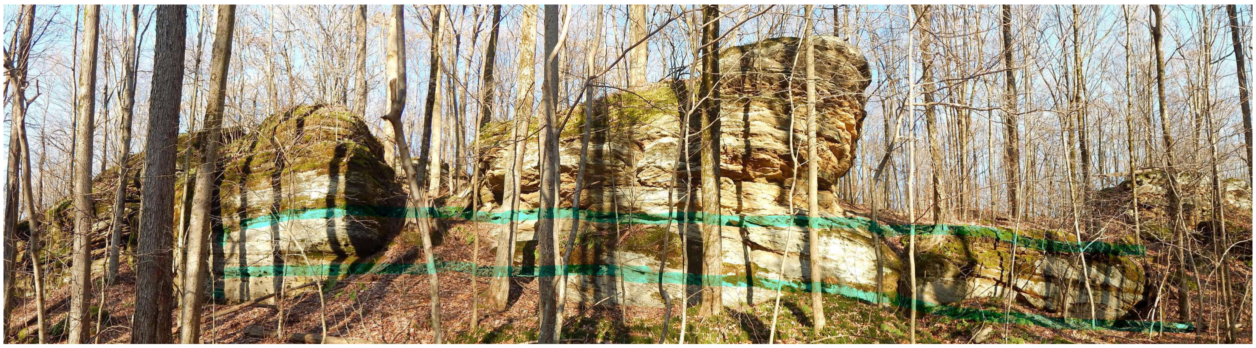

Figure 25. Mouth bar complex forms a bench beneath large channel deposits (outcrop #3 at the south turn on the Rim Trail)

Description

At the base of outcrop #3 where the Rim Trail turns south, three partially-exposed sigmoidal cross-bed sets downlap at low-angles. The overall coset is about 3 meters thick and two major bedding planes which separate sets dip at about 5º. The sigmoidal cross-bed sets average 0.5 - 1 meters thick with low-angle (~ 10-15º) cross-bedding oriented southward. The cross-strata are coarse to very coarse sandstones with thin (~ 4 mm) granule layers usually separating each stratum which average ~ 4 - 6 cm in thickness. Both inverse and normal grading is present with some pebbly strata in places. Set boundaries/major bedding planes are sharp with minor pebble lags and small wave ripples in places on the bases; apparent dip is about 5º. The basal set shows a large (~ 2 m) hummocky form at the updip end and the uppermost set top appears truncated with a large-scale ripple or a small channel. A sinsuous tube-like feature (perhaps a weathering feature) conforms with the top of the basal bed. These sigmoidal cross-beds transition laterally updip (~ 5º) along the major bedding planes into medium to thick-bedded coarse-grained trough cross-beds and HCS which are located directly beneath thick channel (low-angle stratasets dipping eastward) deposits.

Although not directly exposed, thick (2+ m) tapering beds of pebbly, massive and/or partially stratified (e.g., arched pebbles/HCS, Fig.) conglomerate are situated along the outcrop base north and south of this exposure and appear to correlate with it. The base of the conglomerate unit is sharp and irregular and is subdivided in places by bedding planes or thin sandstones; total lateral extent is about 200 meters.

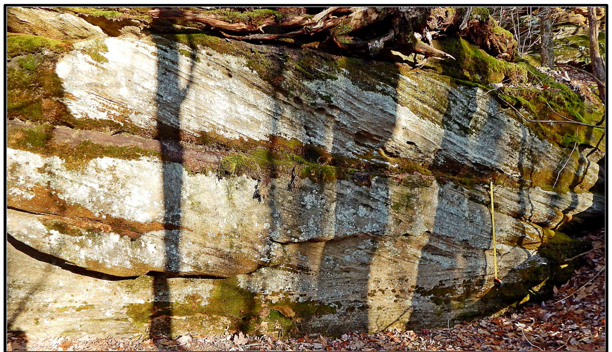



Figure 26. Mouth Bar Deposits: (A) Coset of sigmoidal x-strata; sets are over/downlapping (tape = 1 m). (B) Same as A in a float block. (C) Tapering amalgamated conglomerate beds, 2+ m; fining upward; large (L = 2+ m) isotropic pebble-laminated hummock/HCS suggests storm waves were active during formation/deposition; tape = 1 m. (D) Similar to C but more massive bedding some wavy strata above the erosive base; tape = 1 m.

<u>Interpretation:</u>

The downlapping sets of sigmoidal-cross stratification and associated trough and HCS cross-strata and the massive conglomerate unit, all located subjacent to large fluvial lateral accretion deposits (point bar/stream-reach = N-S orientation) deposits, are interpreted as mouth bar deposits related to sediment-laden floodwaters entering seawater. With stream flow expansion and deceleration, the coarsest sediment is deposited rapidly, i.e., pebbly fluvial bedload (conglomerate). Sequentially, sediment is sorted and deposited via bedload/traction processes (coarse-grained trough cross-beds transition to moderately-sorted sigmoidal cross-beds) with bypass of finer sands, silt, and clay as suspended loads. Both hyperpycnal and hypopycnal (density-constrast) flows may result from sediment-rich floodwaters whereby dense flows provide driving forces for traction and suspended loads and hypopycnal flows carry plumes of suspended sediment basinward. Marine processes, such as waves, tides, longshore, and rip currents, and storm setup with offshore downwelling or a geostrophic flow may then rework and/or redistribute these deltaic sediments as near and offshore marine deposits.

<u>Ancient examples</u> - In an array of process facies models, Tinterri (2011) noted an association of large-scale sigmoidal cross-strata with mouth bars and hyperpycnal flows as generated by sediment-laden floodwaters entering seawater. His sigmoidal cross-strata descriptions and examples are similar to the Salamanca occurrence. And as shown below, he posited a possible genetic connection between sigmoidal (mouth bar) and hummocky (delta-lobe) cross-strata via hyperpycnal and bypass flows and oscillatory pulses of floodwaters. In a polygenetic HCS review, Jelby et al. (2020) proposed a similar process (hyperpycnal flows and pulsations) for the origin of their gravelly “complex” HCS.

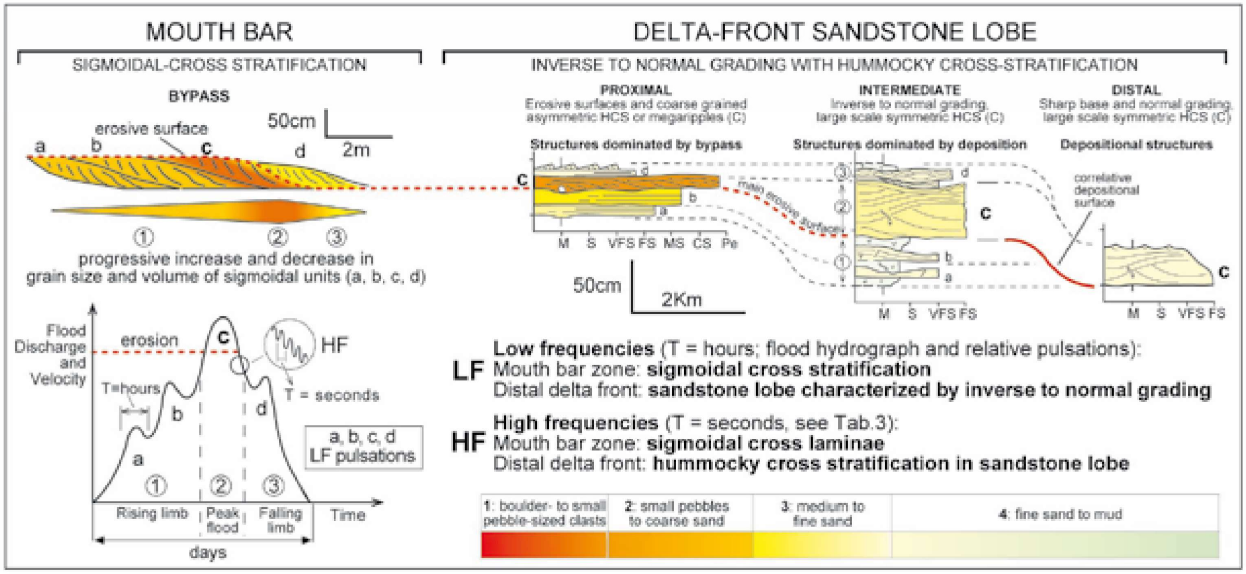

Figure 27. Proposed processes for stratal changes from mouth bar to delta front (Tinterri, 2011)

Modern/recent examples - With separate sets of overlapping/downlapping sigmoidal cross-beds divided by major bedding planes (Fig.) observed at outcrop #3, flooding events are interpreted and the upper bedding plane orientation (~ 5° southward) likely reflects the delta clinoform. Corroborating evidence comes from studies of the Australian Burkedin River and delta which are subjected to episodic flood events, mainly cyclone related (every 1 to 2 years on average) and a wet-dry monsoonal climate. Some floods transport up to several million metric tons of sediment in a few days (Fielding et al., 1999). In a study of the delta, Fielding et al. (2005) surveyed mouth bars with ground penetrating radar (GPR) which showed low-angle dips (1°) on downlapping reflections with steeper dips locally and a major bounding surface, also dipping seaward 1-2°, that separated two distinct inclined bedsets. These results were interpreted as two mouth bars (each ~ 5 m thick), the upper of which overlapped the lower and downlapped along a bounding surface (a low-angle clinoform) in a seaward direction (very similar to the smaller-scale Salamanca examples). They noted that in most places, the mouth bars are sharp-based, typically 4–6 m thick, of moderately sorted sand, and display an equilateral to elongate triangular planform over an area of several square kilometers. Further, mouth bars accumulate on a time scale of 10s of years, with much of the sediment being delivered rapidly during a small number of high-magnitude, short-duration, fluvial discharge events (each of only a few days duration). Supplied by such floods, Burdekin mouth bars aggrade and prograde onto a shallow, low accommodation shelf.

Salamanca delta analogs

The Burdekin River and delta is one of the larger drainage basins on the northeast Australian coast with a drainage area of about 130 km^2 or about 3 times larger than that inferred for the Salamanca study area (and the relative size of the mouth bars roughly scaled)

As noted above, the Arafura Sea and northern Australia is considered a modern analog for the epeiric Catskill sea and coast (Dott and Batten, 1980; Woodrow 1985). Situated on an intracratonic platform, the area is subjected to episodic cyclones (hurricanes; (Fig. 29) and a monsoonal wet-dry climate. In a predictive study of the the Australian coast, Harris et al. (2002) noted that episodic flooding, due largely to cyclones, occurs in generally small, arid, “flashy” drainage basins (50,000 to 130,000 km^2) along the northeast and Gulf of Carpenteria coastlines. Based on estimates of tidal, wave, river power, deltas were characterized as wave and/or tide-dominated. However, they noted that sediment flux rather than river power may exert a primary control on the overall geomorphology of clastic coastal depositional environments, as hypothesized by Dalrymple et al. (1992). The work of Fielding et al. (2005) and others supports this view and hence they classified of the Burkedin delta as “flood-dominated” with wave-influence. Likewise, the Gulf of Carpenteria deltas can be considered flood-dominated, built by rapid sediment flux during large episodic flooding events (major cyclones ~ 5-10 years) and susequently modified by waves and tides (mesotidal; 2 m - 4m range).

Note: A related delta classification was recently proposed by Lin and Bhattacharya (2021), “Storm-flood-dominated delta: A new type of delta in stormy oceans” which helps formalize the “flood-dominated delta” noted above. Their identification criteria include extensive sharp-based planar to hummocky cross-stratified sandstone beds, commonly presenting as large-sized gutter casts. The gutter casts are interpreted as storm channels formed by erosion resulting from offshore-oriented downwelling currents that may include localized rip currents. In conjunction with these gradient (coastal setup) flows, density (hyperpycnal/hypopycnal) flows may form from sediment-laden floodwaters yielding an efficient offshore sediment transport. Whereas gutters and shallow channels may be important to their model, gutters/channels are not distinctive as these occur in many environments (e.g., Amos at al., 2005). In any case, they propose a four-component pyramidal classification scheme of deltaic deposition to highlight storms as a distinct process and which can be ”readily applied to interpret storm-dominated environments and provide new insights into depositional processes of the marine realm.”

Figure 28. Flood-dominated deltas with tidal and wave-influence – Gulf of Carpenteria, N. Australia. The Nicholson, Albert and Leinhardt River deltas coalesce in a complex delta plain with many point and some mid-channel bars; numerous tidal creeks, separate and attached to main drainages and a sandy strandline downcurrent. Episodic flood events (large events = cyclones plus monsoons) on small flashy drainages transport large volumes of sediment to the coast.

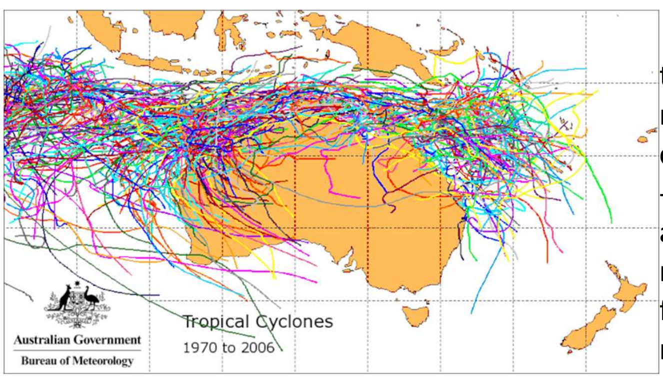

Figure 29.- 35 year track record of Australian tropical cyclones (roughly 10/year; average may be increasing). Per Fielding (2005), a cyclone influences the Burkedin delta every 1 -2 years producing major flooding. Keen et al. (2006; 2012) estimate hurricane recurrence in the Gulf of Mexico at 10 years for event bed deposition and 50 years for major hurricanes depositing thick event beds (e.g., Katrina produced a 0.5 m event bed that thinned to 0.1 m over a distance of 200 K; comparable to thicker storm beds in the geologic record).

<u>Caprock</u>

The caprock is about 15 to 75 cm in thickness, composed largely of pebbles of variable size and coarse sand, and exhibits both massive fabric and crude cross-bedding. And at outcrop #5, the largest caprock pebbles are > 60 mm (Fig. 8) and instead of 99% vein quartz, the clasts include about 25% exotics: red sandstone, metamorphic clasts, including some more equant shapes and large (pebble size) mud chips (the only mud seen thus far) which appears to be a matrix in places. Aligned plant debris is also common, aligned normal to shore, a centimeter or two below the top surface. And the top surface is often irregular and displays hummocks and other relief in places.

Storm waves have been attributed to “beheading” large subaqueous dunes at outcrop #7 as well as another incursion, 1- 2 m below the caprock in places. Given the very large HCS present at or just below the cap and the sudden introduction of exotic pebbles and abundant plant matter, it seems likely that the powerful storm(s) that produced the very large-scale HCS also produced the pebbly caprock (it appears that U_o. > 2 m/s would be required to entrain the large pebbles; the HCS/U_o. chart shows a range of 2- 3 m/s for the larger bedforms). And a flood, possibly coincident, most likely transported the large exotic pebbles and plant debris at least locally with two inferred distributary mouths nearby and less robust caprock elsewhere may have been sorted or winnowed in place.

But what accounts for the drastic and rapid change in sedimentation, apparent wave energy, and biological activity on top of the cap (assuming that the trend continued as examination of float nearby and outcrops elsewhere suggests that it did)? The fossiliferous wave-rippled thin-bedded sandstones record the first fossil occurrences in the sequence; molds of *Productella sp., Camarotechcia sp., and Crytospirifer sp.*). And the caprock appears largely undisturbed and a gradual high energy foreshore/shoreface transgression is not in evidence. Evidence points to a change in relative sea level and an eustatic rise would act to reduce stream gradients for transport of coarse sediment. And It seems that sea level was on the rise with episodes of wave ravinement near the top of sequence. Basin subsidence is appealing for possible rapidity but it does not fit the situation as well. Johnson et al. (1986) noted that sea level rise may occur fairly rapidly with mid-oceanic ridge activity or mid-plate thermal uplift while McClung et al. (2013; 2016) tie their T-R cycles in upper Devonian rocks in West Virginia to glacial cycles.

CONCLUSION

The Salamanca “flat-pebble” Conglomerate records intense episodic events such as storms and floods along a late Devonian coastline as well as periodic and predictable processes such as tides and fair weather sedimentation. Depositional environments include a progradational flood-dominated delta, pebble-rich marine strata with HCS (some very coarse-grained and of large-scale), a sub-aqueous large-scale (~ 5 m) dune field formed by meso-tidal flood tides, and a mixed sand/pebble beach. Much of the sequence records delta progradation by cyclone-driven floods and the generation of large scale, coarse-grained HCS. by storm waves. Sediment transport and redistribution along shore and near-offshore likely occurred mainly by longshore and rip currents from the wave breaker/surf zone to the shoreface. Other sediment transport mechanisms such as downwelling from coastal setup, geostrophic flows, and hyperpycnal flows may carry sorted sediment further offshore. Well-exposed channel deposits near the top of the sequence (which overlie wave-truncated dunes and beach deposits at a similar elevations) suggest an expanding delta until an apparent abrupt rise in relative sea level brought a transgression of shallow marine conditions as recorded by thin-bedded wave-formed strata and an abundant fossil fauna directly overlying a pebbly caprock.

SALAMANCA CONGLOMERATE - ROAD LOG

Meeting Point: Rock City State Forest – Little Rock City Rd. at the DEC sign/State Forest boundary

(two small parking lots; carpool if possible)

Meeting Point Coordinates: 42.225830, -78.710587

Meeting Time: 9:30 AM

Rock City State Forest is off Hungry Hollow Rd. which can be approached from US route-219 or State route-353.

Most outcrops are short hikes on trail or just off the road (longest = outcrop #3, about a mile round trip).

Bring a lunch.

Be prepared for sparse facilities; one porta-loo at road's end/outcrop #7.

Latitude	Longitude	Stop or View Description
42.2280	-78.7092	STOP 1. The "Sentinels" - obvious from the first crest of Little Rock City Rd.; well visited but respect private property. Well-weathered, isolated blocks; interpreted as tidal flats and shoals.
42.2265	-78.7137	STOP 2. The "North Face" - the highest outcrops (+10 m; which includes 4-5 m of marine strata) – shoreface/foreshore/channels
42.2259	-78.7175	STOP 3. NW corner of outcrop belt and points S on the Rim Trail, interpreted as a deltaic channelbelt w/shoreface, mouth bar, channels/point bars.
42.2217	-78.7113	STOP 4. Just east of the first rise (escarpment) on Little Rock City Rd. within RCSF. Some point bars (lateral accretion deposits) and channel fills and large x-beds and capped by beds of very large (> 7m) pebbly hummocky cross-strata (HCS). Interpreted as part of the deltaic sequence with some tidal deposits and storm intrusions.
42.2187	-78.7127	STOP 5. A few 100 meters south of #4, east side of the road, just inside treeline, a series of linear outcrops extend south for 100s of meters. Roughly 2 m exposed, mostly HCS, some wave x-lamination, occ. channels/x-beds; very large and pebbly HCS subjacent to caprock; above cap, a thin-bedded fossiliferous sandstone. Interpreted as intense storms then quiescent shallow

		marine conditions; a distinct change in wave climate perhaps best explained by a relatively rapid change in relative sea level/apparent transgression.
42.2094	-78.7108	STOP 6. East of first campsite/shelter, follow white (NCT) blazes; perhaps the most interesting (confusing?) outcrop area. A large channel complex overlying fine-grained wave-rippled sandstone (delta interpretation) with adjoining cross-strata of all scales.
42.2088	-78.7075	STOP 7. The NCT joins outcrop area #6 with #7 and following it south (white blazes) through Little Rock City provides a representative sampling of large-scale (3-5 m) cross-strata (tidal dune field) followed by HCS (dune beheading by waves), then a smattering of channel deposits (tidal or fluvial), then the caprock (a final leveling by waves).

EPILOGUE

James Hall (1843) poetically summed up these formative geologic processes and the passage of time:

"Here was an ocean supplied with all the materials for forming rocky strata: in its deeper parts were going on the finer depositions, and on its shores were produced the sandy beaches, and the pebbly banks. All, for aught we know, was as bright and beautiful as upon our ocean shores of the present day; the tide ebbed and flowed, its waters ruffled by the gentle breeze, and nature wrought in all her various forms as at the present time, though man was not there to say,

How Beautiful!"

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Escarpment to Escarpment to Escarpment: Three Vistas Reflecting the Subsurface Structure of Western New York

Paul R. Nevergold
984 Parkside Avenue, Buffalo, NY 14216

Joseph A. Butch
202 North Linden St. Amherst, NY 14221

INTRODUCTION

This trip, designed for Earth Science teachers, and the general public, presents a non-technical overview of the structural geology of western New York State. It includes stops at each of the three major escarpments, a brief hike, sampling of the escarpment capping lithologies, "snapshots" of geologic features, guided discussions, and includes historical geology interpretations.

<u>Geological Setting</u>

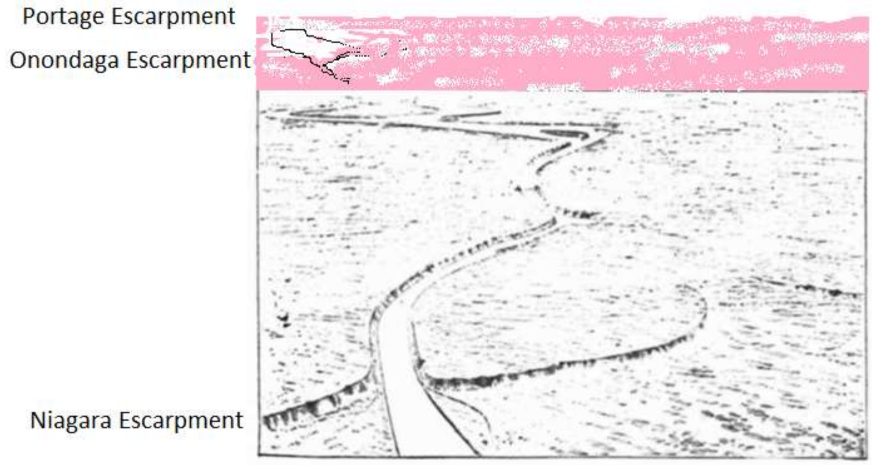
The overall topography in western New York State is cuestaform as a result of Paleozoic sediments dipping about ½ degree to the South and the land surface sloping to the North from 2000 ft above sea level in the Allegheny Plateau to 243 ft above sea level for Lake Ontario. There are three major escarpments: The Portage Escarpment at the edge of the Allegheny Plateau, The Onondaga Escarpment underlain by the Onondaga limestone and the Niagara Escarpment formed by the Lockport Group. As a result of this situation, there about 300 waterfalls within an hour and a half drive from Buffalo (Kershner 1980). Western New York falls range from the stunning Niagara Falls where about 100,000 cubic feet of water per second roar over the lip rock of Lockport dolostone to small intermittent falls of tributaries entering 18 Mile Creek.

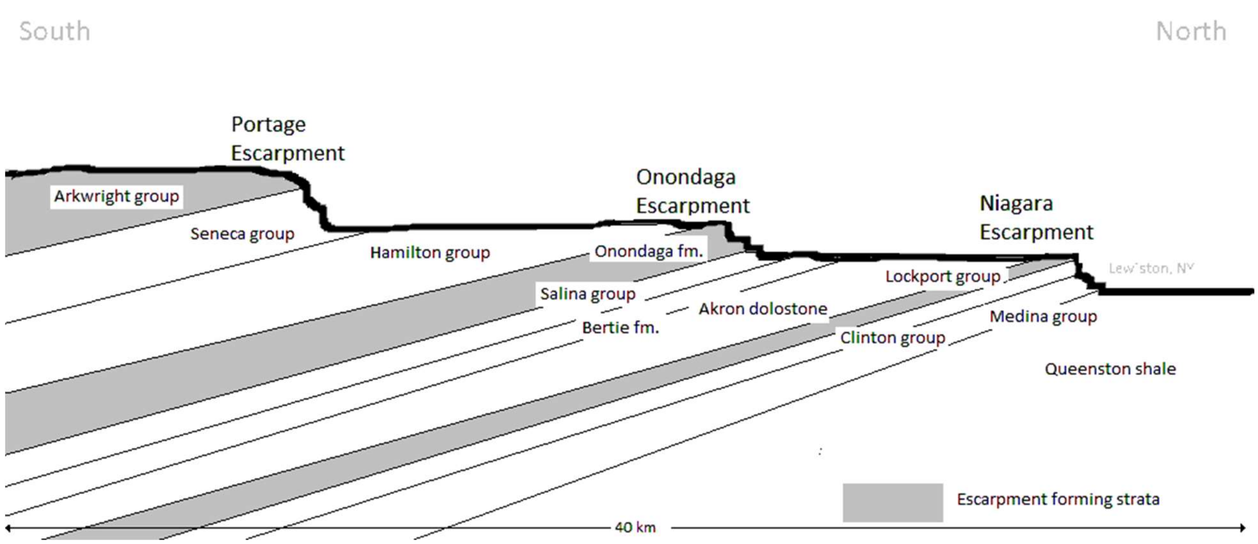
The leaders/authors have scoped out the best points in WNY where you can get a real feel of this cuestaform structure of WNY topography that are readily available to the public. We will also see and collect if you wish, some of the major formations that cap the escarpments.

<u>Previous Investigations</u>

The major escarpments in the physical geography of western New York State were recognized in this block diagram in 1901 by Amadeus W. Grabau.

Grabau 1901 fig. 16 p. 70 presents a view looking south along the Niagara River. Lewiston, New York is in the lower left-hand corner.
A highly stylized cross section from South to North is presented as follows.





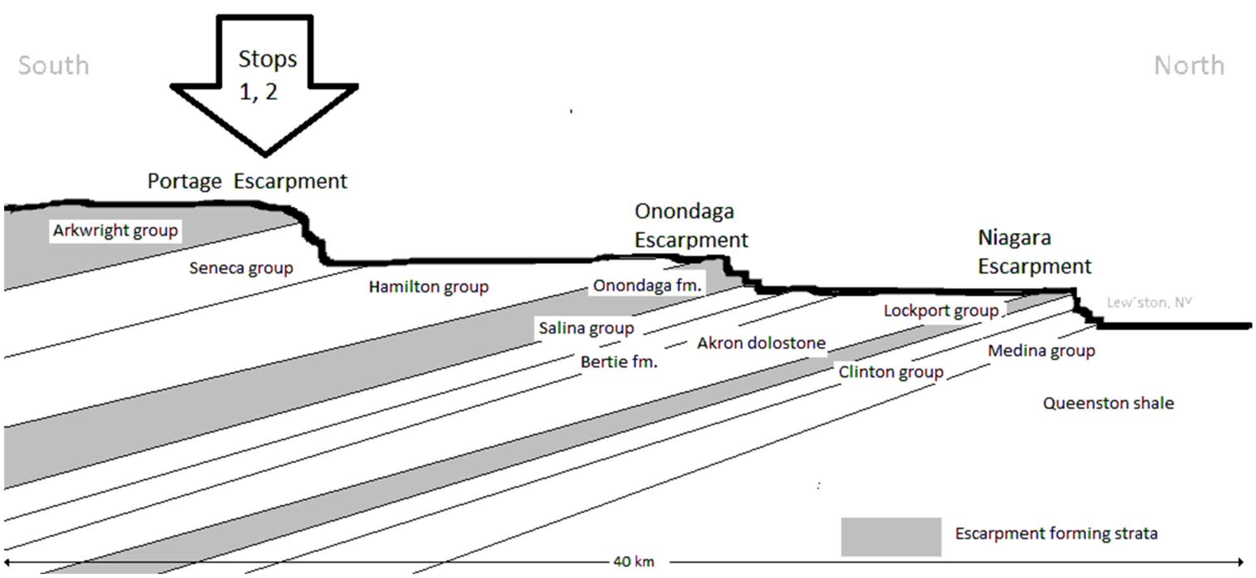
The Niagara Escarpment and the Onondaga Escarpment are underlain by resistant layers. The Portage Escarpment is the remnant of a dissected plateau.

ROADLOG AND STOP DESCRIPTIONS

Meeting Point: Buffalo State College in parking lot I-37 on the East side of the Science Bldg.
Meeting Point Coordinates: 46.56N 78.52W
Meeting Time: 8:00 AM

Distance in miles (km)		Route Description
Cumulative	Point to Point	
0 (0)	0 (0)	Exit Parking Lot, Left on Iroquois Drive to Grant Street Exit right on Grant Street Right on entrance to rt 198 West
1.3(2.1)	0.7(0.9)	Take the Scajaquada Pky. West to I 190 South.
12 (19.3)	10.7(17.2)	left on I 190 South to I 90 West to 219 South
17(27.4)	5.1(8.1)	Take the 219 South to Armor Dulles Road
17.6(28.4)	0.6(0.9)	Take Armor Dulles Road East to Rt 277 South - Chestnut Ridge Rd.
20.6(33.2)	3(4.0)	Take Chestnut Ridge Rd. South to Eternal flame parking lot. 1 mile plus a little past the park entrance.

STOP 1: Chestnut Ridge Park parking lot for the trail to the Falls of the Eternal Flame
Location Coordinates: (42.43N 78.44W)
Meeting Time: 8:40 AM



Field trip location description

Hike to Falls of the Eternal Flame (If, perchance, the eternal flame is out, we will light it and not tell anyone.) You will lose and then gain about 80 feet of altitude on the hike. The trail is rugged and should not be taken unless you are sure you can get back up. Hiking boots best, sturdy sneakers second, sandals or flip-flops forbidden. You will be following a stream bed and may get your feet wet. You should have extra socks. It will take 25 minutes down, and 30 minutes back.

Interesting sedimentary features for photos and samples can be taken. We will be hiking through outcrops of the Upper Devonian Arkwright group's Canadaway formation. The rock formation making up the sides of the small canyon we will be hiking up is the Canadaway formation, the Gawanda member to be specific. The lithology is described as gray and black shale, thin and thick bedded siltstone and occasional limestone concretions and layers. The black shale is highly petroliferous, the likely source of the gas for the "eternal flame" and on calm days can be smelled near the falls. We will point out and discuss sedimentary, topographic, and structural features along the hike.

Distance in miles (km)		Route Description
Cumulative	Point to Point	
21.6(34.8)	1(1.6)	Take Chestnut Ridge Road north to Main Park Entrance
		Turn right and park close to the toboggan slides
22.2(35.8)	0.6(1)	You are now at the edge of the Allegheny Plateau

<u>Meeting Point: STOP 2: Chestnut Ridge Park view from the edge of the Portage Escarpment</u>

Location Coordinates: (42.42N 78.45W)

Meeting Time: 9:50am

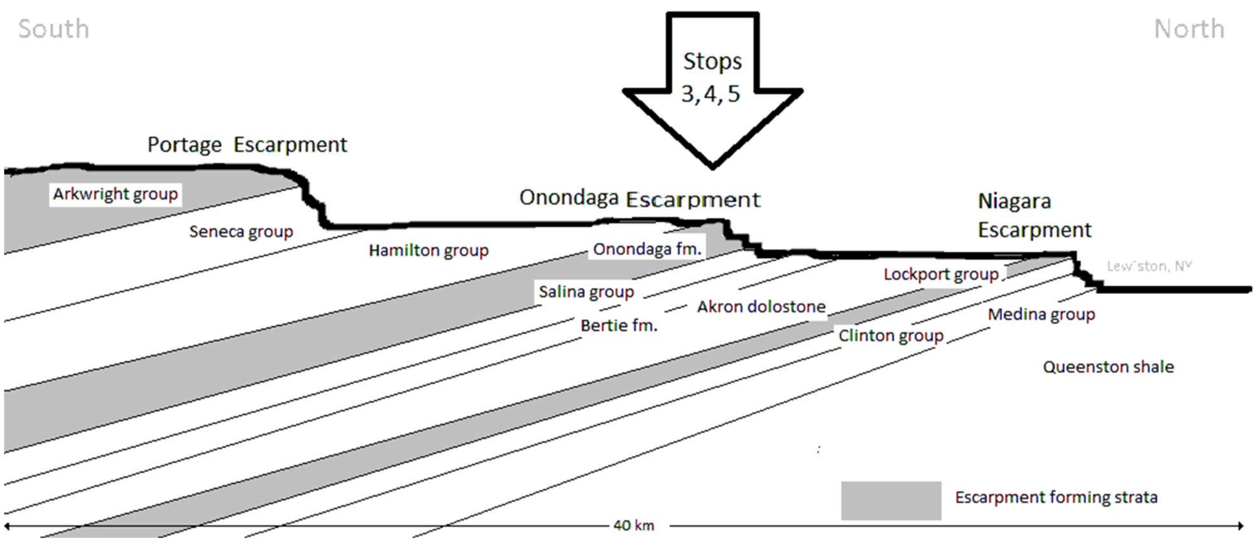
Field trip location description

The view from the edge of the Portage Escarpment will open our discussion on the three topographic vistas of western New York. Looking north you can see UB (South Campus) and the Vet's Hospital on the horizon which is located on the Onondaga Escarpment. If you look a little west (left) of Buffalo, you can see Niagara Falls on a clear day about 30 miles.

Distance in miles (km)		Route Description
Cumulative	Point to Point	
24.2(39.0)	2(3.2)	Take Chestnut Ridge Rd. North and turn left on Amor Dulles Rd.
24.8(39.9)	0.6(1)	Take Armor Dulles Rd. West to Rt 219 North
34.0(54.7)	9.2(14.8)	Take Rt 219 North to I 90 East

44.6(71.7)	10.6(17.0)	Take I 90 East past toll booths to Exit 48 Transit Rd. North
45.2(72.7)	0.6(1)	Take Transit Rd. north to Wehrle Dr. Turn East (right) on Wehrle Dr.
47.5(76.4)	2.3(3.7)	Take Wehrle Dr. East to Barton Rd.
47.9(77.0)	0.4(0.6)	Turn South (right) on Barton Rd and go to gate of quarry.

STOP 3: New Enterprise Stone and Lime Company
Location Coordinates: (42.57N 78.39W)
Meeting time: 11:00 AM



Field trip location description
New Enterprise Stone and Lime Co.
We will meet at 11:am at the quarry gates. All visitors will be required to sign a permission slip before entering the quarry. During the tour, we will stay together as a group and follow all quarry rules.
The quarried area is about one square mile. The rock layer quarried is the Nedrow member of the Onondaga Limestone which is middle Devonian in age. Quarried Onondaga was used as “dimensional stone in most of the bridge abutments of the old railroad bridges in WNY. It was considered strong and resistant enough to be used in the abutments of the Brooklyn Bridge.
It outcrops from Lake Huron and the NW shore of Lake Eri to The Hudson valley. The Nedrow member contains a lot of flint that was used by indigenous peoples for weapons and tools. In Erie County, there are three active quarries on the Onondaga and quite a few abandoned quarries.

Distance in miles (km) Cumulative	Point to Point	Route Description
48.2(77.6)	0.4(0.6)	Take Barton Rd. North to Wehrle Dr turn East (right)
50.7(81.5)	2.4(3.9)	Take Wehrle Dr. East to Shisler Rd (Note outwash deposits)
51.3(82.5)	0.6(1)	Take Shisler Rd. North to first right (Rear entrance to Clarence Town Park)

STOP 4: outcrop south of Clarence Town Park
Location Coordinates: (42.57N 78.36W)
Meeting Time: 11:50 AM to 12:30 PM

Field trip location description
We will take a close look at an outcrop south of Clarence Town Park. We will discuss the nature of the outcrop, collect flint, discuss the weathering process, and interpret how the features observed could have been formed.
Discussion questions:
How does the bedrock outcropping at this stop differ in appearance from the bedrock visible at the New Enterprise Stone and Lime Co. quarry? How would you explain this?
The calcium carbonate in the limestone is stable in an alkaline environment and rapidly decomposes in an acidic environment. The silica in the chert nodules contained in the limestone is stable in an acidic environment. How might a cherty limestone have been formed?

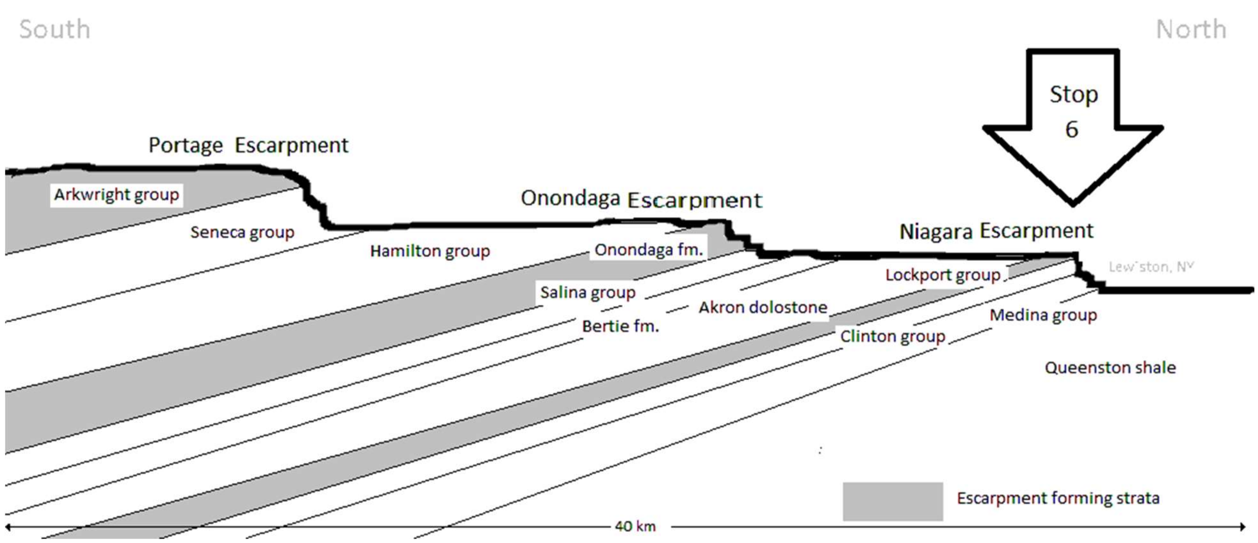
Distance in miles (km) Cumulative	Point to Point	Route Description
51.5(82.9)	0.2(0.4)	Go back to Shisler Rd. and go North (right) to Main St.
57.1(91.9)	5.6(9.0)	Take Main St East (right) to Buell St. – Rt 93 N Turn Left (North) on Buell St.
58.7(94.5)	1.6(2.6)	Take Buell St. North to Mechanic St. North to John St.
59.6(95.9)	0.9(1.4)	Turn East (right) on John St to Airport Rd.

STOP 5: Airport Road in Akron, NY
Location Coordinates: (43.01N 78.29W)
Meeting Time 1:00 PM to 1:10 PM
Field trip location description

A stop along Airport Road in Akron NY affords a view from the Onondaga escarpment and an opportunity to discuss physical geogrphy. The first draft of this field guide instructs us to “take pictures, sing & dance.” The geologic processes forming this escarpment have produced an east-west string of waterfalls across New York State including Williamsville Falls, Akron Falls, Indian Falls, Buttermilk Falls and others.

Distance in miles (km) Cumulative	Point to Point	Route Description
60.5(97.3)	0.9(1.4)	Take John St to Rt 93 North (right).
61.5(99.0)	1.1(1.7)	Follow Rt 93 to Transit Rd Rt. 78 North (right). As we travel north on Transit Road, we will cross low-lying east-west trending ridges. These are terminal moraines.
78.9(127.0)	17.4(28.0)	Take Transit Rd. North to Outwater Dr.
79.4(127.8)	0.5(0.8)	Turn South (right) on Outwater Dr. to Outwater Park.

<u>STOP 6: Outwater Park, 125 Outwater Drive, Lockport, NY</u>
Location Coordinates: (43.01N 78.42W)
Meeting Time: 2:00 PM to 2:20 PM
Field trip location description

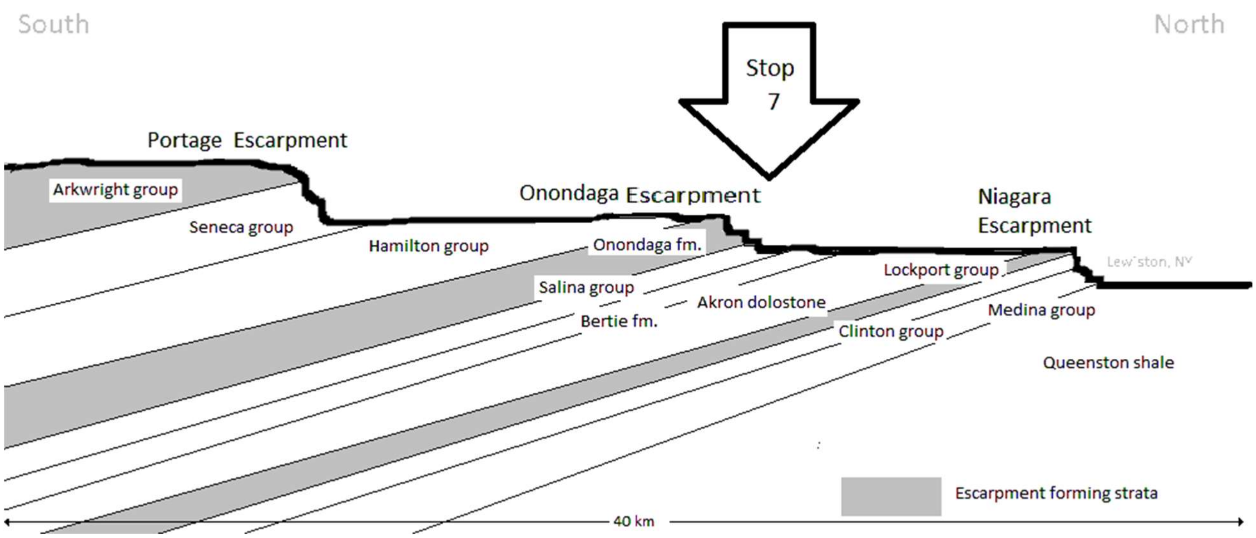


Outwater Park in Lockport New York provides an excellent safe view of Lake Ontario Lowlands. Discuss glacial Lake Iroquois and the carving of the Lake Ontario basin. On a clear day you can see Toronto about 50 miles away.
Since we will be not able to park and walk around at Stop 7, we can take time to discuss this outcrop while we are assembled here.

Distance in miles (km)	Route Description

Cumulative	Point to Point	
80.2(129.0)	0.7(1.2)	backtrack to rt 78 south, Transit Road
86.0(138.4)	5.8(9.4)	South on Transit Rd.
89.1(143.3)	3.0(4.9)	left on Millersport Highway
95.1(153.0)	6.0(9.7)	right onto I 990 south
99.0(159.4)	4.0(6.4)	290 East
102.3(164.6)	3.2(5.2)	90 west

STOP 7: "Folds" in the bedrock exposed along the Kensington Expressway, Buffalo NY.
Location Coordinates: (43.01N 78.29W)
Meeting Time 3:20 PM to 3:30 PM



Field trip location description
This is a "drive by" field trip stop. Do not stop or leave your vehicle. Through the windows of the vehicle, we will observe what appears to be folds. We will point out characteristic features of folds and look for any field expressions of these features. We will also discuss the surface expressions of buried reefs. Note the upper surface of the outcrops. How would you explain why there is bedrock in some places, and soils in others? Note, the places where bedrock is missing are often U shaped, filled in with gravely soil, and always at the ground surface. These features are remnants of pre-glacial stream erosion that are now exposed by the Route 33 roadcut.
William Oliver provides a detailed report of the Onondaga formation (Oliver 1954) and notes the location of bioherms (Oliver 1956). Bioherms in layers just beneath those exposed in the Route 33 roadcut have been reported across New York State. Upon closer inspection of what at first appears to be anticlines, no consistent trends in the symmetries of dips from the "axial centers" can be noted. Additionally, none of the characteristic features of folded terrain are observed in any of the nearby outcrops. With the absence of certain tectonic features, and considering widespread reported subsurface reefs, we can conclude that these "folds" are the

surface expression of differential compaction over buried reefs rather than the result of tectonic activity.

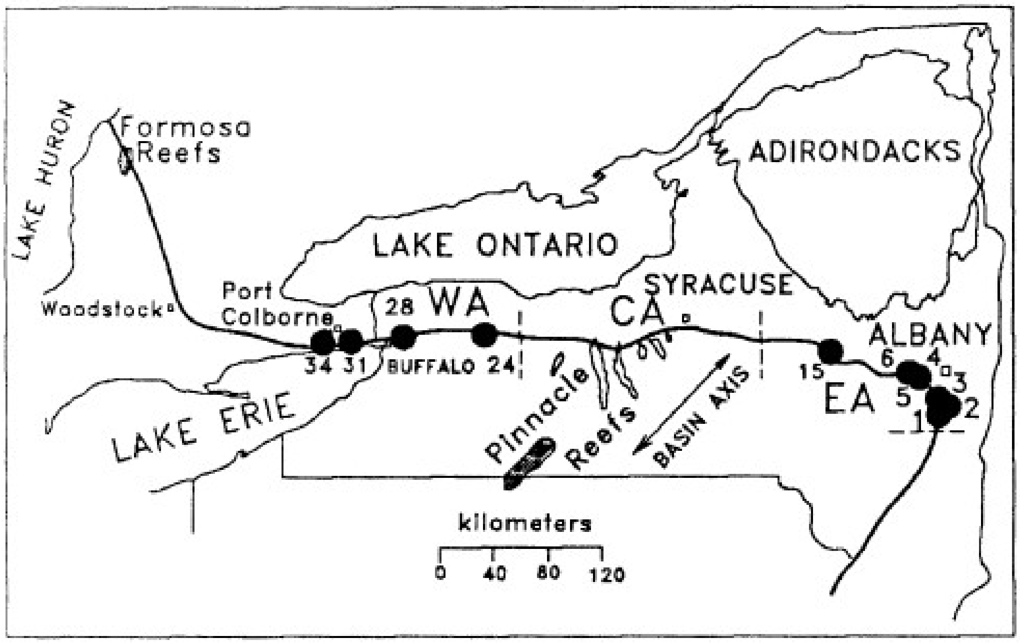


FIGURE 1—Location of reefs along Onondaga strike belt (numbered) in New York and Canada. Reef numbering follows Oliver (1976), but map includes only those reefs referred to in text: 1, Roberts Hill; 2, Albrights; 3, North Coxsackie; 4, New Salem; 5, no longer accessible exposure at south end of Thompson's Lake; 6, Thompson's Lake bioherm; 15, Mt. Tom; 24, LeRoy bioherm; 28, Buffalo Country Club reef; 31, Ridgemount bioherm; 34, reef #34. Pinnacle reefs are subsurface. Formosa reefs are believed to be Edgecliff equivalents. WA = western facies area; CA = central facies area; EA = eastern facies area. Redrawn with modification from Oliver (1976).

Woloz (1992) describes patterns of reef growth in the Edgecliff member as occurring in a variety of structures; mound/bank, thicket/bank, and ridge/bank based on the degree of paleocommunity development.

Distance in miles (km)		
Cumulative	Point to Point	Route Description
106.4(171.2)	1.0(1.6)	Continue on rt 33 west
108.4(174.4)	2.0(3.2)	198 west To Elmwood Avenue Exit
128.9(207.4)	3.2(5.2)	Return to Buffalo State
		Meeting Point: return to Buffalo State
		Meeting Point Coordinates: 46.56N 78.52W
		Meeting Time: 3:30pm

GLOSSARY OF TERMS

Bioherm … a mound like, domelike, lens like, or reeflike mass of rock built up by sedimentary organisms…

Cuesta … a hill or ridge with a gentile slope on one side and steep slope on the other…

Cuestaform … shaped like a cuesta …

Escarpment … a long, more or less continuous cliff or steep slope facing in one general direction…

Peneplain … a region with an almost flat surface …

GENERAL STRATIGRAPHIC COLUMN OF WESTERN NEW YORK

Bedrock geology from the youngest to the oldest units is as follows:

Pleistocene glacial moraine and till
Upper Devonian Arkwright group
Upper Devonian Seneca group
Middle Devonian Hamilton group
Middle Devonian Onondaga formation
Upper Silurian Salina Group, Bertie Formation, Akron Dolostone
Silurian Clinton group
Upper Ordovician – Lower Silurian Medina group
Ordovician Queenston Shale

DETAILED STRATIGRAPHIC COLUMN OF THE ONONDAGA FM IN NEW YORK STATE

Bedrock geology from the youngest to the oldest units of the escarpment forming Onondaga formation is as follows.

Edgecliff member
Moorehouse member
Nedrow member
Seneca member

DETAILED STRATIGRAPHIC COLUMN OF THE LOCKPORT GROUP IN NYS

Bedrock geology from the youngest to the oldest units of the escarpment forming Lockport Group is as follows.

Gasport Dolomite formation
Goat Island Dolomite formation
Eramosa Dolomite formation
Guelph Dolomite formation

RECONSTRUCTED CHRONOLOGY OF EVENTS

All observable geologic features are a direct consequence of the cumulative effects of every past event. In western New York the major events from oldest to youngest, are reconstructed as follows.

Condensation of rocky material and cooling of the protoplanet Earth
Rising of lighter material and sinking of heavier material
Formation of granitic rock crust into shields
Weathering, erosion and transportation of sediments
Formation of rings of sedimentary rocks around shields
Uplift of shields following loss of mass forming concentric rings of progressively younger sediments dipping at low angles away from the shields
Changes in depositional environments resulting in various kinds and amounts of sediments.
Lithification of sediments producing the present stratigraphic column of sandstones, limestones, and shales
Erosion to form a peneplain
Uplift of peneplain
Further erosion and dissection of the peneplain
Continued uplift of shield and tilting of surrounding sedimentary rock formations
Continued erosion with the more resistant layers forming escarpments
Drainage blocked by ice sheets.
Shoreline erosion of resultant glacial lakes defining sharpness of the Onondaga and Niagara escarpments
Advances and pauses in the retreats of Pleistocene ice sheets producing east-west trending moraines.
Deposition of till filling in preserves pattern of pre-glacial drainage.

Collectively, the stops on this trip provide evidence for the events outlined in the chronology above. During the discussions at each stop, we will point out the evidence in the rock, soils, and surface features for each event.

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