New York State Geological Association

93rd Annual Meeting
Field Trip Guidebook

Edited by David W. Valentino

Hosted by the Geological Sciences Program at the State University of New York at Oswego
September 23-25, 2022

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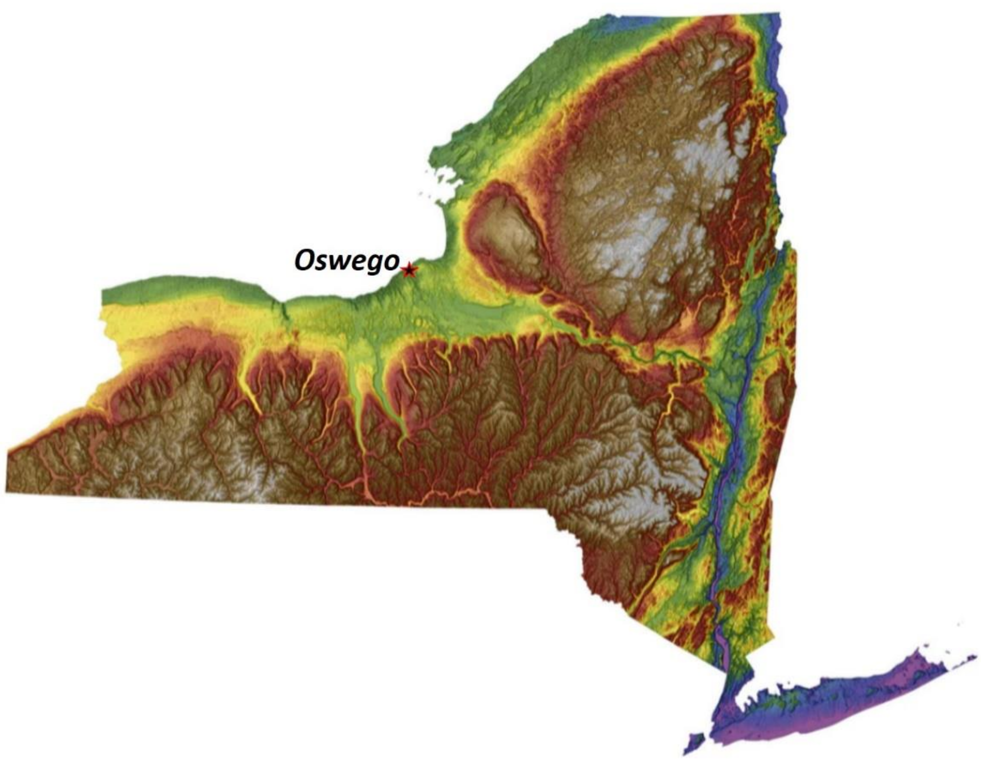

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A PRIMER ON UAV'S FOR GEOLOGICAL RESEARCH: EROSION AT MCINTYRE'S BLUFF, STERLING, NEW YORK

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INTRODUCTION

The terrain of New York state from Rochester to Syracuse is covered with tens of thousands of drumlins, which formed during the last Ice Age (Figure 1). Drumlins are composed of largely unconsolidated Pleistocene glacial till. Intense erosive forces along the southeastern shoreline of Lake Ontario have progressively exposed the interiors of many of the drumlins which intersect the shoreline (Figure 1, red circle). McIntyre's Bluff, at Sterling Nature Center, is one of the more well-exposed shoreline drumlins (Figure 2). Others include Chimney Bluffs and Sitts Bluffs. These bluffs are separated by bays (such as Sodus Bay), ponds, or wetlands (Christensen et al., 1990; McClennen and Pinet,1990).

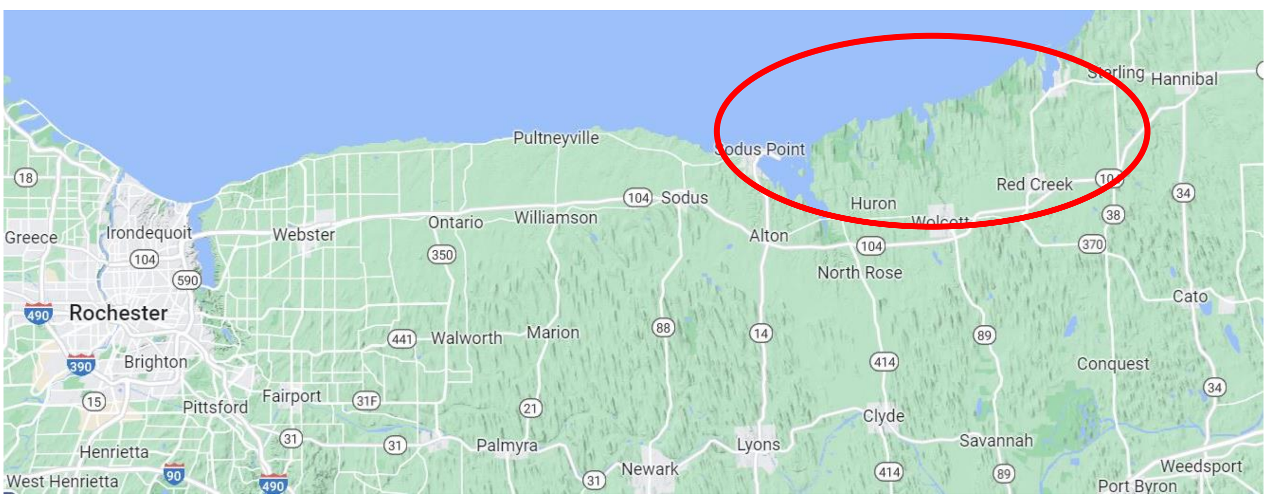

Figure 1. *Thousands of drumlins cover the terrain between Rochester and Syracuse. These drumlins formed at the end of the last Ice Age, as the ice receded north. In the area of Sterling and Sodus Point (red circle) several very large drumlins intersect the lakeshore, and have been significantly eroded by the lake. McIntyre's Bluff, near Sterling, is one of these drumlins.*

Although the bluffs add to the majesty of the Great Lakes region, the intense erosive forces they are susceptible to cause them to degrade rapidly, and significant change can occur even within a single week. This degradation is particularly exacerbated during the winters, when intense freeze-thaw cycles along the cliff face accelerate erosion. Additionally, recurrent high lake levels throughout the year expose the cliffs to considerable wave action which contributes to the erosion process. Unexpected mass failures of these cliffs can cause weeks to months of inaccessibility until nature center trails are cleared and are deemed safe for pedestrian traffic, and can also irrevocably damage sensitive habitats.

The prevalence of coastal erosion by mass wasting and wave activity in this region have been well-documented by Brennan and Calkin (1984), Pinet et al. (1992), and Pinet et al. (1997). Also, surveys conducted by Drexhage and Calkin (1981) found that the erosion and retreat of shoreline bluffs can be attributed to factors such as bluff height and steepness (the steeper the slope, the faster the erosion); bluff orientation (bluffs that face NW appear to erode significantly faster due to the prominence of northwesterly winds in the region); bluff composition (mud-rich till erodes more rapidly); beach width (wide beaches help decrease erosion rate); beach composition (cobble style beaches tend to cause bluffs to retreat more rapidly, due to the promotion of high wave energy); and lake level (undercutting is more severe with higher lake levels and more energetic wave action).

Figure 2. *McIntyres Bluff (red arrow) directly intersects the Lake Ontario shoreline. Erosion has exposed the interior of the drumlin, creating tall bluffs along the shoreline. These bluffs are composed of unconsolidated glacial till, which experiences intense erosion on a daily basis. Sitts bluff (west of McIntyre's Bluff) and Chimney Bluffs (off-map) have undergone similar erosional processes.*

The bluffs are also susceptible to biological erosion. For example, roots from vegetation progress extensively into the bluffs, and several species of birds are known to burrow into and nest within the soft bluff sediment (Figure 3a). Due to these erosive forces, it is not uncommon for the bluffs to undergo partial or total collapse and re-facing up to several times a year, especially following a rise in lake level or during and after rainy, icy or snowy conditions. Bluff faces also experience intense undercutting at times, when wave action on the lake is particularly energetic. In this case, the base of the bluff is more rapidly eroded than the top, leaving the top of the bluff unsupported from underneath, and susceptible to collapse (Figure 3B).

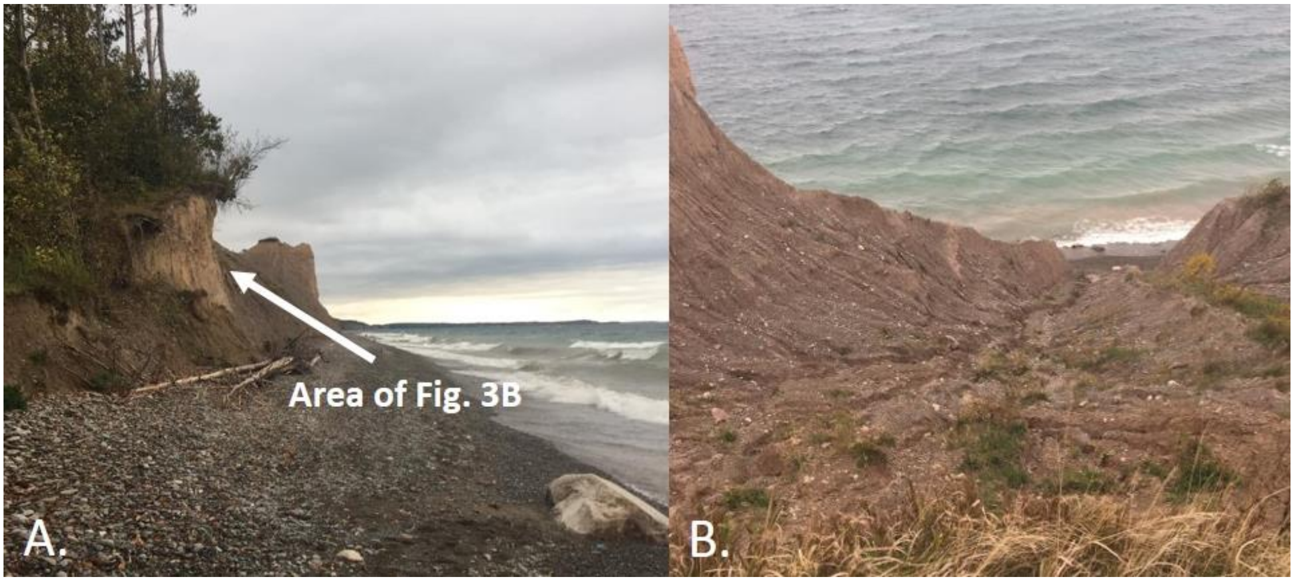

Figure 3. A) *The shoreline at McIntyre's Bluff, early mid-2021. Unconsolidated sediments which make up the bluff continually erode to the beach, and the vegetated top of the bluffs experiences periodic slides and collapses.* ***B)*** *View from the top of the bluff, looking approximately 60 feet down towards the lake. The yellow arrow indicates the location of this view. Small gullies and rivulets formed by rain, snow, and ice can be seen on the drumlin surface.*

The average erosion rate of the bluffs is roughly estimated at 1 meter per year (Pinet et al., 1997); however, this rate was far exceeded during the winter of 2018 to the present. This is partially due to an increase in occurrence and intensity of spring and fall rainfall and several larger-scale lake effect snow storms, coupled with periods of higher lake levels from the implementation of Plan 2014, which aims to restore coastal wetlands through more natural regulation of lake water levels. Collectively, this has resulted in the rapid reorganization of loose glacial sediments that mantle the shoreline, and associated collapse of portions of bluff faces onto the shoreline and beach. The bluffs are progressively receding landward as they erode, making the top surfaces of the bluffs highly susceptible to degradation. Situated at the tops of many bluffs are residences and other infrastructure, which could be threatened by the progressive erosion. Public recreational activities along the shoreline are also significantly impacted. In May of 2018, extensive erosion at Chimney Bluffs State Park forced the main access trail to the bluffs and beach to close, as it became unsafe to traverse. Intermittent closings of the trail and beach have also occurred at McIntyre's Bluff. There are no signs of the erosion rate abating along the shoreline; rather it is becoming continually more pronounced.

Over the last several years, UAV imagery has been used in conjunction with 3D modeling with Agisoft Metashape, in an effort to quantitatively characterize the erosion rate at McIntyre's Bluff (Figure 4, 5). Collected imagery has, thus far, offered critical insight into how various environmental factors, particularly those related to progressive climate change - high lake levels, mild (or harsh) winters, excessive precipitation, etc. - impact the integrity of the bluffs on a regular basis.

Figure 4: A) *Flying the Mavic UAV along the shoreline, collecting imagery of the bluffs* ***B)*** *An example of a 3D model of McIntyre's Bluff created using UAV imagery. The prominent cliff seen in A can also be seen in B. The 3D model was created using Agisoft Metashape software. Models such as this one can be directly compared with others from other time periods, and a percent change in the surface can be estimated.*

FIELD GUIDE

The meeting point is Sterling Nature Center, Sterling NY. From the parking lot of the nature center, take the Shoreline Trail (trail #3). This trail goes down to the lakeshore. Turn left after reaching the lakeshore, and continue along trail #3 (along the beach) until reaching the base of McIntyre's Bluff.

Stop #1: This is the northeastern end of McIntyre's bluff (Figure 6). The bluff is significantly shorter in this area, and severe undercutting of the bluff by Lake Ontario can be seen. Biological erosion, via species of swallows which live in the bluff, is also apparent. This stop is arguably the best place to view the ever-changing surface of the bluff. Note the unconsolidated nature of the glacial till, and the significant difference in size and composition of the sediments. We will also examine the cobblestone beach in front of the bluffs.

Stop #2: These much taller cliffs represent the highest elevation area of the drumlin (Figure 6). Trail #11 goes along the top of the bluff at this location, which can be seen in the photo in figure 3B. The top of this cliff experiences regular mass movements, especially in the winter and spring months, as freeze/thaw processes occur. Walking up into the cliffside, note the significant gullies and rivulets which have been carved out of the cliffside by rain, snow, and groundwater activity. Groundwater periodically outlets from the cliff face here as small springs. The height of the cliffs in this area (60+ feet) make it ideal for UAV imaging, as inaccessible areas can be quickly and easily imaged.

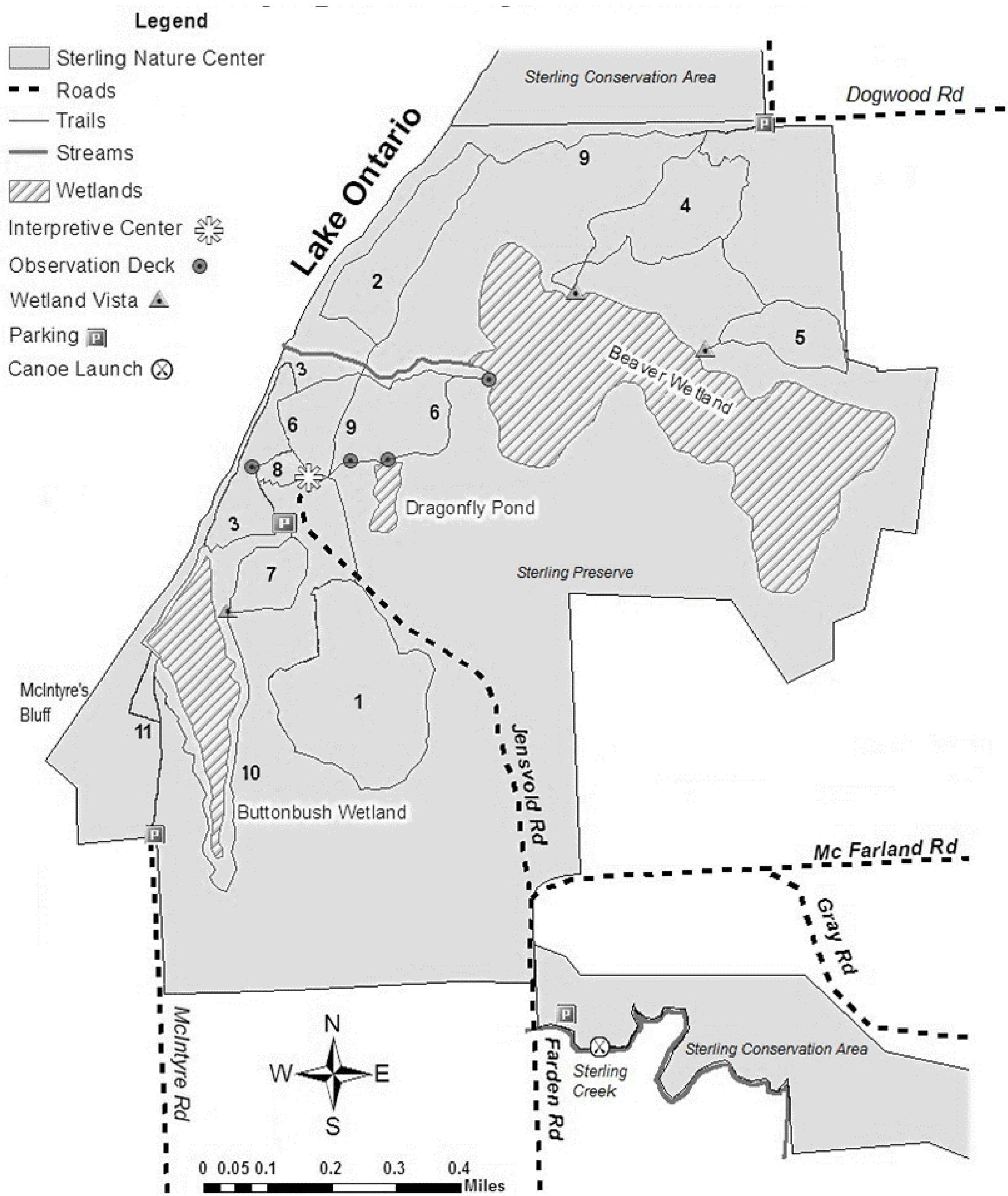

***Figure 5**. Map of Sterling Nature Center, Cayuga County, New York*

Figure 6. *Satellite view of McIntyre's Bluff. Note the intense erosion along the shoreline, which has extended back into the shoreline and up towards the top of the drumlin. Stop #1 and Stop #2 are denoted with red dots.*

REFERENCES CITED

Brennan, S. F. and Calkin, P. E., 1984, Analysis of bluff erosion along the southern coastline of Lake Ontario, New York: Dept. Geol. Sciences, SUNY Buffalo, New York Sea Grant Institute, Albany N. Y., 74 p.

Christensen, S., McClennen, C. E., and Pinet, P.R., 1990, Coastal erosion: southeastern Lake Ontario shore: Geol. Soc. America Abstracts 22, 7.

Drexhage, T. and Calkin, P E., 1981, Historic Bluff Recession along the Lake Ontario Coast, New York. Dept. of Geological Sciences, SUNY Buffalo, New York Sea Grant Institute, Albany, N. Y., 123 p.

McC1ennen, C. E. and Pinet, P. R. 1990, Southeastern Lake Ontario: a most informative field site for teaching coastal geology: Geol. Soc. of America Abstracts 22, 54.

Pinet, P. R., McClennen, C. E., and Frederick, B. c., 1992, Sedimentation-erosion patterns along the southeastern shoreline of Lake Ontario: in April, R. H., ed., New York State Geological Association Guidebook, 64th Annual Meeting, p. 155-169.

Pinet, P. R., McClennen, C. E, and Moore, L. J., 1997, Resolving environmental complexity: A geologic appraisal of process-response elements and scale as controls of shoreline erosion along southeastern Lake Ontario, New York, in (Welby, C. W., and Gowan, M. E, eds.) A Paradox of Power. Voices of Warning and Reason in the Geosciences: Geological Society of America.

Trip A-1

THE SALT OF THE EARTH AND THE SOUL OF THE SILURIAN SEA - THE UPPER SILURIAN SYRACUSE FORMATION AND ADJACENT FORMATIONS IN CENTRAL NEW YORK

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INTRODUCTION

The Syracuse Formation occurs within the Upper Silurian Salina Group in the northern Appalachian Basin. Considerable attention has been paid to this formation in the past based on the occurrence of economic quantities of rock salt (halite) and gypsum, which are of significant importance to the historical and economic development of the region. In addition to rock salt and gypsum, the Upper Silurian rocks of central New York State contain a wide variety of interesting and important geological characteristics. While considerable attention has been paid to the occurrence and distribution of halite and gypsum, less attention has been paid to the sedimentary petrology, geochemistry, and paleontology of the Syracuse Formation. The same is generally true for the underlying Vernon Formation and the overlying Camillus Formation. A significant reason for this may be the general softness/lack of resistance of the various lithologies to the processes of mechanical and chemical weathering, coupled with relatively low relief topography in the central New York outcrop belt, which collectively result in limited numbers of (and quality of) rock outcrops available for study. During recent field work, the author has noted that a significant number of the outcrops previously described in the published scientific literature are currently covered or are otherwise inaccessible due to weathering, subsequent land development activities, or other property ownership or management considerations. Luckily, we can supplement the relative scarcity of accessible rock outcrops by examining rock cores through these formations collected from the subsurface. The New York State Museum/Geological Survey maintains (and makes available for study) a collection of rock cores and well drill cuttings from across the state. Rock cores and drill cuttings are typically obtained by the State Museum from oil and gas exploration and development, salt solution mining, or construction project activities.

The goal of this field trip is to provide participants with an opportunity to visit a few of the more accessible outcrops remaining in central New York State to provide a general introduction to the lithological and paleontological characteristics of these fascinating formations. The locations of stops are shown in Figure 1. We will also stop at Cargill's Cayuga Salt Mine, a currently active salt mine located near Lansing, New York, to receive a presentation from the mine's manager (Mr. Shawn Wilzynski) and his staff regarding the mine's geology and operations. Cargill's presentation will be at the surface and we will not be visiting subsurface portion of the mine during this field trip. We are very grateful to Cargill, Mr. Wilzynski, and his team for their willingness to welcome us to their facility, provide us with a presentation on their operations, and show us some rock samples from their mine. Cargill's assistance to our group will contribute significantly to the continuing professional education of current and future professional geologists in New York State.

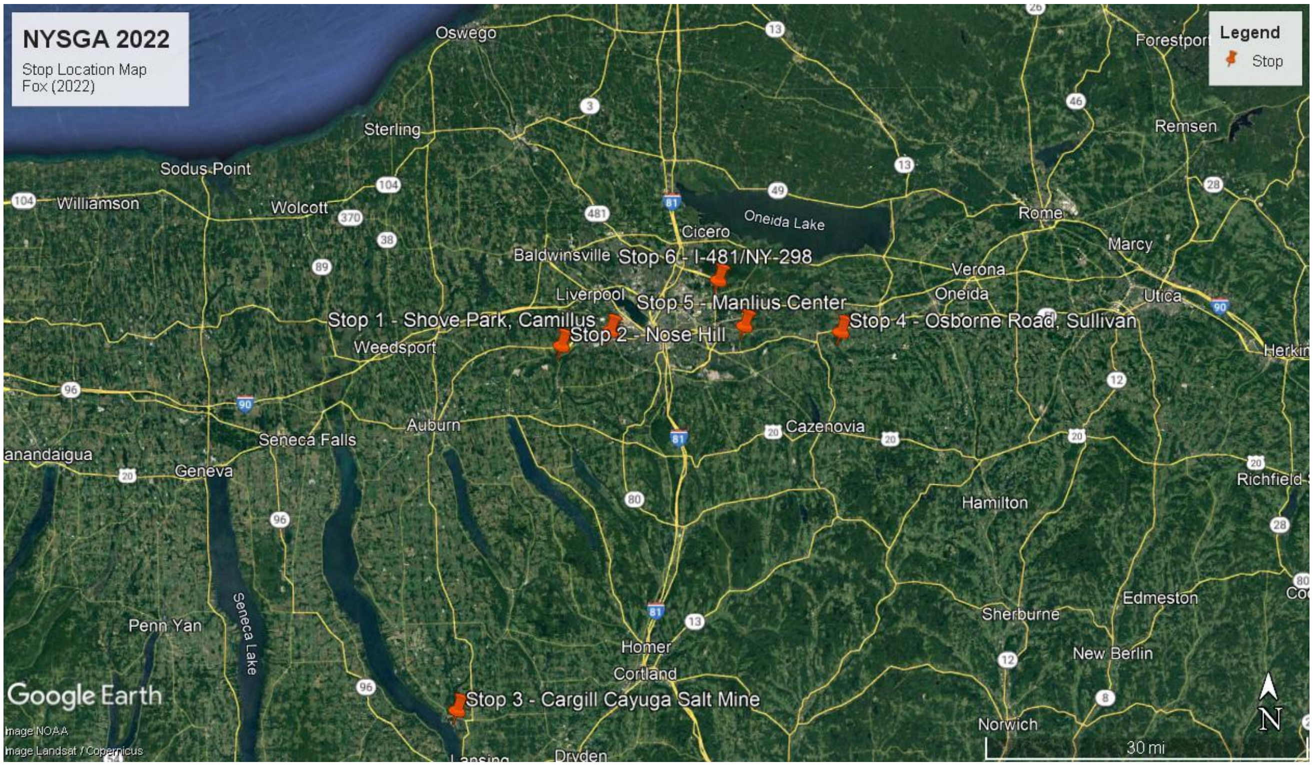

Figure 1. *Google Earth Pro map showing the locations and designations of stops for this field trip.*

GEOLOGIC SETTING

Late Silurian sedimentology and stratigraphy in upstate New York have been studied previously by Alling (1928), Leutze (1955, 1956, 1960, 1964), Fisher (1957), Rickard (1969, 1975), Treesh (1973), Ciurca (1973), Belak (1980), Brett et al. (1990), Ciurca and Hamell (1994), Vrazo et al. (2016), Mayer (2019), and others. The Syracuse Formation and adjacent stratigraphic units were deposited during late Silurian time (Pridoli Epoch) in the northern portion of the Appalachian Basin (Rickard, 1975). The Pridoli Epoch is estimated to have occurred from approximately 419 to 423 ±3 million years before present (Cohen et al., 2013). Figure 2 shows an interpretation of North American paleogeography for 425 million years before present (The Paleontology Portal, 2022). Note the position of central/upstate New York at a paleolatitude somewhere in the range of approximately 25-30° south latitude, placing the study area within Southern Hemisphere subtropical dry belt, a climate belt that is climatologically conducive to the precipitation of evaporite minerals (Sonnenfeld, 1984).

The Appalachian Basin is a foreland basin and consists of an elongate depression in crystalline basement which contains a large volume of predominantly sedimentary stratified rocks. It extends from upstate New York to central Alabama and extends east to west from the west flank of the Blue Ridge Mountains to the crest of the Findlay and Cincinnati arches and the Nashville Dome. It encompasses an area of approximately 207,000 square miles, including all of West Virginia and parts of New York, New Jersey, Pennsylvania, Ohio, Maryland, Virginia, Kentucky, Tennessee, North Carolina, Georgia, and Alabama (Colton, 1961).

The stratified rocks that occupy the basin constitute a wedge-shaped mass whose axis of greatest thickness lies close to and parallel to the east edge of the basin. The maximum thickness of stratified rocks preserved in any one part of the basin today is between 35,000 and 40,000 feet. The volume of the sedimentary rocks is approximately 510,000 cubic miles (Colton, 1961). Sedimentary sequences within the basin are grossly wedge shaped and are thickest along the eastern margin of the basin and thinnest along the western margin.

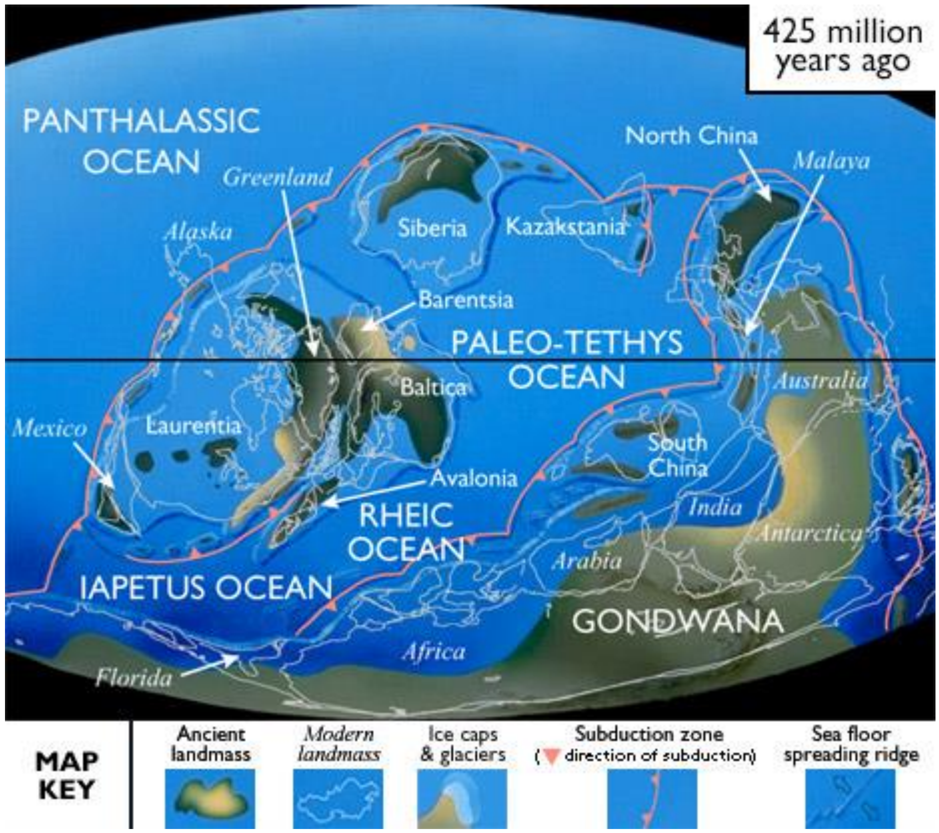

Figure 2. *Interpretation of Late Silurian paleogeography for 425 million years before present (The Paleontology Portal, 2022).*

Major tectonic events that influenced the development of the Appalachian Basin include the Taconian (Ordovician-Silurian), Acadian (Devonian), and Alleghanian (Pennsylvanian) orogenies. These tectonic events were accompanied by the accumulation of large volumes of detrital, siliciclastic sediments in a variety of depositional environments including fluvial, deltaic, beach, shallow shelf, slope, and basin margin settings. Intervening periods of relatively quiet tectonic conditions and corresponding low levels of detrital influx are marked by sequences of carbonate and evaporite sediments that accumulated over widespread areas of the northern Appalachian Basin in a wide variety of depositional environments including supratidal, sabkha, lagoonal, peritidal, shelf, slope, and basin margin settings. The carbonate and evaporite lithologies encountered within the Syracuse Formation record a period of carbonate and evaporite deposition during low periods of detrital influx from the Taconic Mountains to the east.

LITHOSTRATIGRAPHY

The Salina Group occurs within the larger Silurian-Devonian carbonate sequence. Salt-bearing units within the Salina Group occur in the north-central and northwestern parts of the basin, while a thick wedge of fine- to coarse-grained siliciclastic deposits occurs in the northeastern part of the basin (Colton, 1961). The Syracuse Formation and adjacent formations contain all three major types of sedimentary basin fill: siliciclastic, carbonate, and evaporite deposition. Carbonate rock terminology used in this manuscript is consistent with Dunham (1962), Sibley and Gregg (1987), and Mazullo (1992). Evaporite rock terminology is consistent with Sonnenfeld (1984) and Warren (2016). Detrital/siliciclastic rock terminology is consistent with Folk (1954).

Figure 3 presents a generalized stratigraphic column for the Syracuse Formation and adjacent formations based on work performed by Leutze (1955) as a graduate student at Syracuse University. In addition to a very thorough field program in locating and describing outcrops, Leutze clarified the confusing stratigraphic nomenclature previously applied to Salina Group rocks and redefined the Syracuse Formation based on a systematic inspection and evaluation of lithological and paleontological characteristics of the formations.

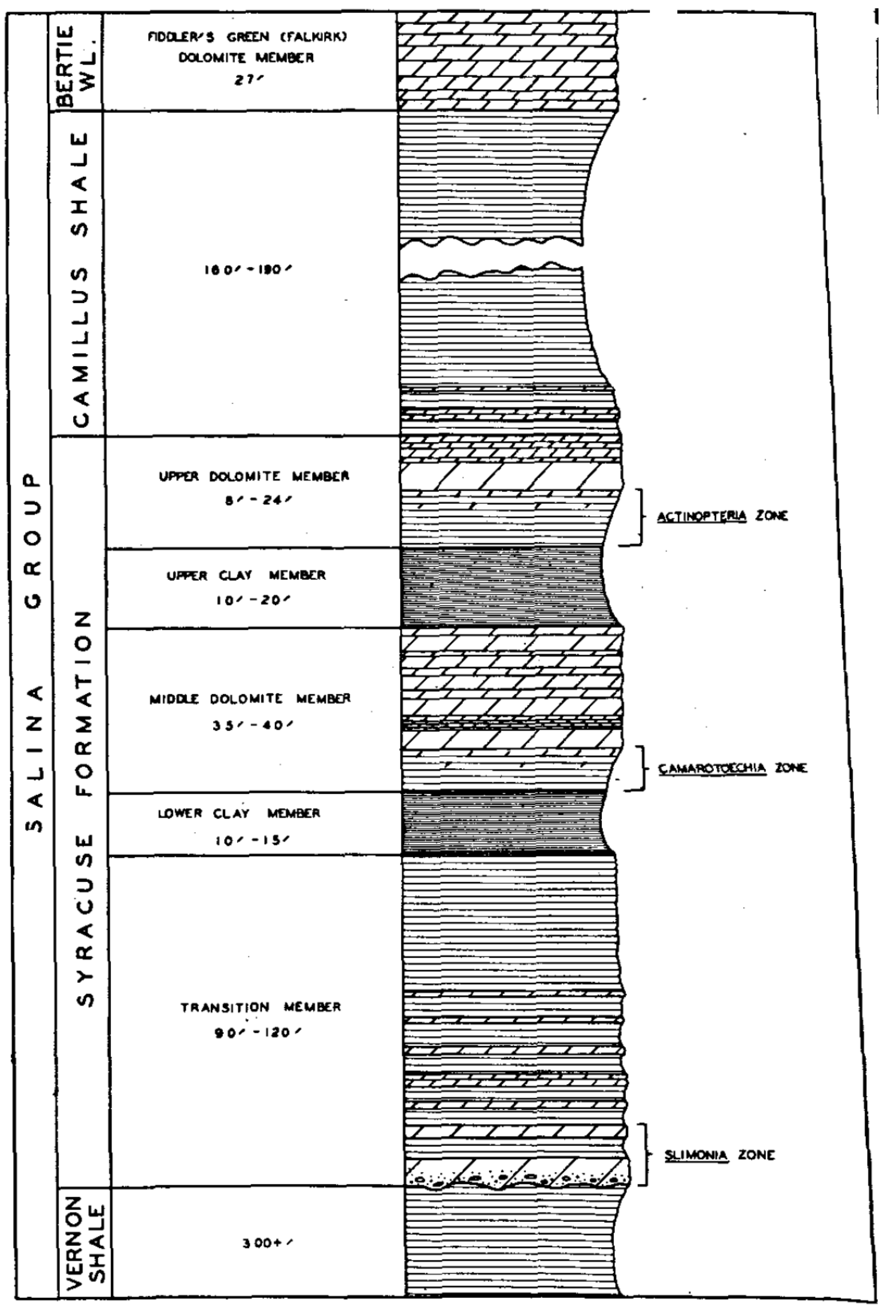

Figure 3. *Generalized stratigraphic column of the Syracuse Formation and adjacent formations in central New York State from Leutze (1956) showing range of thicknesses encountered for the various formations and*

members. Note the disconformable contact between the Syracuse and Vernon formations and the conformable contact between the Syracuse and the Camillus formations. The Slimonia faunal zone was subsequently renamed the Waeringopterus zone in Leutze (1964).

Salt (halite) has not been found to date at Syracuse Formation outcrops, but it is present in significant thicknesses (hundreds of feet) to the south in more basinal settings. Rickard (1969) performed an extensive regional correlation study of the Salina Group in New York, Pennsylvania, Ohio, and Ontario supplemented by a significant amount of subsurface data from oil and gas wells and salt mines and wells. Rickard's work clarified Salina Group stratigraphic relationships across the region and facilitated additional study of the Salina Group. Figure 4 presents an idealized stratigraphic section showing the correlations between surface outcrop units and subsurface units.

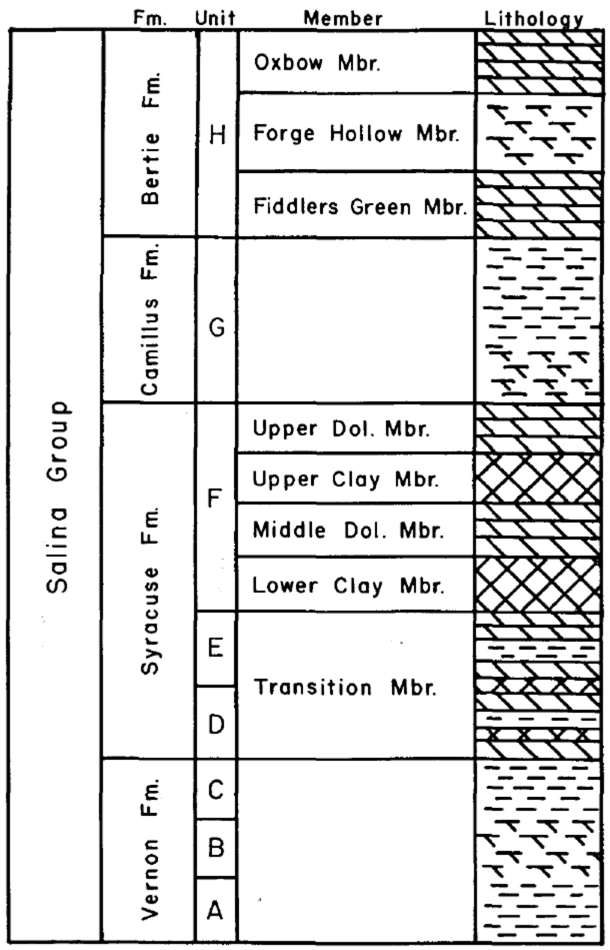

Figure 4. *Idealized stratigraphic section from Treesh and Friedman (1974) showing nomenclature, subdivisions, and correlations between Salina Group surface formations/members from Leutze (1955) and subsurface mapping units from Rickard (1969) listed in the column entitled "Unit".*

The following are brief descriptions of the lithological units at the stops on this trip, based largely on the defining work of Leutze (1955 and 1959). See the section of this manuscript entitled "Field Guide and Road

Log" for figures showing photographs of the various formations and members. Additional information can be obtained from these and other references cited at the end of this manuscript.

Camillus Formation

The Camillus Formation in central New York consists predominantly of light olive gray (5Y 6/1) to greenish gray (5GY 6/1 to 7/1) massive, dolomitic, detrital mudstone. The term mudstone is used rather than shale due to a general lack of fissility at the outcrops visited. The contact with the underlying Syracuse Formation is gradational, and the lower portion of the Camillus Formation typically contains laminated to medium-bedded argillaceous dolostones that are lithologically consistent with the dolostones of the Upper Dolomite Member of the Syracuse Formation (see below). Gypsiferous shale has been reported in the lower portion of the Camillus Formation, particularly west of Syracuse. The base of the Camillus Formation is defined by the lowest occurrence of Camillus-type mudstone.

Syracuse Formation

Upper Dolomite Member

The Upper Dolomite Member consists predominantly of brownish gray (5YR 4/1) to pale yellowish brown (10YR 6/2), laminated, argillaceous dolostone that is typically very fine-grained with non-planar anhedral dolomite texture. The general lack of recognizable carbonate grains (allochems) during petrographic examination suggests original deposition as aphanitic to very-fine grained carbonate mud/matrix, which is consistent with deposition as an evaporitic carbonate unit (Warren, 2016). The lowermost beds contain a zone of bivalve-dominated fauna originally designated by Leutze (1956) as the *Cornellites* zone and later designated by Leutze (1956) as the *Actinoptereia* zone. This member also contains interesting beds of dolostone with spherical- to egg-shaped vug-like porosity that has been described by Leutze (1955) as "vermicular rock" (see Figure 36).

Upper Clay Member

The Upper Clay Member consists predominantly of very light gray (N8) gypsum breccia and medium gray (N5) to dark gray (N3) dolomitized detrital mudstone. These deposits have been interpreted by Leutze (1959) as marking the former horizons of more soluble evaporite beds.

Middle Dolomite Member

The Middle Dolomite Member is generally similar to the Upper Dolomite Member and consists predominantly of yellowish brown (10YR 5/2) to brownish gray (5YR 4/1), laminated, argillaceous dolostone that is typically very fine-grained with non-planar anhedral dolomite texture. The lowermost beds contain a zone of brachiopod-dominated fauna designated by Leutze (1956) as the *Camarotoechia* zone. This member also locally contains a "vermicular rock" zone that is several feet thick and is generally similar to the "vermicular rock" zone in the Upper Dolomite Member, although the voids are typically smaller in size in the Middle Dolomite Member.

Lower Clay Member

The Lower Clay Member is lithologically very similar to the Upper Clay Member and consists predominantly of very light gray (N8) gypsum breccia and medium gray (N5) to dark gray (N3) dolomitized detrital mudstone.

Transition Member

The Transition Member consists predominantly of medium gray (N5) to dusky green (5G 3/2) dolomitic, detrital mudstone that is locally gypsiferous. The lower half of the Transition Member typically contains medium gray (N5) to dark yellowish brown (10YR 4/2), laminated to medium-bedded argillaceous dolostone beds that are similar to dolostone beds in stratigraphically higher members of the formation and which often contain mud cracks. The lowest dolostone bed, the base of which is an unconformable contact with the underlying Vernon Formation, is sandy and often contains pebble-sized clasts of shale/detrital mudstone and localized concentrations of an iron oxyhydroxide mineral. The lowermost two dolostone beds constitute a eurypterid fragment-dominated faunal zone originally designated by Leutze (1955) as the *Slimonia* zone. Leutze (1964) subsequently re-designated this zone as the *Waeringopterus* zone.

<u>Vernon Formation</u>

The Vernon Formation consists predominantly of moderate red (5R 5/4), massive to irregularly stratified, detrital mudstone/shale. It typically contains 5 to 15 percent sub-angular to angular quartz silt and is locally dolomitic and gypsiferous. Certain intervals of the formation in central New York (particularly uppermost beds and the middle portion of the formation) are often pale green (5G 7/2) and occasionally contain mud cracks. There is also a relatively rare medium gray (N5) mudstone/shale at the type locality that is located adjacent to the green facies, and reports of marine fauna in the Vernon Formation at the type locality occur in this gray facies (Fisher, 1957).

PALEONTOLOGY

The Syracuse Formation was deposited during semiarid to arid climatic conditions in and near a restricted marine basin as demonstrated by significant occurrences of primary and secondary evaporite lithologies (Alling, 1928; Leutze, 1959; Rickard, 1969; Treesh, 1973). Given favorable bathymetric profiles, restricted marginal marine basins in semiarid or arid climates typically develop an antiestuarine water flow regime, where normal salinity ocean water flows into the restricted basin at the surface to compensate for water losses due to evaporation and outflow of bottom brines. Surface waters in these restricted marginal marine basins can occasionally contain very high levels of organic productivity (Sonnenfeld, 1984; Warren, 2016).

Overall, Salina Group macrofossils are typically sparse, diversity is relatively low, and preservation of calcareous remains is typically poor. However, some macrofossils are locally to regionally common and are well preserved, particularly in the faunal zones of the Syracuse Formation. The macrofossils found in the Syracuse Formation are typically associated with normal marine salinity conditions, thereby demonstrating the occurrence of relatively normal marine salinity conditions at least temporarily during deposition of parts of the Syracuse Formation. Illustrations of the more common macrofossils from several of the references cited are provided below.

Camillus Formation

As mentioned above, all rocks between the Vernon Formation and the Bertie Formation were previously designated as "Camillus" by early investigators (Swartz et al., 1942). Most reports of fossils occurring in the Camillus Formation are from these early investigations and occur in rocks that are currently defined as the Syracuse Formation. Leutze (1964) stated that no significant fossils have been reported from the Camillus Formation. The author's observations during field work performed to date are consistent with Leutze (1964).

Syracuse Formation

Leutze (1955, 1956, 1959, 1964) reports that three faunal zones typically occur within the Syracuse Formation where fossils are most common and best preserved (Figure 3). These faunal zones are relatively thin (generally 1-4 feet in thickness) and interestingly occur stratigraphically above and in close proximity to the unconformity between the Vernon and Syracuse Formations (the *Wareingopterous* zone) and the Lower Clay and Upper Clay members (i.e., the *Camarotoechia* and *Actinoptera* zones, respectively), which may be the remnants of formerly thick evaporite deposits. Fossils also occur locally in some dolostone or shale beds between these zones, but fauna most typically occurs in one or more of these three faunal zones.

The following illustrations with genus and species identifications show some of the more common macrofossils collected from the Syracuse Formation by Leutze (1955). Other less common fauna that occur are not listed/shown below. Taxonomic level terminology revisions at the order level or higher since Leutze's work have been incorporated into this manuscript; however, no attempt has been made to evaluate for potential revisions at the genus or species levels (i.e., genus and species designations listed below are as listed in Leutze's work from the 1950s and 1960s).

Phylum Brachiopoda

Class Rhynchonellata

Figure 5. *Camarotechia cf. litchfieldensis (2X); internal cast of an adult pedicle valve. Middle Dolomite member.*

Figure 6. *Camarotechia cf. litchfieldensis (2X); internal cast of a small brachial valve. Middle Dolomite member.*

Phylum Mollusca

Class Bivalvia

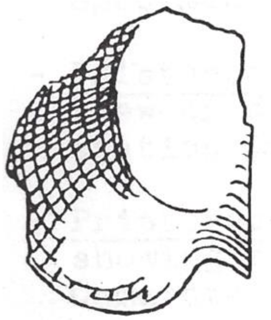

Figure 7. *Cornellites sp. (1X); fragmentary cast of left valve. Upper Dolomite member.*

Figure 8. *Ctenodonta cf. saliensis (1X); cast of left valve. Middle Dolomite member.*

Figure 9. *Rhytimia cf. buffaloensis (1X); cast of left valve. Middle Dolomite member.*

Figure 10. *Cleidophorus ? sp. (1X); cast of left valve. Middle Dolomite member.*

Class Gastropoda

***Figure 11**. Loxonema ? bertiense (2X); composite sketch of several specimens. Middle Dolomite member.*

Figure 12. *Holopea ? sp. (10X); dolomite replacement of original shell material; external shell partly broken away shows columella. Middle Dolomite member.*

Class Cephalopoda

Figure 13. *Pristoceras cf. timidum (2X); side view of conch showing color bands. Middle Dolomite member.*

Figure 14. *Pristoceras cf. timidum (1X); a septum, showing the siphuncle. Middle Dolomite member.*

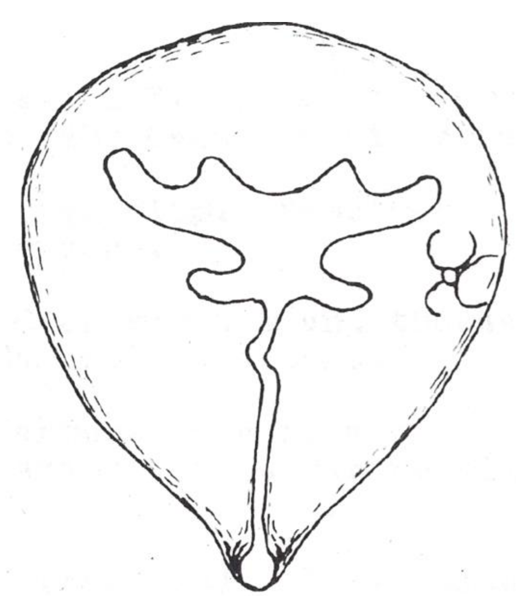

Figure 15. *Pristoceras cf. timidum (3X); cast of an aperture with an ophiuroid (a brittle star, Phylum Echinodermata) attached to the left side. Middle Dolomite member.*

Figure 16. *Pristoceras cf. timidum* (1X); side view of a cast of the living chamber and part of the conch. Middle Dolomite member.

Phylum Arthropoda

Order Eurypterida

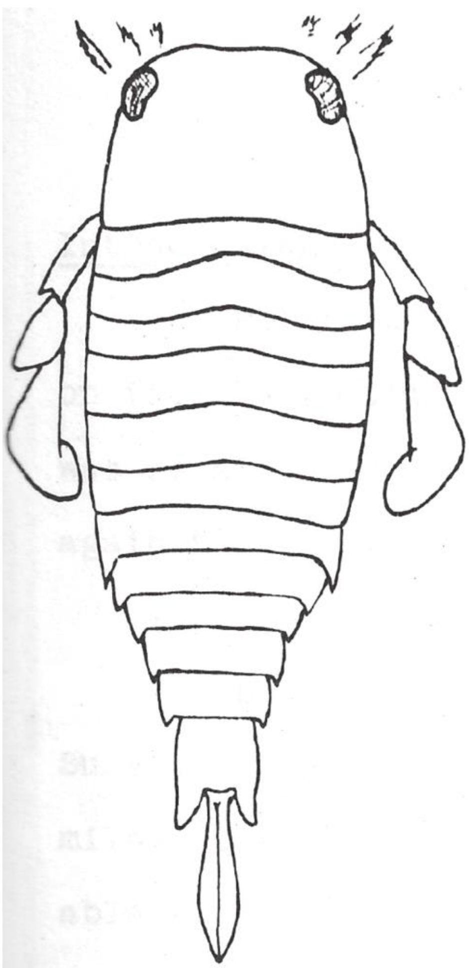

Figure 17. *Waeringopterus apfeli (1X); a composite sketch of specimens. Transition member.*

Figure 18. *Waeringopterus apfeli (2X); mold of right compound eye showing delicate cross hatching. Transition member.*

Class Ostracoda

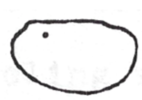

***Figure 19**. Leperditia cf. scalaris (1X); mold of right valve. Middle Dolomite member.*

Phyllum Hemichordata

Subclass Graptolithina

Figure 20. *Medusaegraptus graminiformis (1X); cast of rhabdosome covered with a carbonaceous film. Middle Dolomite member.*

Vernon Formation

The Vernon Formation is generally unfossiliferous with the exception of some isolated beds that generally occur near the middle of the formation (Fisher, 1957). As with the Syracuse Formation, the fauna in the Vernon Formation is marine. Fisher (1957) and Leutze (1960 and 1964) indicate that fauna reported in the Vernon Formation includes bivalves, cephalopods, gastropods, brachiopods, the remains of jawless, armored fishes (Phylum Chordata, Class Pteraspidomorphi, Order Cyathasporida), clawed eurypterids (Family

Pterygotidae), and a siphonophore (Phylum Cnidaria, Class Hydrozoa). Overall, the fauna in the Vernon and Syracuse formations appear to be generally similar.

The following illustrations with genus and species identifications show some of the more common macrofossils collected by others from the Vernon Formation as illustrated and described in Fisher (1957) and Flower and Wayland-Smith (1952). Taxonomic level terminology revisions at the order level or higher since the 1950s have been incorporated into this manuscript; however, no attempt has been made to evaluate for potential revisions at the genus or species levels.

Phylum Brachiopoda

Class Lingulata

***Figure 21**. Lingula allingi (2X); note the foramen, the completely enclosed slit-like passage for the opening of the pedicle, slightly apical of the pedicle valve. Also note the prominence of the radially-arranged costellae along the valve margin.*

Phylum Mollusca

Class Gastropoda

***Figure 22.** Poleumita vernonensis (2X); apical view of an immature specimen showing four whorls.*

Class Bivalvia

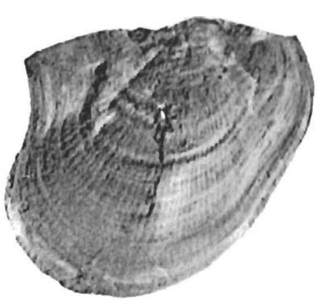

Figure 23. *Pterinia wayland-smithi (2X); mold of left valve; typical cancellate pattern is well shown in this earliest pectinate genus.*

Class Cephalopoda

Figure 24. *Spyroceras sp. (2X); portion of conch showing narrow camerae and moderately-arched septa.*

Phylum Chordata

Class Pteraspidomorphi

Order Cyathasporida

Figure 25. *Vernonaspis leonardi (2X); interior of dorsal plate showing pineal body, branchial impressions, semicircular canals, and impression of brain.*

Figure 26. *Archegonaspis ? sp. (2X); interior of ventral plate, whitened, etched, showing cancellous layer and the interrupted canals of the lateral line system; in the lower left quadrant, the plate is removed completely, showing the external mold of the pattern of the dentin area.*

Phylum Arthropoda

Class Ostracoda

Figure 27. *Leperditia scalaris (3X); mature specimen.*

ACKNOWLEGEMENTS

Sincere gratitude and appreciation are offered to Shawn Wilczynski and the entire team at Cargill for allowing us to visit the Cayuga Salt Mine and for their time and effort in providing us with a very interesting and informative presentation. Thanks also are offered to Brian Slater and Marian Lupulescu of the New York State Museum/Geological Survey, who provided access to and shared their significant knowledge of rock core, drill cuttings, and mineral samples from the phenomenal collections of the New York State Museum, which assisted ongoing research and the development of this field trip. Thanks also to Dave Valentino (SUNY Oswego Department of Atmospheric and Geological Sciences) and Nick DiFrancesco (University at Buffalo Department of Geology) for interesting discussions that initiated and facilitated the development of this field trip and for reviewing the draft manuscript. Thanks also go to Trisha House of SUNY Oswego's Department of Atmospheric and Geological Sciences for administrative and logistical assistance in setting up the field trip, and to SUNY Oswego geology students Alex Robinson and Nathan Drake for assisting their fellow geology students by agreeing to drive the college vans procured for this field trip, and for enhancing the sustainability aspects of this field trip by resulting in a reduced number of vehicles and associated reduction in carbon emissions to the atmosphere.

FIELD GUIDE AND ROAD LOG

General Suggestions: Parking is limited at some stops, so please carpool with others as much as possible. Many of our stops are along active roads, so please remain alert for vehicular traffic at all times. **It is required that you bring and wear a high-visibility fluorescent safety vest at the stops**, and hard hats are recommended as well if you plan to locate yourself beneath or close to rocks that are above you. Please remember to dress for the weather and wear waterproof field boots. Portions of trails can be narrow and slippery, so please watch your step and tread safely. Please bring your own rock hammer and hand lens. Access to some beds at outcrops can be facilitated if you have a camper's shovel to make "stairs" for yourself through talus that occurs at the base of many outcrops. Please stay hydrated and protect yourself from biting insects with whatever means you deem appropriate. In my experience, four very effective techniques (when used together) include wearing pre-treated clothing (e.g., Insect Shield®), tucking your pants into your boots, tucking your shirt into your pants, and tucking your shirt sleeves into your gloves. Guidance for protection from biting insects is available from the Centers for Disease Control here: Avoid bug bites | Travelers' Health | CDC. Please bring your own lunch and non-alcoholic beverages. We will try to stop as a group occasionally for bathroom breaks along the route, as restrooms generally are not available at most of the stops. If you are having any issues during the field trip, please let me know and we'll do our best to assist you.

Meeting Point (Stop 1): Shove Park, Town of Camillus, Onondaga County, New York - Look for and park near a blue Honda Pilot with New York license plate number BEB-5439.

Meeting Point (Stop 1) Coordinates: 43° 2'21.13"N, 76°15'6.91"W

Meeting Date/Time: 24 September 2022/8:30 a.m.

We'll walk together from the meeting point upstream through the woods along Geddes Brook to an outcrop which exposes the upper portion of the Syracuse Formation (Upper Dolomite Member) and the overlying Camillus Formation (mostly covered). The contact between these two formations is gradational and mostly covered at this location, but there is good exposure of the Upper Dolomite Member here. Relatively resistant dolostone beds and less resistant argillaceous dolostones below form a low-relief waterfall near the top of the Upper Dolomite Member.

Figure 28. *View looking southwest at Stop 1 along Geddes Brook at a portion of the Upper Dolomite Member of the Syracuse Formation.*

Distance (miles)

Cumulative Mileage	Point-to-Point	Route Description
0.0	0.0	Meeting Point (Stop 1)
0.2	0.2	Head northeast on Shove Park Dr. toward Slawson Dr.
0.5	0.3	Turn left onto Slawson Dr.
0.6	0.1	Turn right onto Rowena Dr.
3.5	2.9	Turn left onto West Genesee St.
5.7	2.2	Turn left onto NY-174 S/Elm St; safely pull off the road far to the right at the intersection with Forward Road (Stop 2).

Stop 2: Nose Hill, Town of Camillus, Onondaga County, New York

Stop 2 Coordinates: 43°1'10.43"N, 76°20'27.85"W

This stop provides an excellent opportunity to observe stratigraphic units in the upper portion of the Syracuse Formation, including the Upper Clay and Upper Dolomite members. The contact with the base of the overlying Camillus Formation is also exposed. The Middle Dolomite Member of the Syracuse Formation is also present in this outcrop; however, it is mostly covered by talus near the base of the outcrop.

Figure 29. *View looking west at the outcrop exposed at Stop 2, which shows in ascending order the Middle Dolomite Member (largely covered by talus; "MD"), the Upper Clay Member ("UC"), and the Upper Dolomite Member ("UD") of the Syracuse Formation. The overlying Camillus Formation ("CM") is also shown.*

Distance (miles)

Cumulative Mileage	Point-to-Point	Route Description
5.7	1.1	Turn right onto Forward Road
6.8	5.8	Turn left onto NY-321 S
12.6	2.0	Slight right onto Old Seneca Turnpike
14.6	1.0	Turn left onto County Line Rd.
15.6	2.6	Turn right at the first cross street onto US-20 W
18.2	1.4	Turn left onto Town Hall Rd.

19.6	1.3	Turn right onto Melrose Rd.
20.9	1.2	Turn left onto Oakridge Rd
22.1	0.6	Turn right onto E Lake Rd./Owasco Rd.
22.7	0.3	Turn left onto White Bridge Rd.
23.0	1.4	At the traffic circle, take the second exit onto Sand Beach Rd.
24.4	27.2	Turn left onto NY-34 S
51.6	0.7	Continue straight onto NY-34B N
52.3	0.9	Turn left onto Portland Point Rd
53.2	0.0	Turn right into the Cargill Cayuga Salt Mine facility and safely park, being careful not to park in a manner or location that may disrupt traffic flow for any facility vehicles (Stop 3).

Stop 3: Cargill Cayuga Salt Mine, Town of Lansing, Tompkins County, New York

Stop 3 Coordinates: 42°31'57.05"N, 76°31'45.08"W

Cargill Deicing Technology operates this mine, providing customers with deicing technology and road salt that saves lives, enhances commerce, and reduces environmental impact.

In 1915, John Clute organized the Rock Salt Corporation on Portland Point in Lansing. In 1916, a shaft was sunk to -1500 feet, but the salt was of poor quality. By 1918 the mine was still not producing well and was shut down. In 1921, Frank L. Bolton and John W. Shannon bought the mine and further sank the shaft to -2000 feet to a better vein of salt which was 99% pure. The operation was managed by Frank Bolton, and then his wife Lucie when he died. The mine was operated by the Cayuga Rock Salt Company, Inc. until Cargill acquired the mine in 1970 and modernized the mine with new beltlines for salt haulage, ventilation updates, a new shaft, and new diesel-powered equipment. The mine produces approximately 2 million tons of road salt annually that is shipped to more than 1500 locations throughout New York State and the northeastern United States. The mine employs more than 200 full-time employees in operations, maintenance, engineering, finance, management and support positions.

Currently, the mine is advancing north up Cayuga Lake and is approximately one mile past Taughannock Point. The salt is mainly sold in the road deicing market in the Northeast (including New York, Vermont, and Pennsylvania) but is also sold under the Diamond Crystal name as residential deicing salt.

The Cayuga Mine operates under a Mined Land Reclamation Permit issued by the DEC in 1975. Permit renewals were approved in 1985, 1997, 2003, 2007, 2012, 2019 and modifications were approved in 2013, 2015 and 2017. In each case, DEC determined that there would be no significant adverse environmental impact from mining.

In the mid-1990s, Cargill undertook a thorough environmental assessment of the region to verify Cargill's understanding of the geology. This work has served as a critical foundation for the continued evaluation and study of mine stability and operations. Data collected through these continuing studies is regularly evaluated and verified by numerous third-party experts employed by DEC and Cargill.

Figure 30. *Salt-mining operations within Cargill's Cayuga Salt Mine.*

Since 2002, the Cayuga Mine has submitted an annual report on the status of special conditions outlined in the permit. The primary focus is on mine stability, three-year mine plans and water management. DEC utilizes the services of Dr. Vincent Scovazzo, an internationally recognized mine stability expert from the John T. Boyd Company, for mine stability review and participation in the annual on-site visit and inspection of the mine. Over the years, Cargill has relied on several similarly credentialed experts to assist its own mine stability and other technical evaluations. Each year Cargill performs an internal risk review with leaders to evaluate and manage the risks associated with mining.

In addition, the federal Mine Safety & Health Administration (MSHA) oversees mine safety such as adequacy of underground ventilation and emergency access. MSHA inspects the mine every calendar quarter. Each of these inspections typically lasts six to eight weeks.

The mine's manager, Mr. Shawn Wilczynski, and his staff, will provide us with a presentation on the mine's geology and operations. I have requested that they also show us specimens of rock salt and other lithologies from the Syracuse Formation from their mine. The total thickness of the Syracuse Formation at the mine exceeds 800 feet, including a total salt thickness exceeding 400 feet. The mine historically has produced salt from either the F1 unit (No. 4 Salt) or the D unit (No. 6 Salt) of the Syracuse Formation. Stratigraphically, the F1 unit is interpreted to occur within the Lower Clay Member and the D unit is interpreted to occur within the Transition Member of the Syracuse Formation (Figure 3). Cargill is currently mining the D unit (No. 6) salt.

Distance (miles)

Cumulative Mileage	Point-to-Point	Route Description
53.2	0.9	Turn left onto Portland Point Rd.
54.1	1.2	Turn right onto NY-34B S
55.3	10.4	Keep right to continue on NY-34B S/Peruville Rd. (NOTE: we are really heading east on this leg of the trip, so don't be confused by "NY-34B S").
65.7	0.9	Turn left onto School St.
66.6	3.5	Continue onto McLean Rd.

70.1	1.2	Turn left onto Luker Rd.
71.3	2.4	Turn left onto NY-281 N/West Rd.
73.7	0.5	Turn right
74.2	0.7	Follows signs for I-81 N
74.9	19.5	Merge onto 1-81 N
94.4	0.4	Take exit 15 toward US-11/US-20/Lafayette
94.8	0.2	Merge onto US-11 S
95.0	9.8	Turn left onto US-20 E
104.8	4.7	Turn left onto Pompey Center Rd.
109.5	1.3	Turn right onto Enders Rd.
110.8	5.2	Turn right onto NY-173 E
116.0	0.5	Turn right onto Madison St.
116.5	0.7	Continue onto Perryville Rd.
117.2	1.1	Turn left onto Osborne Rd.
118.3	0.0	Safely pull far off the road to the right (Stop 4).

Stop 4: Osborne Road, Town of Sullivan, Madison County, New York

Stop 4 Coordinates: 43° 2'15.24"N, 75°49'51.85"W

The Middle Dolomite Member of the Syracuse Formation is well exposed at this locality, and this locality reportedly is one of the better locations in central New York to find fossils from the *Camarotoechia* zone (Leutze, 1956). The fossils are generally well preserved in gray shale near the base of the Middle Dolomite Member. The fauna is concentrated in the lowermost few feet of the member and fossils become rare higher in the section. Wavy, parallel bedding and lamination, which are locally discontinuous, are commonly observed at this locality.

Figure 31. *View looking east-southeast at a portion of the outcrop at Stop 4 showing the Middle Dolomite Member of the Syracuse Formation. More resistant, wavy-bedded dolostones near the top of the outcrop are gradationally underlain by argillaceous dolostones, which are gradationally underlain by predominantly shale/detrital mudstone. The contact with the underlying Lower Clay Member may be covered by talus near the base of the outcrop.*

Distance (miles)

Cumulative Mileage	Point-to-Point	Route Description
118.3	1.1	Head southwest on Osborne Rd toward Evans Rd.
119.4	0.7	Turn right onto Perryville Rd.
120.1	0.5	Continue onto Madison St.
120.6	3.2	Continue straight onto NY-5 W/West Genesee St.
123.8	4.1	Turn right onto NY-290 W
127.9	0.6	Continue straight, road turns into NY-257 S
128.7	0.0	Carefully cross over and pull off safely on the left (south) side of the road into the gravel parking lot (Stop 5).

Stop 5: Manlius Center, Town of Manlius, Onondaga County, New York

Stop 5 Coordinates: 43° 2'45.38"N, 76° 0'25.21"W

We will walk from the parking lot north along the former railroad bed for approximately one tenth of a mile. The former railroad bed is somewhat overgrown with vegetation, but there are excellent exposures of the Syracuse Formation, particularly on the eastern hillside. Although there is no formal type section designated for the Syracuse Formation and the lowermost member of the formation (the Transition Member) is not exposed here, this large outcrop was cited by Leutze (1959) as an ideal reference section for the Syracuse Formation given its accessibility, relative completeness, and clarity of the stratigraphic section. The exposed portion of the Syracuse Formation at this location is approximately 64 feet thick and the stratigraphic units exposed here include (in ascending order) the Middle Dolomite Member (37 feet thick), the Upper Clay Member (11 feet thick), and the Upper Dolomite Member (16 feet thick). Talus typically obscures much of the Middle Dolomite Member. Access to exposed rock can be facilitated by digging "steps" in the talus using a camping shovel.

Figure 32. *View looking east standing on the former railroad bed at Stop 5 showing an exposure of the upper portion of the Syracuse Formation.*

Figure 33. *Photo showing the upper portion of the Syracuse Formation including (in ascending order) the Middle Dolomite Member (mostly covered by vegetation; "MD"), the Upper Clay Member ("UC"), and the Upper Dolomite Member ("UD").*

Figure 34. *Closer view of the Upper Clay Member showing gray to dark gray dolomitic carbonate mudstone and light-colored gypsum (selenite). Note the presence of wavy, parallel bedding suggesting preservation of*

some primary depositional texture. However, some discontinuous, wavy, non-parallel bedding and brecciated texture is also present, suggesting leaching, dissolution, and recrystallization (secondary evaporite mineralization).

Figure 35. *Closer view showing the Upper Dolomite Member at Stop 5. The thickness of the exposed interval shown in this photograph is approximately 15 feet.*

Figure 36. *Close-up view of "vermicular" dolostone collected from the Upper Dolomite Member at Stop 5.*

Distance (miles)

Cumulative Mileage	Point-to-Point	Route Description
128.7	0.6	Turn right out of the parking area to head north on NY-257 N
129.3	1.7	Turn left onto NY-290 W
131.0	3.3	Turn right onto Fremont Rd.
134.3	1.0	Turn left onto NY-298 W
135.3	0.1	Turn right onto the I-481 S ramp
135.4	0.0	Pull off far to the right and park prior to the curve in the on-ramp (Stop 6).

Stop 6: Intersection of NY Route 298 and I-481, Town of DeWitt, Onondaga County, New York

Stop 6 Coordinates: 43° 6'25.24"N, 76° 3'15.56"W

Our last stop on this trip will be a brief stop at a small exposure of the Vernon Formation. The Vernon Formation in central New York is often moderate red (5R 5/4; see Figure 37); however, certain intervals of the formation (particularly the middle portion in central New York) are often pale green (5G 7/2) as seen at this stop (Figure 38). Given its softness and lack of resistance to weathering, significant outcrops of the Vernon Formation are rare, particularly in areas of relatively low topographic relief.

Figure 37. *Photo showing an exposure of predominantly moderate red Vernon Formation. This outcrop is not visited on this field trip and is shown for information purposes only. Intersection of NY-51 and County Highway 16, Town of Litchfield, Herkimer County, New York.*

Figure 38. *Photo showing an exposure of predominantly pale green Vernon Formation at Stop 6, Town of DeWitt, Onondaga County, New York.*

REFERENCES CITED

Alling, H.L., 1928. The geology and origin of the Silurian salt of New York State. New York State Museum Bulletin 275, 139 p.

Belak, R., 1980. The Cobleskill and Akron members of the Rondout Formation: Late Silurian carbonate shelf sedimentation in the Appalachian Basin, New York State. Journal of Sedimentary Research, v. 50, p. 1187-1204.

Brett, C.E., Goodman, W.M., and LoDuca, 1990. Sequences, cycles and basin dynamics in the Silurian of the Appalachian Foreland Basin. Sedimentary Geology, v. 69, p. 191-244.

Ciurca, S.J. Jr., 1973. Eurypterid horizons and the stratigraphy of the Upper Silurian and ?Lower Devonian of western New York State. New York State Geological Association 45th Annual Meeting Guidebook, p. 90-105.

Ciurca, S.J. Jr. and Hamell, R.D., 1994. Late Silurian sedimentation, sedimentary structures and paleoenvironmental settings within an eurypterid bearing sequence (Salina and Bertie Groups), western New York State and southwestern Ontario, Canada. New York State Geological Association 66th Annual Meeting Guidebook, p. 457-488.

Cohen, K.M., Finney, S.C., Gibbard, P.L., and Fan, J.-X., 2013 (updated February 2022). The International Commission on Stratigraphy International Chronostratigraphic Chart. Episodes, v. 36, p. 199-204; URL: http://www.stratigraphy.org/ICSchart/ChronostratChart2022-02.pdf

Colton, G.W., 1961. Geologic summary of the Appalachian Basin, with reference to the subsurface disposal of radioactive waste solutions. United States Geological Survey Open File Report 62-28, Washington (D.C.), 121 p.

Dunham, R.J., 1962. Classification of carbonate rocks according to depositional textures. *In* Ham, W.E. (ed.), *Classification of Carbonate Rocks*. American Association of Petroleum Geologists Memoir No. 1, p. 108-121.

Flower, R.H. and Wayland-Smith, R., 1952. Cyathaspid fishes from the Vernon shale of New York. Bulletin of the Museum of Comparative Zoology, v. 107, no. 6, Harvard University, Cambridge, Massachusetts, p. 355-387.

Folk, R.L., 1954. The distinction between grain size and mineral composition in sedimentary rock nomenclature. Journal of Geology, v. 62, p. 344-359.

Kjellesvig-Waering, E.N. and Caster, K.E., 1955. The Pterygotidae of the Silurian Vernon shales of New York. Journal of Paleontology, v. 29, p. 1042-1047.

Leutze, W.P., 1955. Stratigraphy and paleontology of the Middle Salina in central New York. Unpublished M.S. thesis, Syracuse University, Syracuse, New York, 193 p.

Leutze, W.P., 1956. Faunal stratigraphy of Syracuse Formation, Onondaga and Madison counties, New York. American Association of Petroleum Geologists Bulletin, v. 40, p. 1693-1698.

Leutze, W.P., 1959. Stratigraphy and paleontology of the Salina Group in central New York. Unpublished Ph.D. dissertation, The Ohio State University, Columbus, Ohio, 463 p.

Leutze, W.P., 1960. Silurian fish fossils in the Salina Basin. Geological Society of America Bulletin, v. 71, p. 215-218.

Leutze, W.P, 1964. The Salina Group. In Prucha, J. J. (Ed.), New York State Geological Association, 36th Annual Meeting Guidebook, p. 57–65.

Mayer, S.M., 2019. Paleoecology and taphonomy of some eurypterid-bearing horizons in the Finger Lakes Region of New York State. New York State Geological Association, 91st Annual Meeting Guidebook, p. 47-73.

Mazzullo, S.J., 1992. Geochemical and neomorphic alteration of dolomite: a review. Carbonates and Evaporites, v. 7, p. 21-37.

Rickard, L.V., 1969. Stratigraphy of the Upper Silurian Salina Group: New York, Pennsylvania, Ohio, Ontario. New York State Museum and Science Service Map and Chart Series No. 12, Albany, New York, 57 p.

Rickard, L.V., 1975. Correlation of the Silurian and Devonian rocks in New York State. New York State Museum and Science Service Map and Chart Series Number 24, 16 p.

Sibley, D.F. and Gregg, J.M., 1987. Classification of dolomite rock texture. Journal of Sedimentary Petrology, v. 57, p. 967-975.

Sonnenfeld, P., 1984. Brines and Evaporites. Academic Press, Inc., Orlando, Florida, 613 p.

Swartz, C.K., Alcock, F.J., Butts, C., Chadwick, G.H., Cumings, E.R., Decker, C.E., Ehlers, G.M., Foerste, A.F., Gillette, T., Kindle, E.M., Kirk, E., Northrop, S.A., Prouty, W.F., Savage, T.E., Shrock, R.R., Swartz, W.H., Twenhofel, W.H., and Williams, M.Y., 1942. Correlation of the Silurian formations of North America. Geological Society of America Bulletin, v. 53, p. 533-538.

The Paleontology Portal, 2022. The Silurian - North America (accessed 5 August 2022). University of California Museum of Paleontology, the Paleontological Society, the Society of Vertebrate

Paleontology, and the United States Geological Survey. http://paleoportal.org/index.php?globalnav=time_space&sectionnav=period&period_id=14#29

Treesh, M.I., 1973. Depositional environments of the Salina Group (Upper Silurian) in New York State. Unpublished Ph.D. dissertation, Rensselaer Polytechnic Institute, Troy, New York, 189 p.

Treesh, M.I. and Friedman, G.M., 1974. Sabkha deposition of the Salina Group (Upper Silurian) of New York State. Fourth International Symposium on Salt, Northern Ohio Geological Society, Cleveland, p. 35-46.

Vrazo, M.B., Brett, C.E., and Ciurca, S.J. Jr., 2016. Buried or brined? Eurypterids and evaporates in the Silurian Appalachian Basin. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 444, p. 48-59.

Warren, J. K., 2016. Evaporites – A Geological Compendium (Second Edition). Springer, Switzerland, 1813 p.

ASSESSMENT QUESTIONS FOR CONTINUING PROFESSIONAL EDUCATION CREDIT (FOR NEW YORK-LICENSED PROFESSIONAL GEOLOGISTS)

We will verbally address these questions and discuss the answers in the field at the end of the field trip.

1. During what period of geologic time and approximately how long ago (in millions of years before present) was the Syracuse Formation deposited?
2. True or false: the Syracuse Formation contains siliciclastic, carbonate, and evaporite lithologies.
3. Describe the most predominant color(s) for each of the following formations:
    a. Camillus Formation
    b. Syracuse Formation
    c. Vernon Formation
4. What is the name of the faunal zone contained within the Transition Member of the Syracuse Formation, and what type of animal is the zone named after?
5. What is the approximate annual production of rock salt (in tons) at Cargill's Cayuga Salt mine?

Trip A-2

IGNEOUS AND METAMORPHIC ROCKS OF THE BLACK AND MOOSE RIVER VALLEYS, WESTERN ADIRONDACK HIGHLANDS, NY

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INTRODUCTION

Trip A-2 IGNEOUS AND METAMORPHIC ROCKS OF THE BLACK AND MOOSE RIVER VALLEYS, WESTERN ADIRONDACK HIGHLANDS, NYThe Black and Moose Rivers, the largest rivers in the western Adirondack Highlands, have exposed numerous Middle Proterozoic metamorphic rocks, with the former largely paralleling the unconformity between Ordovician sedimentary rocks of the Tug Hill Plateau and Proterozoic gneisses of the Adirondack Highlands (to the east; Figure 1). Consequently, the region has attracted sedimentologists, stratigraphers, paleontologists, and metamorphic & igneous petrologists. Our trip, however, will focus almost entirely on the Middle Proterozoic meta-igneous and meta-sedimentary rocks.

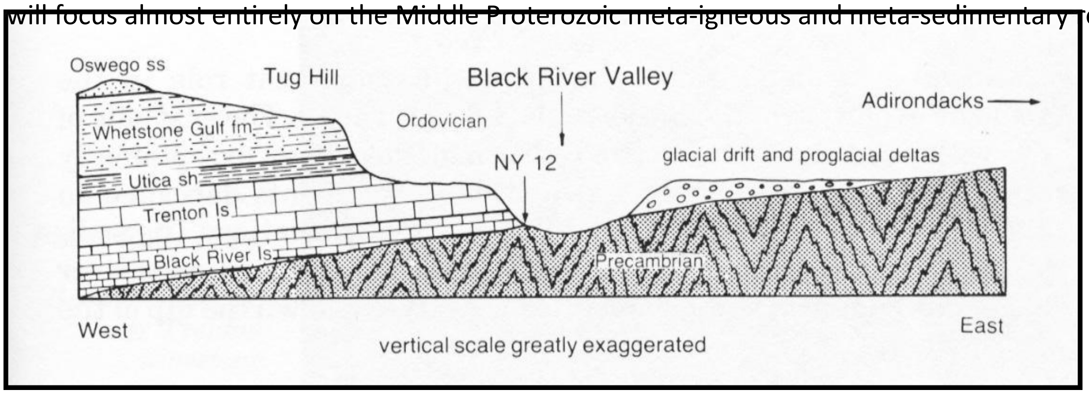

Figure 1. *Simplified west-east cross section across the Black River Valley showing relationship of Ordovician sedimentary rocks of the Tug Hill Plateau to Middle Proterozoic metamorphic rocks of the Adirondack Highlands. From Van Diver (1985).*

The Adirondack region of northern New York State is a roughly circular dome of high-grade metamorphic and igneous rocks that form a southeast extension of the Grenville Province in Canada (inset of Figure 2). The Adirondack Highlands comprise mostly meta-igneous rocks (anorthosites, charnockites, mangerites, gabbros, and granites) whereas the Adirondack Lowlands comprise mostly meta-sedimentary rocks (calc-silicates, marbles, metapelites). The Lowlands are separated from the Highlands by the Carthage-Colton mylonite zone (CCMZ), which shows extensive down-to-the-northwest relative motion from the collapse of the Ottawan phase of the Grenville Orogenic cycle (Rivers, 2008). The CCMZ is a metamorphic age and facies boundary as well, with upper amphibolite facies and *Shawinigan age* (1190-1140 Ma) metamorphic rocks in the Lowlands and granulite facies and *Ottawan age* (1090-1020 Ma) metamorphic rocks in the Highlands (Darling and Peck, 2016)

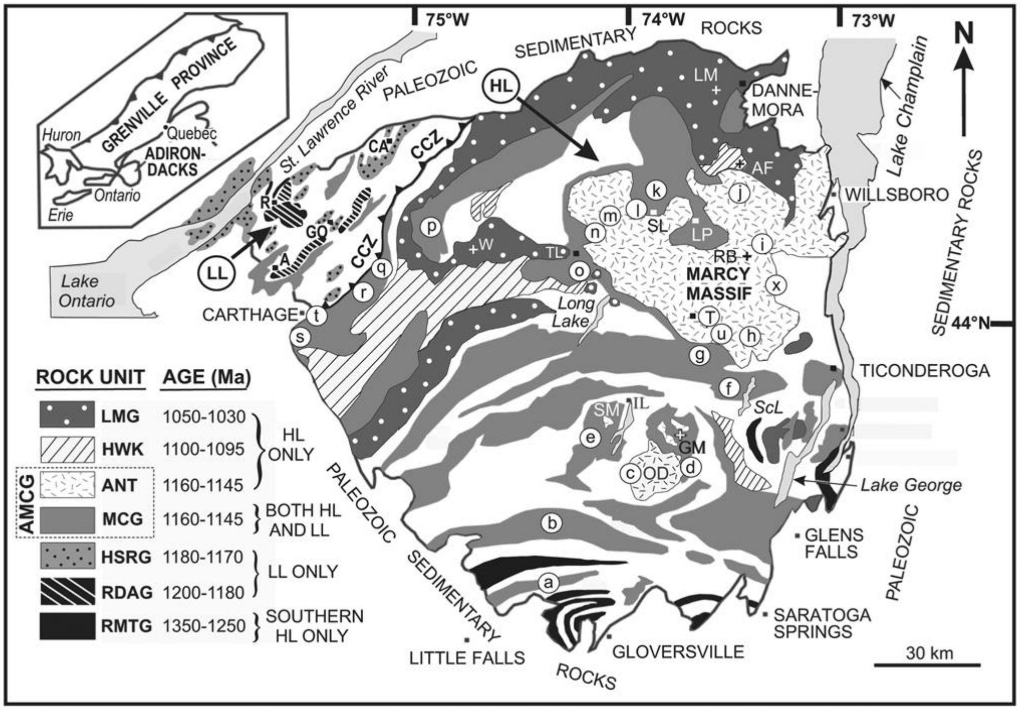

Figure 2. *Generalized geologic and geochronological map of the Adirondacks from McLelland et al. (2004). Ages of meta-igneous rocks shown in legend. Refer to McLelland et al. (2004) for sample locations and unit descriptions.*

As shown in Figure 2, Middle Proterozoic rocks in the western Adirondacks are characterized by mostly charnockites and granites, with lesser amounts of amphibolite and meta-sedimentary rocks (metapelites and calcsilicates). Radiometric dating in the area has largely concentrated on the Lyon Mtn granite (McLelland et al., 2001, 2002b) and surrounding metapelites (Florence et al., 1995). The radiometric dates of zircon fall into two general categories, those associated with the Ottawan phase of the Grenville Orogeny (ca. 1050-1090 Ma) and those associated with earlier anorthosite-mangerite-charnockite-granite (AMCG) magmatism (1145-1160 Ma).

Metapelitic rocks in the western Adirondack Highlands have been the focus of three metamorphic studies (Florence et al., 1995; Darling et al., 2004; Darling, 2013) and pressure-temperature (PT) conditions of >780^{o}C, 6.0 ± 0.5 kbar near Port Leyden, NY, and 830-870^{o}C and 6.0-7.2 kb near Moose River, NY, have been determined. These PT conditions are well into the granulite-facies, but the lower than average pressures reported by Florence et al. (1995) and Darling (2013) suggest mid-crustal burial depths with an elevated geotherm. The metamorphic temperatures determined in the aforementioned studies are considerably higher than those projected by Bohlen et al. (1985) for the western Adirondacks. A summary of Adirondack metamorphic conditions is provided by Darling and Peck (2016), and new evidence of ultrahigh temperature metamorphism (UHT, >900°C) has recently been described in the central Adirondacks by Shinevar et al. (2021), Ferrero et al. (2021), and Metzger et al. (2022).

On this trip, we will visit a number of unusual meta-igneous and meta-sedimentary rock types. The basic itinerary is as follows:

Stop 1.— Ordovician-age spheroidal weathering at the Knox unconformity.

Stop 2.— Port Leyden nelsonite.

Stop 3.— Two-pyroxene amphibolite at Lyons Falls, NY

------------ Lunch --

Stop 4.— Hydrothermal quartz + sillimanite veins and pegmatite at Lyonsdale, NY

Stop 5.— Prismatine locality at Moose River, NY

------------ Head back west on Moose River Road ---------------------

More detailed rock descriptions and interpretations are included under each of the five stops.

FIELD GUIDE AND ROAD LOG

Meeting Point: Parking lot of the Edge Hotel, Route 12, Lyons Falls, NY.

The Edge Hotel is located 40 miles north of Utica on Route 12.

Meeting Point Coordinates: 43.6217°N, 75.3717°W

Meeting Time: 8:30 AM

From the Edge Hotel parking lot, turn right and proceed north on Rt. 12 for 9.0 miles and turn right onto Cannan Rd. Proceed for 0.2 miles and park on the left side next to Roaring Brook.

STOP 1. -- Ordovician spheroidal weathering in Middle Proterozoic gneiss

Parking Coordinates: 43.7427°N, 75.4251°W

Site Coordinates: 43.7425°N, 75.4251°W

Walk down to the exposures of pink feldspathic gneiss along Roaring Brook. Here fractured bedrock contains very good examples of spheroidal weathering preserving between the fractures sets (Figure 3). The spheroidal weathering is characterized by closely spaced (3-4 mm) bands of iron hydroxide, the bands extending into the gneiss a few centimeters from the fracture sets. Microscopically, the bands are characterized by fine-grained iron-hydroxide, calcite, chlorite, and possibly serpentine. Locally, the bands are filled with medium-grained calcite, suggesting open fracture deposition.

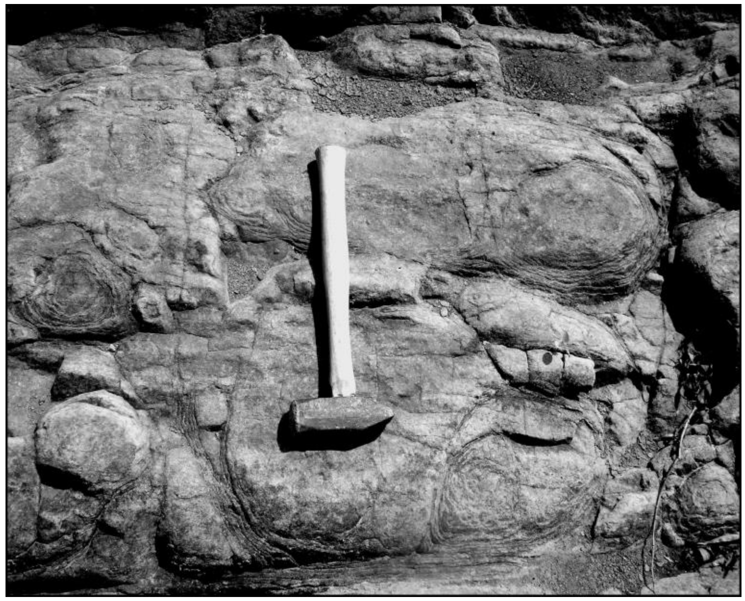

Figure 3. *Vertical view onto surface of Ordovician–age, spheroidal weathering preserved in middle Proterozoic felsic gneiss just below the Knox unconformity at Roaring Brook (Stop 1). Hammer for scale.*

From this location, walk upstream about 30 meters and observe the lowermost red and green sands and pebble conglomerates of the Pamelia Formation (Middle Ordovician) resting directly on top of Middle Proterozoic gneiss. This is the widely known *Knox Unconformity* and represents nearly 600 million years of missing rock record. The unconformity is easily observed in the stream bed. Spheroidal weathering also occurs directly below the nonconformable contact, but is observed only during low water levels (normally late summer). The spheroidal weathering directly below the nonconformity and ~30 meters downstream (location of Fig. 3) are the only locations where it has been observed. Both are located within one vertical meter of the nonconformity. Exposures of felsic gneiss farther downstream, which are a few meters below the projection of the unconformity, show little or no evidence of spheroidal weathering. The proximal relationship between the nonconformity and spheroidal weathering is interpreted as evidence of Ordovician-age chemical weathering. Middle Ordovician time, therefore, was likely tropical or sub-tropical, which is consistent with paleomagnetic studies (Niocaill, et al., 1997).

Because the closely spaced bands of iron hydroxide (in the spheroidal weathered portion of the gneiss) contain chlorite, the rocks must have been buried to "chlorite-grade" depths following middle Ordovician deposition. This is interpreted to have occurred during the late Paleozoic Alleghanian Orogeny (Isachsen et al., 1991). Consequently, the chlorite here cannot be associated with retrograde metamorphism that occurred during exhumation of Middle Proterozoic rocks following the Grenville Orogeny.

From STOP 1, get back on Rt. 12 and head south for 12.1 miles to the village of Port Leyden and turn left at the traffic light. Proceed for 0.2 miles on East Main Street and turn left into the former Port Leyden Elementary School and park. Walk northeastward across the lawn and into the woods about 100 meters and look for a water-filled, mine shaft. Beware of old barbed wire and poison ivy!

<u>STOP 2. – Port Leyden nelsonite</u>

Parking Coordinates: 43.5842°N, 75.3418°W

Site Coordinates: 43.5851°N, 75.3411°W

This is one of two occurrences of nelsonite in New York State (Darling and Florence, 1995). The other occurs near Cheney Pond, 110 km to the northeast, in the High Peaks region of the Adirondacks (Kolker, 1980; 1982). At Port Leyden, the nelsonite occurs as a dike about 3 to 4 meters wide and is traceable for about 30 meters on the surface. The host rock is metapelitic gneiss comprising K-feldspar, quartz, garnet, biotite, sillimanite and spinel.

Nelsonites are unusual igneous rocks comprising apatite and Fe-Ti oxides such as ilmenite, or magnetite and rutile. They are one of the few igneous rocks on Earth that have little or no silicate minerals. The Port Leyden nelsonite has 32-50% magnetite, 8-15% ilmenite, 30-45% apatite, and 5-11% pyrite (Darling and Florence, 1995). Chlorite is present as well and easily observed in hand specimens, whereas garnet, zircon and monazite are best observed in thin section. The pyrite commonly defines planar bands and lenses in what could be described as a foliation.

A characteristic feature of nelsonite is its overall, fine-grained texture compared to other rocks of the anorthosite suite (A. Philpotts, personal communication,) see Figure 4. However, ilmenite in the Port Leyden nelsonite actually forms much larger, oikocrystic grains up to 5 cm across (Figure 5). Although hard to observe in Figure 4, all of the ilmenite shown is actually part of one larger grain. Oikocrysts are *large crystals that enclose smaller crystals to form a poikilitic texture.* At Port Leyden, the ilmenite fills voids between millimeter scale apatite and magnetite grains (Figure 4). Nearly all ilmenite in the Port Leyden nelsonite is oikocrystic and is best observed by reflections from the (0001) parting on freshly broken or cut surfaces (Figure 5). Ilmenite also occurs as oxy-exsolution lamellae in magnetite (Fenner et al., 2018).

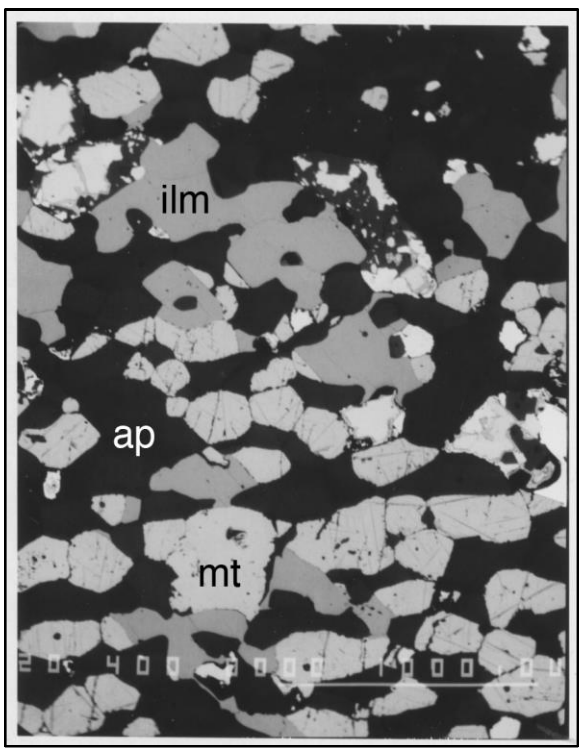

***Figure 4.** Backscattered electron image of the Port Leyden nelsonite. Bar scale at bottom is 1 mm. Note distinctive fine-grained texture. ap = apatite; mt = magnetite; ilm = ilmenite. Brightest grains are pyrite (unlabeled).*

Figure 5. *Cut slab of the Port Leyden nelsonite showing reflections from (0001) parting of single oikocrystic ilmenite grain (center). Slab width is 4 cm.*

It is tempting to infer that the ilmenite oikocrysts have a late igneous origin, but Duchesne (1996) argues against late liquids with a pure ilmenite composition. Although the texture is indicative of late crystallization, the origin of interstitial ilmenite could be the result of post-igneous deformation instead (Duchesne, 1996).

Nelsonites are normally associated with anorthosite-suite rocks (like at Cheney Pond) and there are two proposed theories on their origin. They are believed to form by either magmatic immiscibility (Philpotts, 1967; 1981) or by cumulate processes (Dymek and Owens, 2001; Duchesne and Liégeois, 2015), but in both cases the source rocks are either oxide-apatite gabbro norites or jotunites of the anorthosite suite. Neither of these two rocks occurs within the vicinity of Port Leyden, so the parent rock of the Port Leyden nelsonite has thus far not been identified. It is conceivable that the parent rocks once existed in the Port Leyden area and were eroded away. This is plausible because Philpotts (1981) suggests that nelsonites actually intrude downward in the crust due to their high liquid density (~4.0 gms/cm^3). It is also possible that anorthosite-suite rocks currently located near Carthage, NY (~50 km to the north-northwest; see Buddington, 1939) could be a potential source rock as the Adirondack Lowlands once existed structurally on top of the Adirondack Highlands and slid in a NW direction along the CCMZ late in the history of the Grenville Orogeny (Rivers, 2008). This would require greater than 50 km of horizontal displacement along the CCMZ, however. Lastly, the most likely and nearest source rocks (stratiform anorthosites and jotunites) occur as little as 25 km to the northeast (near Crooked Creek, roughly along strike) in the Number Four 15' quadrangle (Whitney et al., 2002). These could be potential source rocks that were later separated from the nelsonite during regional Ottawan shearing.

Darling et al. (2018) provide radiometric dates on zircon from the Port Leyden nelsonite. Two zircon textures were observed: 1) highly oscillatory-zoned cores with homogeneous rims, and 2) homogeneous grains. The concordant age on oscillatory zoned cores (n = 14) is 1145.5 ± 4.2 Ma, whereas the concordant age on homogeneous grains and rims (n = 16) is 1036.8 ± 3.8 Ma. Interestingly, the oscillatory zoned cores contain 2-10 micrometer size inclusions of K-feldspar and quartz (Darling, 2016), mineral phases generally not associated with oxide-apatite gabbro norites or jotunites. Therefore, the zircon cores are interpreted as inherited xenocrysts and may suggest mixing of felsic melts with that of the nelsonite parent melt (Darling, 2016). The identification of the included phases was based on EDS analysis so it is plausible that the tiny grains in zircon could actually be kokchetavite and cristobalite, the high temperature, metastable polymorphs of K-feldspar and quartz, respectively, typical of nanogranite melt inclusions (Ferrero et al., 2016). The *homogeneous grains and rims* are interpreted to have formed during igneous crystallization of the Port Leyden nelsonite. The 1036.8 ± 3.8 Ma age of igneous crystallization is similar to an obtained zircon age (1032.1 ± 3.8 Ma) on the aforementioned stratiform jotunite / anorthosite body, occurring along strike of the regional foliation, 25 km to the northeast, near Crooked Creek (Whitney et al. 2002). The similarity of ages strongly suggests that regionally extensive, folded jotunite / anorthosite layers mapped in the western and central Adirondacks are a likely parental magma source for the Port Leyden nelsonite.

The inferred Ottawan age of the Port Leyden nelsonite intrusion has some interesting consequences. First, the nelsonite dike cross-cuts the regional foliation of the metapelite host rocks which suggests the foliation predates the intrusion. If so, when was the nelsonite separated from its inferred parent rocks 25 km to the northeast? Also, if the pyrite banding represents a metamorphic foliation, when was it produced? Is the pyrite banding parallel to the foliation in the country rocks? Only an oriented core of the nelsonite would answer this question. Lastly, the nelsonite does contain metamorphic garnet inferred to have formed by the reaction plagioclase + Fe-oxide → garnet (McLelland and Whitney, 1977). This suggests that garnet-producing metamorphic reactions continued after 1036 Ma.

Darling et al. (2018) report that apatite from the Port Leyden nelsonite, contains 1.1 to 2.1 wt.% total rare earth element oxides, which is about 5 to 10 times more than reported REE contents of apatites from some Norwegian nelsonites (Duchesne, 1999). Chondrite normalized plots of REE abundances in apatite and zircon show: 1) LREE enrichment in apatite, 2) a similar (in magnitude) negative Eu anomaly for apatite and zircon, 3) HREE enrichment in zircon, and 4) a positive Ce anomaly in zircon. No systematic difference was recognized in the REE patterns between the aforementioned zircon types. Furthermore, the REE patterns are typical for igneous zircon (Hoskin and Schaltegger 2003). The negative Eu anomalies of both phases suggest significant plagioclase crystallization in the parent magma. Although the Port Leyden nelsonite contains small amounts of metamorphic garnet, the HREE enrichment in zircon indicates no zircon grew in the presence of garnet.

Small lenses (2 cm width) of nelsonite are locally observed in the surrounding metapelitic gneiss. One such lens occurring in rock excavated from the small hydroelectric plant on the Black River at the end of North St. was found to contain small (1 mm), highly fractured sapphires (Darling et al., 2019). Their rich blue color is attributed to the presence of high amounts (up to 0.22 wt.%) of TiO_2, which is not surprising given the composition of the host lens. Although these sapphires are not gem quality, they are the only reported sapphires in New York State, and their presence here demonstrates a previously undescribed geological occurrence in nelsonite.

From STOP 2, go west on East Main Street 0.2 miles and turn right at the traffic light onto on Rt. 12 north. Proceed 2.1 miles and turn right onto Franklin Street Extension. After 350 feet, turn left onto River Road. Proceed 0.3 miles and turn right onto Laura Street. Proceed for 0.2 miles (you'll cross over the Black River) and turn left onto Lyons Falls Rd. Process 0.3 miles (you'll cross over the Moose River now) and turn left onto the parking area. Walk down the path to Lyons Falls.

STOP 3. – Two-pyroxene amphibolite at Lyons Falls

Parking Coordinates: 43.6186°N, 75.3568°W

Site Coordinates: 43.6185°N, 75.3574°W

Lyons Falls drops about 63 feet here and has been harnessed as a power source since the mid-1800's. The base of the falls served as an initial settlement for French explorers of the "Castorland Company," in June of 1794 (Hough, 1860).

Lyons Falls occurs here because of a ~100 meter wide band of highly resistant amphibolite gneiss that strikes northeast, normal to the course of the Black River. The amphibolite here is strongly lineated with discontinuous, thin, plagioclase-rich bands. Foliation is poorly developed. D. Valentino (personal communication) describes it as an L-tectonite. The unit has a sharp contact with feldspar-quartz gneiss to the south (observed at the foot of the upstream board dam) but the north contact is not exposed here.

Petrologically, the unit is best described as a medium-grained, two-pyroxene amphibolite. Plagioclase, hornblende, clinopyroxene, and orthopyroxene are the dominant mineral phases. A red-brown (presumably Ti-rich) biotite and opaque magnetite occurs as well (Figure 6). Quartz has been observed but is uncommon. Unlike the central and eastern Adirondacks, the amphibolite at Lyons Falls contains no garnet, despite the fact they are compositionally similar.

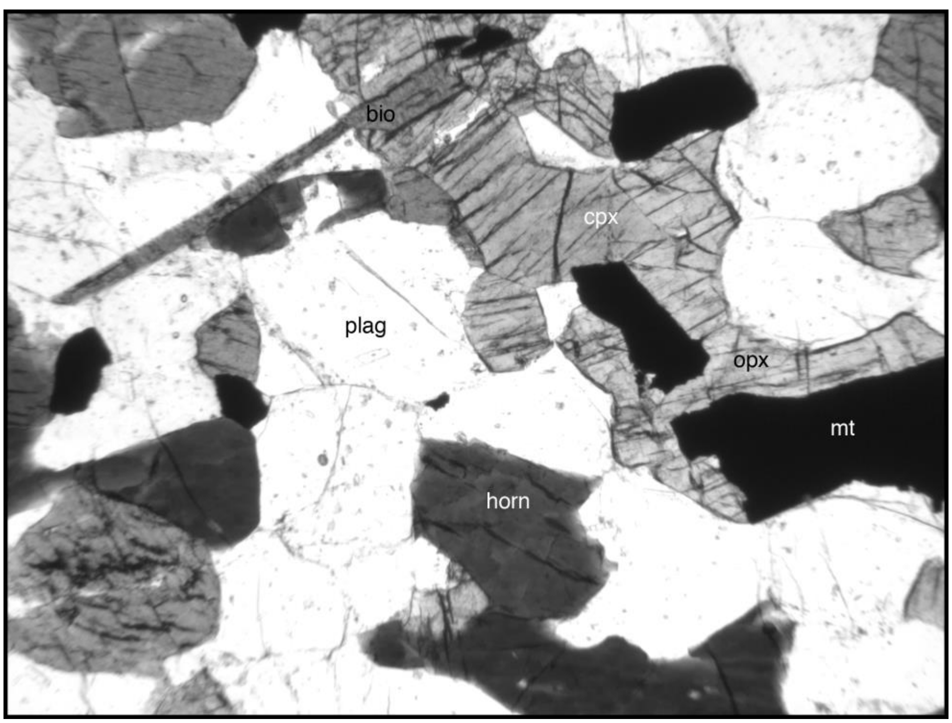

Figure 6. *Photomicrograph of amphibolite at Lyons Falls. Note granoblastic texture and medium grain size. (plag = plagioclase; horn = hornblende; cpx = clinopyroxene; opx = orthopyroxene; bio = biotite; mt = magnetite) Note absence of garnet (see text for discussion). Field of view = 2.5 millimeters width.*

The absence of garnet in amphibolites from the western Adirondacks has been known for a long time (Buddington, 1939; de Waard, 1967 and references therein). Its absence is due to lower metamorphic pressures in this region of the Adirondacks as compared to the central and eastern Adirondack Highlands.

Figure 7 shows that the amphibolite of Lyons Falls is located to the west of the garnet + clinopyroxene isograd. In mafic rock compositions, this isograd is based on the reaction:

orthopyroxene + anorthite = clinopyroxene + garnet + quartz (1),

where garnet is present on the higher pressure side of the reaction. The famous garnet amphibolites (e.g. Gore Mtn.), so common in the central and eastern Adirondacks could not form in the western Adirondacks simply because the rocks were not buried deep enough.

The garnet + clinopyroxene isograd across the central Adirondacks has received little attention since first mapped by de Waard (1967). Why the isograd makes two nearly right angle and opposite turns as it sinuates from south to north is unknown. Also unknown is the inferred temperatures and pressures along the isograd and how this compares to other thermobarometric estimates from non-mafic lithologies. The southern terminus of the isograd should have been affected by both the Moose River Plain and the Piseco Lake shear zones (Gates et al., 2004) unless shearing pre-dates garnet formation in mafic lithologies.

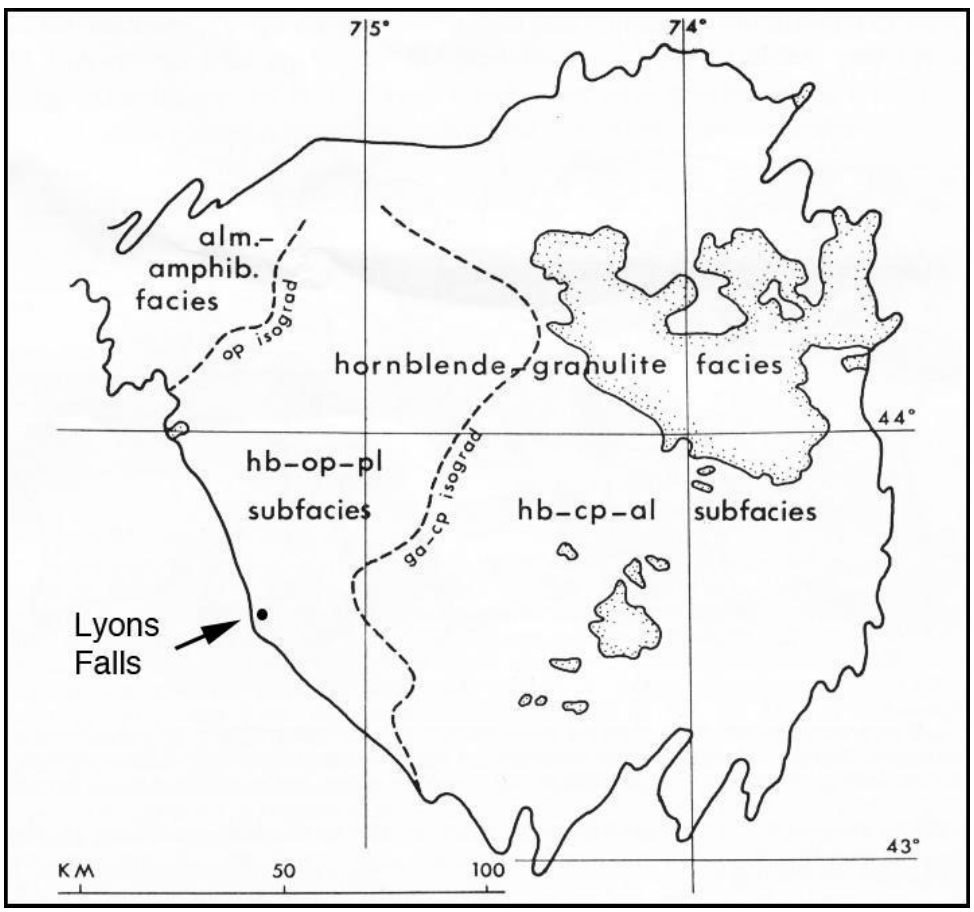

Figure 7. *Simplified metamorphic isograd map for the Adirondacks modified from de Waard (1967). Note the amphibolite at Lyons Falls is located west of the garnet + clinopyroxene isograd.*

Also occurring at Lyons Falls are some meter-scale potholes carved into the amphibolite. These are best observed during low-water levels and closer to the waterfall.

For our lunch stop, we'll drive to Stewart's on Rt. 12. From STOP 3, go back 0.3 miles to Laura St. (while again crossing the Moose River) and turn right, proceed 0.2 miles (while again crossing the Black River) and turn right onto Franklin St. Continue for 0.5 miles as Franklin St. turns into Center St. while passing through the village of Lyons Falls. At the end of Center St. turn left onto McAlpine St. and proceed for 0.2 miles and turn left at the flashing light onto Cherry St. After 0.2 miles turn right into the Stewart's Shop.

Parking Coordinates for Stewart's Shop: 43.6164°N, 75.3682°W

From the Stewart's Shop, get on Rt. 12 and head southeast for 0.5 miles and turn left onto Franklin St. Ext. After 350 feet, turn left onto River Road. Proceed 0.3 miles and turn right onto Laura Street. Proceed 0.8 miles and turn left onto Lyonsdale Rd. Continue for 2.4 miles. After passing the Twin Rivers Paper Co. (on the left), the road turns into the Marmon Rd. Continue 0.3 miles and turn left onto Hunkins Rd. Continue 1.1 miles and turn left onto Fowlersville Rd. Continue 2.5 miles (while again crossing the Moose River) and turn left onto Fowler Rd. Continue 1.4 miles and turn left onto Lowdale Rd. Continue 0.8 miles and park at the end of Lowdale Rd. Follow the trail south of Lowdale Rd. to the Moose River bedrock exposures just below the former bridge location.

STOP 4. – Undeformed pegmatite and hydrothermal sillimanite + quartz veins in Lyon Mtn. granite.

Parking Coordinates: 43.6194°N, 75.3030°W

Site Coordinates: 43.6196°N, 75.3035°W

This stop demonstrates important igneous and hydrothermal features of Lyon Mtn. granite along the Moose River. These exposures were first recognized by Buddington and Leonard (1962, p. 93) and studied extensively by McLelland et al. (2001, 2002a, 2002b) and Selleck et al. (2004). Their overall interpretation is that Lyon Mtn. granite experienced contemporaneous intrusion and hydrothermal alteration at about 1035 Ma. The hydrothermal activity leached large cations (K^+, Na^+) from the granite but left behind Al^{3+} and Si^{4+} to form quartz-sillimanite veins. Early vein sets were ductilely deformed (due to magmatic flow or tectonic shear) and younger veins sets formed afterward.

These rocks were included in a large unit of mapped metapelites in Figure 1 of Florence et al. (1995) but the composition and textural features are more consistent with altered igneous rocks. Some of the country rocks into which the Lyon Mtn. granite intruded are indeed metapelites, and numerous exposures of sillimanite + garnet + hercynite + quartz + K-feldspar gneiss occur in the area.

Looking west from the former bridge on the northern side of the Moose River, one can observe an undeformed pegmatite cutting hydrothermally altered Lyon Mtn. granite. The pegmatite is shown in Figure 8 and is compositionally zoned with an uncommon magnetite-rich core. McLelland et al. (2001) dated well-zoned igneous zircons from this pegmatite at 1034 ± 10 Ma. Because the pegmatite is undeformed, the 1034 Ma zircon age has been interpreted as the terminus of Ottawan deformation in this part of Adirondacks (Orrell and McLelland, 1996).

Figure 8. *Undeformed pegmatite just west of Lyonsdale bridge. Taken from Figure 4a of McLelland et al. (2001). See text for discussion.*

The bedrock exposed farther downstream show excellent examples of quartz + sillimanite veins hosted by Lyon Mtn. granite (see Figure 9). These veins occur in two prominent orientations, N20E and N50E (McLelland et al., 2002a). At other locations in the area (e.g. Ager's Falls), the quartz-sillimanite veins are nodular in shape and strongly deformed (McLelland et al., 2002a,b). Figure 10 shows the rock texture in thin-section. Note the presence of tartan-twinned microcline; this is in contrast to many nearby metagranites that are characterized by a hypersolvus quartz-mesoperthite mineralogy.

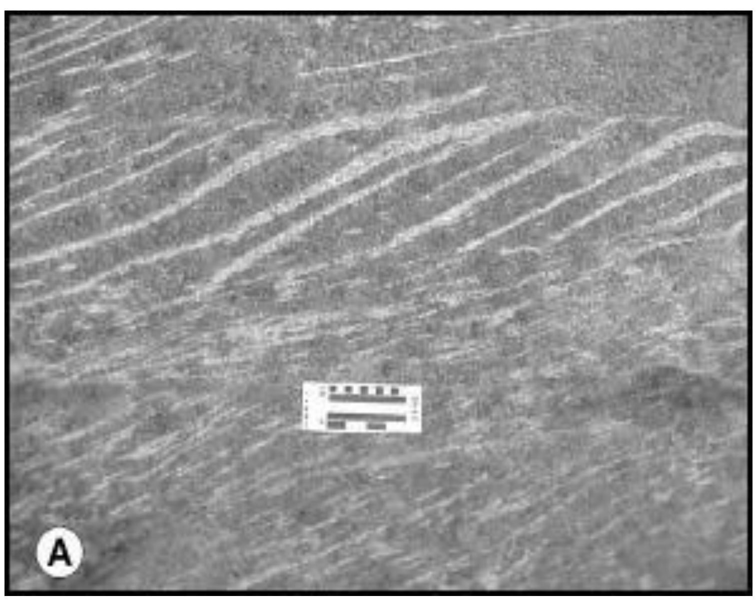

Figure 9. *Hydrothermal quartz + sillimanite veins in Lyon Mtn. granite at Stop 4. Taken from Figure 2a of Selleck et al. (2004).*

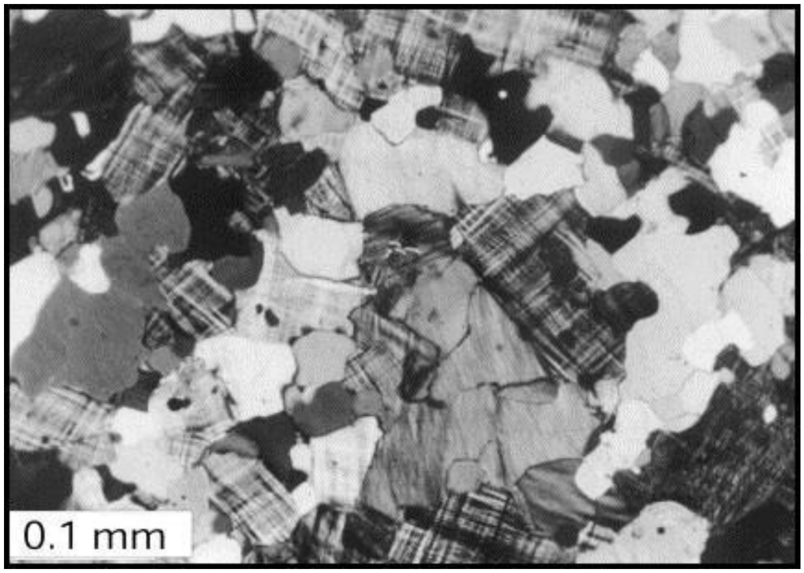

Figure 10. *Photomicrograph of Lyon Mtn. granite. Note little or no grain shape fabric of quartz and microcline. From Figure 11 of McLelland et al. (2002b).*

The hydrothermally altered Lyon Mtn. granite here at Lyonsdale is not the only occurrence of this unusual Adirondack rock. Identical sillimanite + quartz vein-hosted granite occurs in a number of locations along strike (to the SW) until ending at Johnson Dam, on the Black River 4.6 kilometers to the southwest.

From STOP 4, go back along Lowdale and Fowler Roads (2.2 miles) to Fowlersville Rd and turn right. Proceed south for 2.8 miles to Penney Settlement Rd. and turn left. Proceed 0.5 miles and turn right onto North-South Rd. Proceed 2.0 miles and turn left onto Moose River Rd. Follow Moose River Rd. for 7.1 miles, turn left and park in the sandy area close to the Moose River. From this point, follow the all-terrain vehicle path downstream for about 200 meters. The path passes through the stone foundations of the former Moose River tannery and then follows rapids as the Moose River flows southwest. Here, the river cuts through northwest-dipping, calc-silicate gneisses and quartzites. Stay high on the river bank until the rapids disappear. The path will descend and cross a small, wet, muddy creek bed. Afterwards, the Moose River pools and turns north and the first outcrops on the west side of the river are the prismatine-bearing rocks.

STOP 5. – Moose River Prismatine locality.

Parking Coordinates: 43.6080°N, 75.1624°W

Site Coordinates: 43.6101°N, 75.1669°W

Please exercise caution while walking among the river boulders and talus at the base of the outcrops. Also, please DO NOT USE HAMMERS at this stop and refrain from collecting prismatine specimens unless you're planning to study them scientifically; a future petrologist or mineralogist will be grateful someday. Moreover, none of the prismatine grains are gem-quality, they are too dark and too fractured!

Prismatine, the boron-rich endmember of the kornerupine solid solution (Grew et al., 1996; ideally $Mg_3Al_6Si_4BO_{21}(OH)$) occurs in metapelitic and quartzitic rocks along the Moose River. Kornerupine-group minerals are generally rare, having been described from nine localities in the Grenville Province (Grew, 1996; Darling et al., 2004; Korhonen and Stout, 2005) including two in the Adirondacks (Farrar and Babcock, 1993; Farrar, 1995; Darling et al., 2004).

Prismatine is identified here by X-ray diffraction, electron microprobe analyses, SIMS analyses and laser Raman microspectroscopy (Bailey et al., 2019). However, it was first suspected by observing its interfacial angles of about 80 and 100 degrees between the {110} prism faces, values in between those of an amphibole (56° and 124°) and a pyroxene (87° and 93°).

Along the Moose River, prismatine occurs at two locations (separated by about 400 meters, Figure 11) within a unit of heterogeneous metasedimentary rocks (Figure 11, unit BL) mapped by Whitney et al. (2002). This unit comprises mostly quartzite and biotite-quartz-plagioclase gneiss with lesser amounts of calc-silicate rocks, and minor amphibolite, quartzofeldspathic gneiss, and calcite marble (Whitney et al., 2002). These rocks are interlayered with other metasedimentary and meta-igneous rocks (Figure 11). These units occur in a complex, southeast-verging overturned synform bordered on the northwest by a tabular, northwest-dipping body of charnockitic gneiss (CG) several kilometers thick, and on the southeast by a domical body of batholithic proportions consisting of relatively leucocratic CG (Whitney et al, 2002). Although the granitic and charnockitic rocks have not been dated, they are lithologically and geochemically similar to felsic rocks of the ca. 1150 Ma anorthosite-mangerite-charnockite-granite (AMCG) suite found throughout much of the Adirondack Highlands (McLelland et al., 2001; Whitney et al., 2002).

In addition to prismatine-bearing assemblages, the surrounding rocks contain metapelitic assemblages of a) cordierite + spinel + sillimanite + garnet + plagioclase + quartz + ilmenite + rutile +/- biotite, b) cordierite + orthopyroxene + biotite + K-feldspar + quartz, and c) orthopyroxene + plagioclase + K-feldspar + quartz +/- biotite, +/- garnet (Darling et al., 2004).

The feature of geologic interest at STOP 5 are the exceptionally well-developed prismatine crystals in coarse-grained, feldspathic lenses. Here, prismatine crystals form dark greenish-black, euhedral, elongated grains (up to 10 cm in length). Ed Grew (personal communication) indicates that only the prismatine crystals from the Larsemann Hills, Antarctica (Grew and Carson, 2007) are comparable in length to those at Moose River. The prismatine commonly displays radiating patterns in feldspathic lenses one to three cm thick (Figure 12A).

The prismatine crystals *appear* to have grown only within the plane of the foliation. However, upon closer examination, the prismatine grains are seen to be arranged randomly, but the longest and best-developed crystals formed parallel to the foliation plane. Because of this, Darling et al. (2004) inferred that nondeviatoric pressure conditions prevailed locally during prismatine formation. It should also be noted that a number of prismatine-bearing feldspathic lenses are located adjacent to fine-grained tourmaline + plagioclase + biotite–rich zones near the north end of the exposed rocks. In these locations, the prismatine-bearing feldspathic lenses texturally embay, cross-cut earlier foliation, and appear to form at the expense of the tourmaline-bearing zones (Figure 12B). The embayed country rocks, coarser grain size, and the random arrangement of the prismatine crystals led Darling et al. (2004) to interpret the feldspathic lenses and the prismatine found in them to be of anatectic origin.

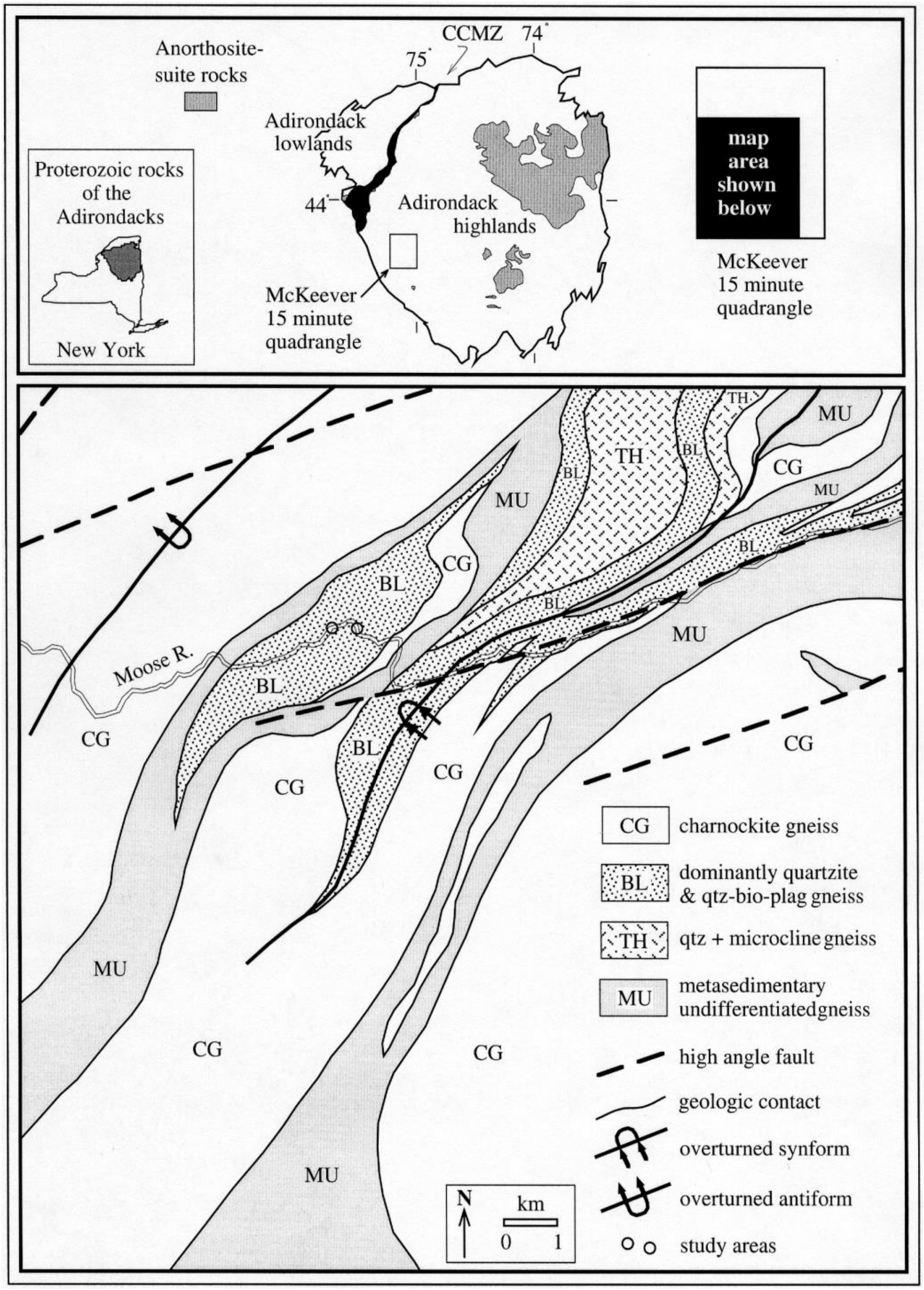

Figure. 11. *Map showing location of prismatine-bearing metapelites and quartzites (open circles on Moose River) and surrounding bedrock geology. Stop 5 is the easternmost open circle. Geologic map units, structures, and relations illustrated are from Whitney et al. (2002). Taken from Darling et al. (2004). Bio—biotite; CCMZ—Carthage-Colton mylonite zone; pl—plagioclase; qtz—quartz.*

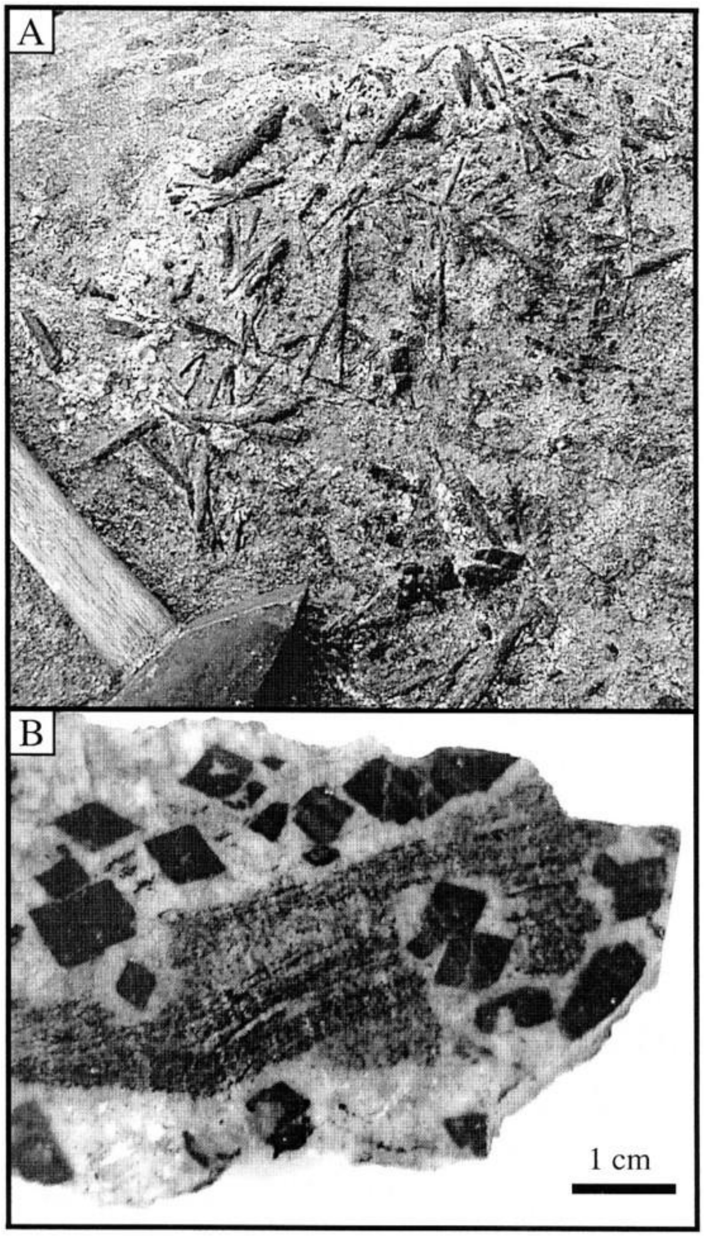

Figure 12. *(A) Prismatine crystals (black) in coarse-grained feldspathic lens, taken parallel to plane of lens. Hammer for scale. (B) Euhedral prismatine (black) in coarse-grained feldspathic lens embaying fine-grained, foliated tourmaline + biotite + plagioclase–rich zones. Taken from: Darling et al. (2004).*

Plagioclase, K-feldspar, minor quartz and rutile are the most common phases associated with prismatine, but biotite, cordierite, garnet and rarely sillimanite occur locally as well. The prismatine contains 0.73 to 0.79 formula units of B (out of 1.0) and has Mg / Mg + Fe between 0.70 and 0.73 (Table 3 of Darling et al., 2004). After determining the associated mineral compositions, Darling et al. (2004) proposed the following prismatine-forming reaction:

Tourmaline + sillimanite + biotite + cordierite → prismatine + rutile + melt (2).

Reaction 2 is similar to a number of proposed prismatine-forming reactions from other granulite terranes (Grew, 1996), including those in sapphirine-free rocks found in the Reading Prong, New Jersey (Young, 1995), and in Waldheim, Germany (Grew, 1989). In those cases, garnet rather than cordierite was a proposed reactant. Because prismatine is inferred to form as a direct result of anatectic melting, it is considered a peritectic phase.

Metamorphic temperatures and pressures are difficult to estimate from prismatine-bearing mineral assemblages as little is known about the stability of boron-rich kornerupine at pressures less than 10 kb (Schreyer and Werding, 1997). However, the prismatine occurs in proximity to low-variance metapelitic assemblages in the surrounding rocks. Specifically, thermobarometry calculations from net transfer and exchange equilibria record temperatures and pressures of 850° ± 20°C and 6.6 ± 0.6 kilobars for orthopyroxene + garnet assemblages and 675° ± 50°C and 5.0 ± 0.6 kilobars for cordierite + garnet + sillimanite + quartz assemblages (Darling et al., 2004). The former assemblage is interpreted to have formed during partial melting whereas the latter assemblage is interpreted to have formed on the early retrograde metamorphic path (Darling et al., 2004). The ~850°C temperatures derived from the orthopyroxene + garnet assemblage are reasonable for partial melting conditions. Although the cordierite + garnet + sillimanite + quartz assemblage occurs at STOP 5, it and the orthopyroxene + garnet assemblage are better developed farther downstream at the second prismatine location (the westernmost open circle in Figure 11). These exposures can be reached by following the footpath on the south bank of the Moose River for a distance of about 400 meters.

Because many of the prismatine-bearing feldspathic lenses are saturated in both quartz and rutile, Storm and Spear (2009) intensely studied the prismatine-bearing lenses as part of a natural test of the titanium-in-quartz geothermometer of Wark and Watson (2006). Storm and Spear (2009) determined a wide range of metamorphic temperatures, specifically from 630 + 63 / -86 to 879 ± 8 °C, but most determinations fell between 700°C and 880°C (see Figure 8a of Storm and Spear, 2009). This is in good agreement with metamorphic temperatures determined by the aforementioned methods (Darling et al., 2004). Storm and Spear (2009) also provide convincing textural evidence that prismatine was locally replaced by leucosomatic quartz, most likely during melting of prismatine. Interestingly, it was the leucosomatic quartz that yielded the highest Ti-in-quartz temperatures (800-880°C; Figure 8a of Storm and Spear, 2009).

The age of partial melting is unknown at this time but is likely associated with either intrusion of the AMCG suite at ~ 1160-1145 Ma, or burial associated with the Ottawan phase of the Grenville Orogenic cycle and the associated intrusion of Lyon Mtn. granite at ~ 1050-1030 Ma (McLelland et al., 2010).

Lastly, and perhaps most interestingly, prismatine from the Waldheim granulite, in Germany, was recently found to contain tiny mineral inclusions of coesite, the high pressure polymorph of silica. Moreover, almandine garnet from these same rocks were found to contain the first ever described inclusions of stishovite, the super high pressure polymorph of silica (Thomas et al., 2022)!

REFERENCES CITED

Bailey, D.G., Lupulescu, M.V., Darling, R.S., Singer, J.W. and Chamberlain, S.C., 2019, A review of boron-bearing minerals (excluding tourmaline) in the Adirondack region of New York State: Minerals, 9(10), p. 644. doi.org/10.3390/min9100644

Bohlen, S.R., Valley, J.W., and Essene, E.J., 1985, Metamorphism in the Adirondacks. I. Petrology, pressure, and temperature: Journal of Petrology, v. 26, pp. 971–992.

Buddington, A.F., 1939, Adirondack igneous rocks and their metamorphism: Geol. Soc. Amer. Mem. 7, 295 p.

Buddington, A.F. and Leonard, B.F., 1962. Regional geology of the St. Lawrence county magnetite district, northwest Adirondacks, New York: U.S. Geological Survey Professional Paper 376, US Govt. Print. Off., 145 p.

Darling, R.S., and Florence, F.P., 1995, Apatite light rare earth chemistry of the Port Leyden nelsonite, Adirondack Highlands, NY: Implications for the origin of nelsonite in anorthosite suite rocks: Economic Geology, v. 90, p. 964-968.

Darling, R.S., Florence, F.P., Lester, G.W., Whitney, P.R., 2004. Petrogenesis of prismatine-bearing metapelitic gneiss along the Moose River, west-central Adirondacks, New York. In: Tollo, R.P., Corriveau, L., McLelland, J., Bartholomew, M.J. (Eds.), Proterozoic Tectonic Evolution of the Grenville Orogen in North America, Geological Society of America Memoir 197, Boulder, CO, pp. 325–336.

Darling, R.S. and Peck, W.H., 2016, Metamorphic conditions of Adirondack rocks: Adirondack Journal of Environmental Studies, 21(1), p.7.

Darling, R.S., 2016, Felsic mineral inclusions in zircon from the Port Leyden nelsonite, Western Adirondacks: A product of magma mixing?: Geological Society of America, *Abstracts with Programs.* v. 48, n. 2. doi: 10.1130/abs/2016NE-272695

Darling, R.S., Lupulescu, M.V., and Chiarenzelli, J.R., 2018, Rare earth element composition of apatite, zircon, and monazite, and U-Pb zircon age of, the Port Leyden Nelsonite, Western Adirondack highlands, NY: Geological Society of America, paper No. 40-3. doi: 10.1130/abs/2018NE-310477

Darling, R.S., Gordon, J.L. and Loew, E.R., 2019. Microscopic blue sapphire in nelsonite from the western Adirondack Mountains of New York State, USA. Minerals, 9(10), p.633.

Duchesne, J.C., 1996. Liquid ilmenite or liquidus ilmenite: a comment on the nature of ilmenite vein deposits. In: Demaiffe D. (ed) Petrology and geochemistry of magmatic suites of rocks in the continental and oceanic crusts. A volume dedicated to Professor Jean Michot, Université Libre de Bruxelles, Royal Museum for Central Africa (Tezvuren), p. 73-82.

Duchesne, J.C., 1999, Fe-Ti deposits in Rogaland anorthosites (South Norway): geochemical characteristics and problems of interpretation: Mineralium Deposita, 34(2), p. 182-198.

Duchesne, J-C., and Jean-Paul Liégeois, J-P., 2015, The origin of nelsonite and high-Zr ferrodiorite associated with Proterozoic anorthosite: Ore Geology Reviews, v. 71, p. 40-56. doi.org/10.1016/j.oregeorev.2015.05.005.

Dymek, R.F., and Owens, B.E, 2001, Petrogenesis of apatite-rich rocks (nelsonites and oxide-apatite gabbronorites) associated with massif anorthosites: Economic Geology v. 96, p. 797-815

Farrar, S.S., 1995, Mg-Al-B rich facies associated with the Moon Mountain metanorthosite sill, southeastern Adirondacks, NY: Geological Society of America Abstracts with Programs, v. 27, no. 1, p. 42–43.

Farrar, S.S., and Babcock, L.G., 1993, A sapphirine + kornerupine–bearing hornblende spinel periodotite associated with an Adirondack anorthosite sill: Geological Society of America Abstracts with Programs, v. 25, no. 6, p. A265.

Fenner, E., Balzani, P., Andersen, A.K., and Singer, J., 2018, Origin of the Port Leyden nelsonite: New insights from oxide chemistry, petrography and geophysics: Geological Society of America, *Abstracts with Programs.* Vol. 50, No. 2. doi: 10.1130/abs/2018NE-310986

Ferrero, S., Ziemann, M.A., Angel, R.J., O'Brien, P.J. and Wunder, B., 2016. Kumdykolite, kokchetavite, and cristobalite crystallized in nanogranites from felsic granulites, Orlica-Snieznik Dome (Bohemian Massif): Not evidence for ultrahigh-pressure conditions. *Contributions to Mineralogy and Petrology*, *171*(1), pp.1-12.

Ferrero, S., Wannhoff, I., Laurent, O., Yakymchuk, C., Darling, R., Wunder, B., Borghini, A. and O'Brien, P.J., 2021, Embryos of TTGs in Gore Mountain garnet megacrysts from water-fluxed melting of the lower crust: *Earth and Planetary Science Letters, 569*, p.117058.

Florence, F.P., Darling, R.S., and Orrell, S.E., 1995, Moderate pressure metamorphism and anatexis due to anorthosite intrusion, western Adirondack Highlands, New York: Contributions to Mineralogy and Petrology, v. 121, p. 424–436.

Gates, A.E., Valentino, D.W., Chiarenzelli, J.R., Solar, G.S. and Hamilton, M.A., 2004, Exhumed Himalayan-type syntaxis in the Grenville orogen, northeastern Laurentia: Journal of Geodynamics, 37(3-5), p. 337-359.

Grew, E.S., 1989, A second occurrence of kornerupine in Waldheim, Saxony, German Democratic Republic: Zeitschrift fur Geologische Wissenschaften, Berlin, v. 17. p. 67–76.

Grew, E.S., 1996, Borosilicates (exclusive of tourmaline) and boron in rockforming minerals in metamorphic environments, in Grew, E.S., and Anovitz, L.M., eds., Boron mineralogy, petrology and geochemistry: Washington, D.C., Mineralogical Society of America, Reviews in Mineralogy, v. 33, p. 387–480.

Grew, E.S., Cooper, M.A., and Hawthorne, F.C., 1996, Prismatine: Revalidation for boron-rich compositions in the kornerupine group: Mineralogical Magazine, v. 60, p. 483–491

Grew, E.S. and Carson, C., 2007, A treasure trove of minerals discovered in the Larsemann Hills: Australian Antarctic Magazine, no. 13, p. 18-19.

Hoskin, P.W. and Schaltegger, U., 2003, The composition of zircon and igneous and metamorphic petrogenesis: Reviews in Mineralogy and Geochemistry, 53(1), p .27-62.

Hough, F.B., 1860, A history of Lewis County, in the state of New York, from the beginning of its settlement to the present time: Munsell & Rowland, Albany NY, 319 p.

Isachsen, Y.W., Landing, E., Lauber, J.M., Rickard, L.V., and Rogers, W.B., 1991, Geology of New York: A Simplified Account, New York State Museum, Educational Leaflet No. 28, 284 p.

Kolker, Allan, 1980, Petrology, geochemistry and occurrence of iron-titanium oxide and apatite (nelsonite) rocks: MS Thesis, Univ. of Massachusetts, 156p.

Kolker, A., 1982, Mineralogy and geochemistry of Fe-Ti oxide and apatite (nelsonite) deposits and evaluation of the liquid immiscibility hypothesis: Economic Geology, v. 77, p. 1146-1158.

Korhonen, F.J., and Stout, J.H., 2005, Borosilicate- and phengite-bearing veins from the Grenville Province of Labrador: evidence for rapid uplift: Journal of Metamorphic Geology, v. 23, p. 297–311.

McLelland, J.M. and Whitney, P.R., 1977, The origin of garnet in the anorthosite-charnockite suite of the Adirondacks: Contributions to Mineralogy and Petrology, 60(2), p. 161-181.

McLelland, J.M., Hamilton, M.A, Selleck, B.W., McLelland, Jo. M, and Walker, D., 2001, Zircon U-Pb geochronology of the Ottawan orogeny, Adirondack Highlands, New York; Regional and tectonic implications: Precambrian Research, v. 109, p. 39-72.

McLelland, J., Goldstein, A., Cunningham, B., Olson, C., and Orrell, S., 2002a, Structural evolution of a quartz-sillimanite vein and nodule complex in a late- to post-tectonic leucogranite, western Adirondack Highlands, New York: Journal of Structural Geology, v. 24, p. 1157–1170.

McLelland, J., Morrison, J., Selleck, B., Cunningham, B., Olson, C., and Schmidt, K., 2002b, Hydrothermal alteration of late- to post-tectonic Lyon Mt. Granitic Gneiss, Adirondack Highlands, New York: Origin of quartz sillimanite segregations, quartz-albite lithologies, and associated Kiruna type low-Ti Fe-oxide deposits: Journal of Metamorphic Geology, v. 20, p. 175–190.

McLelland, J.M., Bickford, M.E., Hill, B.M., Clechenko, C.C., Valley, J.W., and Hamilton, M.A., 2004, Direct dating of Adirondack massif anorthosite by U-Pb SHRIMP analysis of igneous zircon: implications for AMCG complexes: Geological Society of America Bulletin, v. 116, p. 1299-1317.

McLelland, J.M., Selleck, B.W., and Bickford, M.E., 2010, Review of the Proterozoic evolution of the Grenville Province, its Adirondack outlier, and the Mesoproterozoic inliers of the Appalachians, *in* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds., From Rodinia to Pangea: Lithotectonic Record of the Appalachian region: Geological Society of America Memoir 206, p. 21-49.

Metzger, E.P., Leech, M.L., Davis, M.W., Reeder, J.V., Swanson, B.A. and Waring, H.V., 2022, Ultrahigh-temperature granulite-facies metamorphism and exhumation of deep crust in a migmatite dome during late-to post-orogenic collapse and extension in the central Adirondack Highlands (New York, USA): Geosphere, 18(1), p. 261-297.

Niocaill, C.N., van der Pluijm, B.A., and Van der Voo, R., 1997, Ordovician paleogeography and the evolution of the Iapetus ocean: Geology, v. 25; no. 2; p. 159–162.

Orrell, S., and McLelland, J., 1996, New single grain zircon and monazite U-Pb ages for Lyon Mt. granite gneiss, western Adirondack Highlands, and the end of the Ottawan orogeny. Geol. Soc. Am. Abs. Prog. 28, p. 88.

Philpotts, A.R., 1967, Origin of certain iron-titanium oxide and apatite rocks: Economic Geology, v. 62, p. 303-315.

Philpotts, A.R., 1981, A model for the generation of massif-type anorthosites: Canadian Mineralogist, v. 19, p. 233-253.

Rivers, T., 2008, Assembly and preservation of upper, middle, and lower orogenic crust in the Grenville Province – Implications for the evolution of large, hot, long duration orogens: Precambrian Research, v. 167, p. 237-259.

Schreyer, W., and Werding, G., 1997, High-pressure behaviour of selected boron minerals and the question of boron distribution between fluids and rocks: Lithos, v. 41, p. 251-266.

Selleck B.W., McLelland J.M., Hamilton M.A. 2004, Magmatic-hydrothermal leaching and origin of late- to post-tectonic quartz-rich rocks, Adirondack Highlands, New York, in Tollo R.P., et al. eds., Proterozoic tectonic evolution of the Grenville orogen in North America: Geological Society of America Memoir 197, p. 379–390.

Shinevar, W.J., Jagoutz, O. and VanTongeren, J.A., 2021, Gore Mountain garnet amphibolite records UHT conditions: Implications for the rheology of the lower continental crust during orogenesis: Journal of Petrology, 62(4), p.egab007

Storm, L.C. and Spear, F.S., 2009, Application of the titanium-in-quartz thermometer to pelitic migmatites from the Adirondack Highlands, New York: Journal of Metamorphic Geol., v. 27, p. 479–494.

Thomas, R.; Davidson, P.; Rericha, A.; Recknagel, U., 2022, Discovery of Stishovite in the Prismatine-Bearing Granulite from Waldheim, Germany: A Possible Role of Supercritical Fluids of Ultrahigh-Pressure Origin: Geosciences, 12, p. 196. doi.org/10.3390/geosciences12050196

Van Diver, B.B., 1985, Roadside Geology of New York: Mountain Press, Missoula, MT, 411 p.

de Waard, D., 1967, The occurrence of garnet in granulite-facies terrane of the Adirondack Highlands and elsewhere, an amplification and a reply: Journal of petrology, v.8, p. 213-232.

Wark, D. and Watson, E.B., 2006, TitaniQ; a titanium–in–quartz geothermometer: Contributions to Mineralogy and Petrology, v. 152, p. 743–754.

Whitney, P.R., Fakundiny, R.F., and Isachsen, Y.W., 2002, Bedrock geology of the Fulton Chain-of-Lakes area, west-central Adirondack Mountains, New York: Albany, New York State Museum Map and Chart 44, 123 p. with map.

Young, D.A., 1995, Kornerupine group minerals in Grenville granulite facies paragneiss, Reading Prong, New Jersey: Canadian Mineralogist, v. 33, p. 1255–1262.

Trip A-3

THE PISECO LAKE SHEAR ZONE: A SHAWINIGAN CRYPTIC SUTURE IN THE SOUTHERN ADIRONDACK MOUNTAINS, NEW YORK

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ABSTRACT

This field trip was first run in 2005 for NYSGA and was reorganized for the 2018 NYSGA-NEIGC joint field conference to show progress in our understanding of the complex geology of the southern Adirondacks. However, in 2018 the trip canceled just prior to the conference due to the sudden tragic loss of Dr. Brian Hough that occurred a few days before the conference. Brian was a colleague of the lead author. Four years later the trip is finally being run again, and it is dedicated to the memory of Brian. Although his tenure at SUNY Oswego was not long, and was tragically cut-short, the impact that he had on many student and colleagues will last a lifetime.

Highly deformed Piseco granitic gneisses occur in an arching east-west transpressional ductile shear zone (Piseco Lake shear zone) that spans the width of the exposed southern Adirondacks. The highly deformed granitic gneisses have restricted silica content, are metaluminous, alkali-calcic to calc-alkalic, continental arc trace element signatures. These granitic rocks intruded supracrustal gneisses resulting in extensive Shawinigan partial melting. The Piseco Lake shear zone (20-30 km wide) formed in this belt of granitic rocks and correlate with a pronounced arcuate-shaped high magnetic anomaly. The magnetic anomaly extends well beyond the exposed Adirondack basement window.

The shear zone is 20-30 km wide and is believed to be the location of a cryptic suture because it occurs between the Adirondack Highlands (underlain primarily by anorthosite and related granitic rocks, AMCG suite) and the Southern Adirondack Terrane (underlain by calc-alkaline tonalitic arc rocks) (Valentino et al., in press). Within the shear zone, the original megacrystic granite contains lineated quartz and rodded feldspar aggregates up to a meter long in places. Along the axis of the shear zone there are thick (1-2 km), subvertical zones of granitic L-S and L-tectonites. The northern domain of the zone is defined by large foliation domes that are cored by L-tectonite. The southern limbs of the domes steepen toward the south and merge with a wide zone (up to 15 km) of steeply dipping granitic mylonite. Overall, the shear system (domes and steep mylonite zone) forms the core of a region of intense ductile deformation with left-lateral kinematic indicators and subhorizontal E-W ribbon lineations.

The Piseco granitic suite are highly deformed suture-stitching arc plutons that intruded within a sinistral, oblique-convergent, shear system in the deep crust during the Shawinigan orogeny. This is ductile shear zone is the most continuous and largest in the entire Adirondack massif. The shear zone, associated granitic rocks, and the magnetic anomaly abruptly trends toward the south in the eastern Adirondacks. Just beyond this location, the magnetic anomaly is truncated by a branch of the NY-AL magnetic lineament. Following the

trace of the magnetic anomaly toward the west, suggests that the shear zone continues for a considerable distance beyond the Adirondack window. The size, in addition to the magnitude and extend of the associated magnetic anomaly, suggests that the Piseco shear zone penetrates the Moho.

The current field trip is an update on our very long research project, and it's geared toward an undergraduate student audience. All field locations were picked to accommodate large student groups. Sampling in the Adirondack Park is generally prohibited by NYS law, and we encourage future instructors to help preserve the field locations presented herein by showing and discussing, and not removing the spectacular bedrock features. Note that the field guide included here in was pirated and modified from the Friends of the Grenville field conference run by D. Valentino, J. Chairenzelli, D. Piaschyk, L. Williams and R. Peterson in 2008.

INTRODUCTION

The Adirondack Mountains are a relatively recently uplifted (Roden-Tice and Tice, 2005), domal exposure of Mesoproterozoic high-grade gneisses that are part of the contiguous Grenville Province (Figure 1). Sharing many similarities to nearby rocks in Ontario and Quebec, and Grenville basement inliers in the Appalachians (McLelland et al., 2010), the rocks exposed in the Adirondacks record processes in the deep crust related to a series of orogenic events collectively known as the Grenville Orogenic Cycle (McLelland et al., 1988). Undergoing deformation spanning a period of over 250 million years (ca. 1250-1000 Ma), the Grenville Province of Laurentia is a small part of a world-wide system of orogenic belts whose assembly led to the eventual formation of the supercontinent Rodinia.

The Adirondacks have been subdivided into the Lowlands and Highlands based on difference in metamorphic grade, predominance of supracustal verses metaigneous rocks, and topography. The Carthage-Colton shear zone in the northwestern Adirondacks (Geraghty et al., 1981; Wiener, 1983), is the boundary between these terranes. The shear zone displays evidence of one or more ductile, high-grade events, as well as, later brittle remobilization and intrusion by leucogranites (Selleck et al., 2005). It has been interpreted as a late, brittle, normal fault accommodating orogenic collapse and, although the earlier ductile history is obscured, thrust and strike-slip kinematics have been noted (Baird and MacDonald, 2004; Wiener, 1983). Traditionally, lithologic similarities with the Central Metasedimentary Belt in Canada has led to inclusion of the Adirondack Lowlands as part of this terrane; whereas, the Adirondack Highlands has been equated to the Central Granulite Terrane of Quebec, again based on lithologic similarities and metamorphic grade. More recently Rivers (2008) has proposed orogen-wide subdivisions based on geochronology, and metamorphic and structural data. He suggested that the Adirondacks were part of a terrane accreted during the Shawinigan Orogeny (ca. 1140-1200 Ma) and subsequently were part of the orogenic lid; part of a medium to low pressure belt of allochthonous rocks that lack the widespread deformation that occurred elsewhere during the Ottawan Orogeny (ca. 1020-1080).

Tectonic models in the South-Central Grenville Province suggest that southeastern margin of Laurentia (present coordinates) has been the site of subduction and accretionary processes for much of the period between 1200-1500 Ma (Carr et al., 2000; Hanmer et al, 2000; Rivers and Corrigan, 2000). Part of this pre-orogenic history includes a period of back-arc, failed rift spreading in the Central Metasedimentary Belt (Dickin and McNutt, 2007) and the opening of a marginal sea within the current location of the Trans-Adirondack Back-arc Basin (Chiarenzelli et al., 2012). This began ca. 1300 Ma and was terminated by the Elzevirian Orogeny at ca. 1240 Ma. The continental arc developed along the leading edge of Laurentia (ca.1300-1350 Ma) was dismembered and dispersed by this subsequent rifting. Fragments of this arc are now found in Ontario, Quebec, the Green Mountains of Vermont, and in the Southern and Eastern Adirondacks, and were accreted to Laurentia during the Shawinigan Orogeny (Figure 1). Collectively these

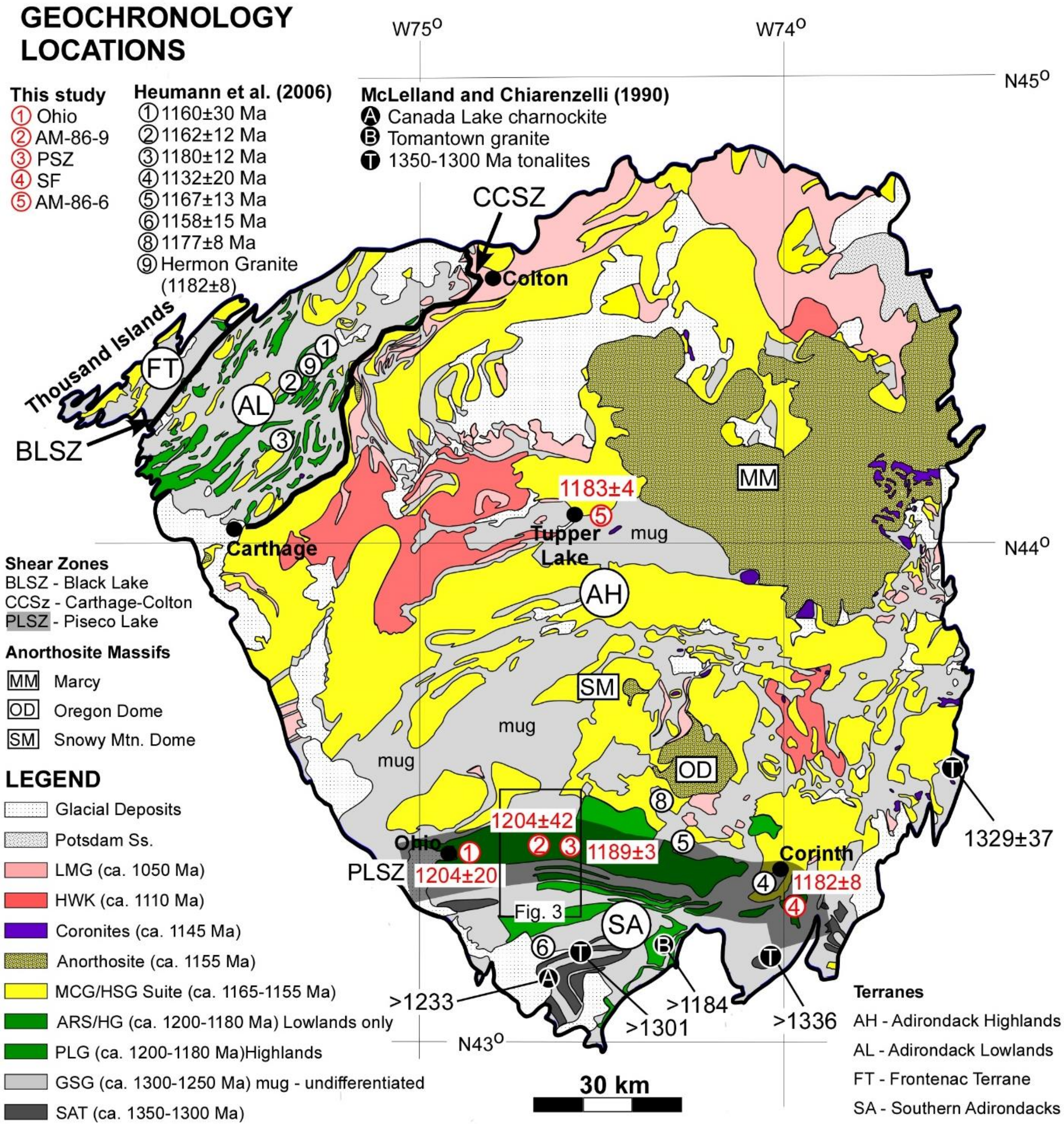


Figure 1. *Simplified geological map showing rock suites (from Valentino et al. – in press): ARS – Antwerp-Rossie suite; GSG – metasedimentary rocks of the Grenville Supergroup; HG – Hermon granitic gneiss; HSG – Hyde School gneiss; HWK – Hawkeye granite; LMG – Lyon Mountain granite; MCG – Mangerite-charnockite-granite suite (granitoid part of AMCG suite); mug- mixed metasedimentary and metaigneous rocks that are most likely correlative with GSG; PLG – Piseco Lake gneisses; SAT – Southern Adirondack tonalite suite. U-Pb zircon analysis results (red circles – this study; white circles – Heumann et al., 2006; black circles – McLelland and Chiarenzelli, 1990). Numbered red circles (1-5) are geochronology locations for this study. Location of Figure 3 show as thin black rectangle.*

rocks have been called the Dysart-Mt. Holly Complex (Hanmer et al., 2000) after locations in Ontario and Vermont, respectively. The ca. 1300-1350 Ma tonalitic rocks in the Southern Adirondacks (McLelland and Chiarenzelli, 1990) and the Green Mountains of Vermont (Ratcliffe et al., 1991) are considered part of this arc or arcs.

THE PISECO GRANITOID AND SHEAR SYSTEM

The Piseco granitoid suite (Figure 2), located in the southern Adirondacks, was originally thought be part of a basement complex to the Adirondack supracrustal sequence (McLelland and Isachsen, 1986), and were later characterized as granitic members of the AMCG suite (McLelland et al., 1988; 2004; Hamilton et al., 2004). Most recently, evidence supports an independent origin, and slightly older age for these rocks, which are exclusively found within, and along strike, of the highly tectonized Piseco shear zone (Valentino et al., in press). Their ubiquitous high strain prohibits detailed characterization of primary textures at most locations. However, the bulk mineralogy and detailed geochemistry, in addition to very large recrystallized mineral aggregates suggest that these rocks were predominantly igneous, megacrystic, and associated with arc Shawinigan plutononism (Valentino et al., in press).

In contrast to the dominant northeast structural trends throughout most of the Grenville Province, the south-central Adirondack Highlands structural grain is predominantly east-west (Figure 2), including the belt of Piseco granitoids. Across strike, where the overlying Paleozoic cover rocks have been stripped away, this region is greater than ~150 km wide and displays general parallelism of geologic contacts, fold axes, compositional layering, strike of foliation, the trend of mineral elongation lineations and substantial (~5-10 km wide) zones of mylonite. Several large (>20 km across) structural domes, cored by rheologically rigid anorthosite, lie within the zone (Chiarenzelli et al., 2000; Gates et al., 2004; Valentino et al., 2004; Valentino et al., 2008), and kinematic investigations indicate that this zone is dominated by left-lateral transpressive shear (Chiarenzelli et al., 2000; Gates et al., 2004; Valentino et al., 2008). There are a number of large-scale features, such as drag folds and rotated giga-scale clasts (i.e. Snowy Mountain anorthosite body), which are consistent with the abundant meso- and micro-scale kinematic indicators including S-C fabrics, shear bands, and rotated porphyroclasts of various minerals (Gates et al., 2004). Here we designate this east-west striking, broad zone of gneisses as the Central Adirondack Shear System (CASS), and the crust-scale structures of the CASS are interpreted as the consequence of transpressional modification of earlier recumbent folds, analogous to those exposed in the Adirondack Lowlands (Chiarenzelli et al., 2000). The CASS is superimposed on rocks that contain widespread granulite-facies mineral assemblages, deformed migmatitic gneisses, and substantial volumes of supracrustal rocks, all supporting earlier history of compressional tectonism as described above. However, discrete mylonite zones within the CASS were shown to contain retrograde deformation fabrics, with some containing fabric forming greenschist facies minerals, such as biotite, chlorite and muscovite (Price et al., 2003; Valentino et al., 2008). These relationships suggest that structural activity within the CASS outlasted high-grade conditions and continued through denudation, uplift and cooling of the central Adirondacks. It was previously proposed that the locus of deformation within the CCAS, the Piseco Lake shear zone, marks the boundary of oblique-slip convergence between the southeastern margin of Laurentia and the Southern Adirondack arc terrane (Gates et al. 2004; Valentino et al., 2008).

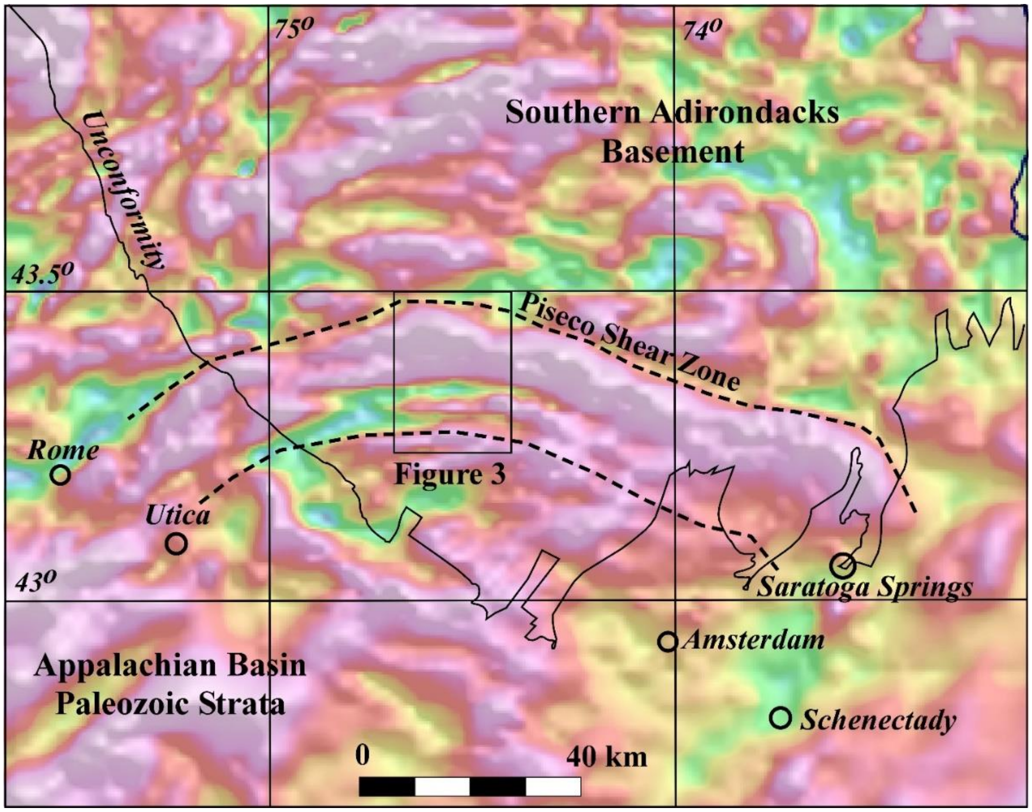


Figure 2. *Magnetic anomaly map of the southern Adirondack region. Location is shown on Figure 1.*

THE PISECO LAKE SHEAR ZONE

Several discrete ductile shear zones occur within the CASS (Gates et al., 2004; Valentino et al., 2008; Weimer et al., 2001), with the Piseco shear zone as the most continuous and largest spanning the entire 120 km width of the Adirondack massif and upward of 30 km wide (Figures 2 and 3). A prominent magnetic anomaly correlates with the Piseco Lake shear zone (PLz) and it appears that the zone extends well beyond the exposed limits of the Adirondack massif (Figure 2). Strong deformation in the area surrounding Piseco Lake was first noted by Cannon (1937). Fakundiny (1996) and Fakundiny et al. (1994) proposed a fundamental structural discontinuity in this area based largely on geomorphological criteria and the trace of the Prospect Fault (Valentino et al., 2012). In addition, fundamental lithologic differences have been noted across the boundary as AMCG rocks are rare or absent to the south, whereas tonalitic rocks (ca.1300-1350 Ma) of the Southern Adirondack arc terrane have not been recognized north of it. The PLz is developed in a suite of granitoids (Piseco Lake Granitoids – PLG) that span the width of the Adirondack massif outcrop belt, but more importantly the belt strongly correlates with the pronounced linear magnetic anomaly.

The PLz is defined by spectacular L-S, L>S and L-tectonites developed in the Piseco granitoids, with an east-west arching trend. This trend continues smoothly to the eastern margin of the Adirondack Dome, where before plunging beneath Paleozoic cover rocks to the east at Spier Falls on the Hudson River, where the lineation and foliation gradually transitions to a north-south orientation defining a broad (10's of km) open vertical fold. The PLz comprises parallel structural domains that developed contemporaneously: 1) a broad (10-15 km wide) tabular steeply dipping zone of granitic mylonite (southern domain); 2) a series of flanking upright (5-10 km wide) foliation domes (northern domain) (Cannon, 1937; Glennie, 1973; McLelland, 1984; Wiener et al., 1984). There is no apparent structural discontinuity between the foliation of the southern mylonite zone and the foliation in the dome region. As well, lineations are consistent in orientation and defined by the same minerals within both domains. Collectively, these structural domains make up a zone of ductile deformation that is upward of 30 kilometers wide, crossing the exposed width of the boundary between the Central and Southern Adirondacks.

The Southern Domain – Shear Zone

There is a well-defined textural transition that occurs in variably deformed granitoids in the southern limit of the PLz. From south to north, the granitic rocks exhibit moderately deformed megacrysts of K-feldspar, well developed mylonite with remnant K-feldspar grains (~5-10 mm in diameter) and finally domains of ultramylonite (Valentino et al., 2008). Penetrative foliation is defined by dynamically recrystallized quartz, K-feldspar and plagioclase, and alignment of micas. Rocks within the zone are made of fine-grained aggregates of these minerals in strong alignment. Locally there are 2-6 cm long K-feldspar megacrysts and/or porphyroclasts preserved which are consistent with a plutonic origin. Pegmatitic and aplitic layers provide evidence of strong transposition of primary contacts (Piaschyk et al., 2005; Valentino et al., 2008). Rocks with steeply dipping penetrative mylonitic fabric persist over an across strike distance of more than 10 km, eventually merging with the highly deformed granitoids in the northern domes. Lineations in this zone are subhorizontal and defined by dynamically recrystallized quartz, K-feldspar, plagioclase, and accessory mafic minerals such as hornblende, biotite or chlorite. Shear sense indicators are abundant and occur at the meso- and microscopic scale. K-feldspar megacrysts form Type I S-C fabrics,□□- and □- porphyroclasts and domino-structures (Figure 4). Where the L-S fabrics are well developed, consistently the shear sense indicators are sinistral across the entire 10 km wide zone of the southern domain.

Northern Domain – Dome

Penetrative foliation and lineations define several upright antiformal domes (Cannon, 1937; Glennie, 1973; Weiner et al., 1984; Valentino et al., 2012) that flank the north side of the steeply dipping mylonite zone (southern domain). These domes have subhorizontal arching axes that trend approximately 110^{o} in the east, 090^{o} in the central region, and 080^{o} in the west. The largest of these domes occurs in the vicinity of Piseco Lake. The foliation on the dome limbs dips moderately with mineral elongation lineations that trend at 110^{o} and plunge about 10^{o} eastward. In the crest of the domes, foliation is not well developed and penetrative lineations are defined by mineral rods, rods of mineral aggregates, and mineral ribbons. Lineations in the domes are intensely developed, and in many places the linear fabric is dominant over the weak foliation (L>>S) with grain aggregate aspect ratios upward of 60:1, in the L-parallel and S-perpendicular plane, as originally noted by McLelland (1984). Some rocks, of considerable thickness ~1-2 km, in the core of the domes lack foliation altogether and are true L-tectonites. Microscopic examination showed that the lineations are defined by dynamically recrystallized ribbons and rods of quartz, K-feldspar, plagioclase, in addition to streaks of magnetite, biotite, chlorite and occasionally muscovite (Valentino et al., 2008).

Like the rocks in the southern domain, shear-sense indicators are abundant in the dome rocks that are not dominated by L-tectonite. They include Type I S-C mylonite, σ - and δ -porphyroclasts of K-feldspar, and asymmetric polymineralic tails around porphyroclasts (Figure 4). These kinematic indicators reveal a consistent sinistral-shear sense on both the north- and south-dipping L-S tectonite domains of the domes.

Magnetic Anomalies

The magnetic anomaly map of North America (U.S.G.S. – Mineral Resources Online Spatial Data) has an interesting distribution of high and low anomalies in the Adirondack region that roughly correlate with major metaigneous and metasedimentary rock bodies (Figure 2). The corresponding magnetic anomaly for the anorthosite bodies is low, while vast regions underlain by charnockitic gneiss express high anomalies. Regions of the Adirondacks with substantially thick sequences of supracrustal rocks generally have low magnetic anomaly signatures, similar to the anorthostie bodies. Magnetic anomaly patterns in the Adirondack lowlands are parallel to the overall northeastern striking geologic structures that control the geographic distribution of various metaigneous and metasedimentary rock bodies. Within the Adirondack Highlands, the northern region can be characterized as having magnetic anomalies with a nebulous, or unorganized, structural pattern, most likely the result of the vast anorthosite bodies that make up the Mount Marcy massif.

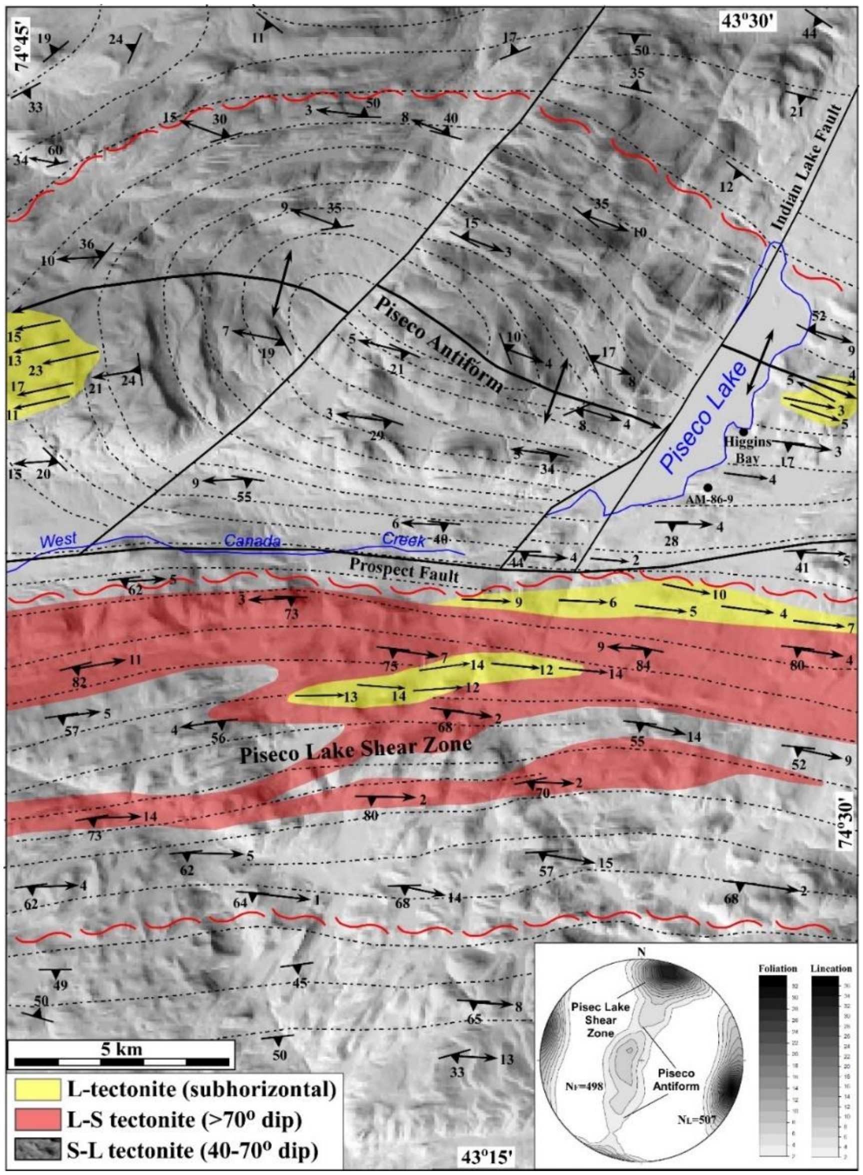


Figure 3. *Simplified structural geology map for the Piseco Lake shear zone and dome plotted on a gray-scale digital elevation model. Location is shown on Figure 2. Inset shows lower hemisphere contour stereogram for poles to foliation and lineation data collected in the mapped region. Note that the lineation data is a composite plot for both the antiform and shear zone domains; foliation is designated separately. The sinuous dashes show the approximate boundaries of the Piseco Lake shear zone in this area. The axis of the Piseco antiform is shown in addition to Mesozoic faults (heavy black lines) that offset the antiform axis near Piseco Lake (Valentino et al., 2012).*

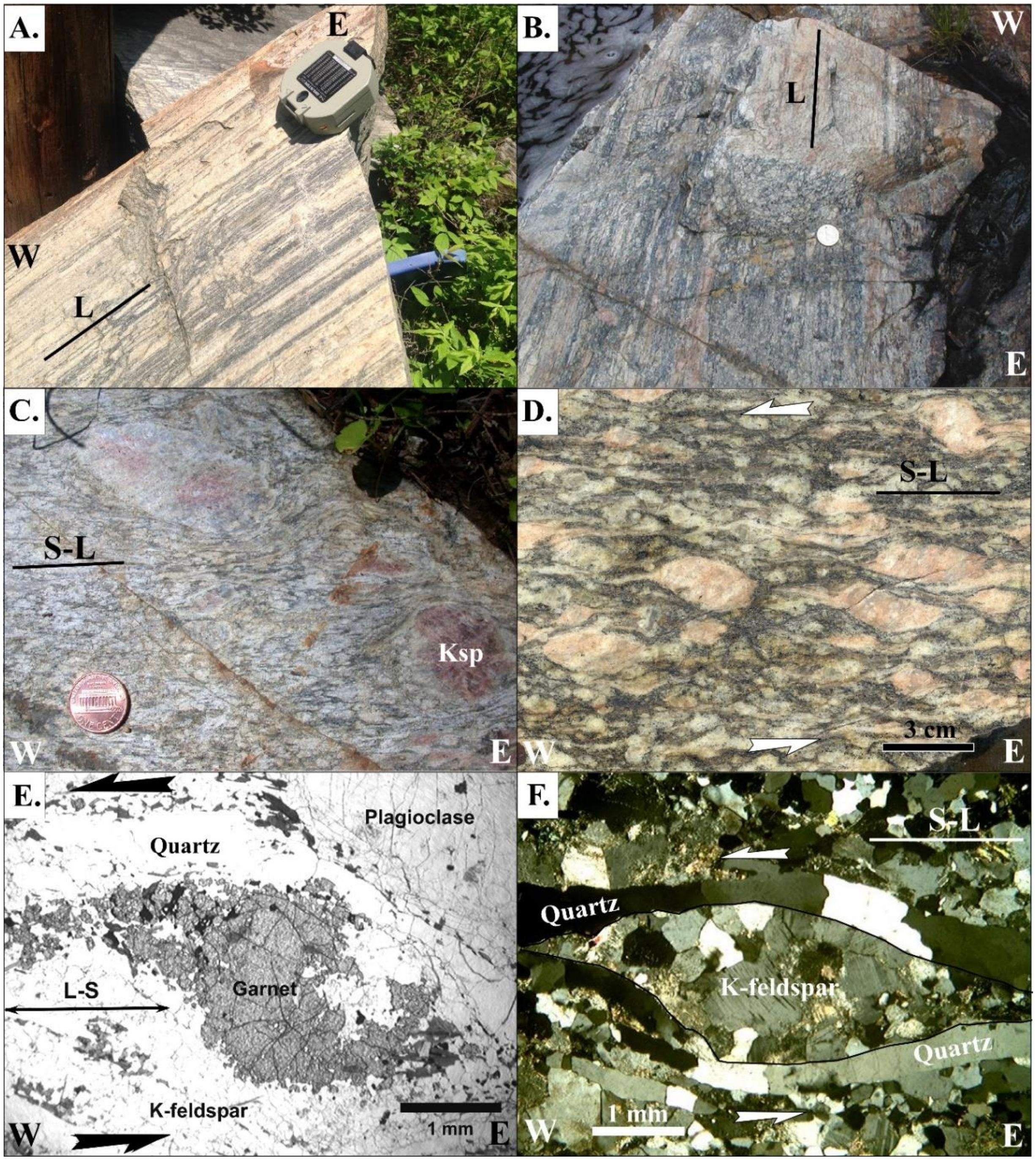


Figure 4. *Examples of deformation fabrics in the Piseco shear zone. A. Strongly lineated granitic mylonite from the southern flank of the structural dome near Piseco Lake. B. Tectonite dominated by mineral aggregate elongation lineations from the core of the dome. C. Well developed granitic mylonite from the southern steeply dipping shear zone. D. Augen mylonite with relict K-feldspar megacrysts forming asymmetric ⍰-porphyroclasts on the northern flank of the dome. E. Photomicrograph of asymmetric garnet enveloped in quartz ribbons, and recrystallized K-feldspar and plagioclase (ppl). F. Photomicrograph of highly deformed granitic gneiss from the Piseco shear zone (xpl) displaying polygonal grains of dynamically recrystallized K-feldspar and quartz ribbons. The aggregate of K-feldspar (center) is surrounded by quartz ribbons and the overall shape forms a recrystallized asymmetric sigma-porphyroclast.*

With the exception of small anorthosite bodies with ovoid low magnetic anomalies, the structural pattern of anomalies in the central and southern Adirondacks is strikingly regular and trends roughly east-west parallel to the structures of the Central Adirondack Shear System (Chiarenzelli et al., 2000; Gates et al., 2004; Valentino et al., 2008). One of the most pronounced high magnetic anomalies extends unbroken across the entire exposed width of the southern Adirondacks, forming a broad open arch that curves to a north-south trend in the east and appears to be sharply truncated near Saratoga Spring, NY. This anomaly extends west of the Adirondacks where it bifurcates and continues beneath the Appalachian basin of central NY. This high magnetic anomaly has the most structural continuity for any anomaly associated with the exposed basement geology, and it directly correlates with the highly deformed Piseco granitoid suite and related shear system (McLelland and Isachsen, 1986; McLelland et al., 1988; 2004; Gates et al., 2004; Hamilton et al., 2004; Valentino and Chiarenzelli, 2008).

ORIGIN OF THE PISECO LAKE GRANITOIDS

The limited compositional range, megacrystic texture, coarse grain-size, and large volume of the Piseco Lake granitoids suggest they are an intensely deformed and deep-seated suite of granitic plutons. The occurrence of deformed and transposed cross-cutting aplites and pegmatites is in concert with this interpretation (Piaschyk et al., 2005; Price et al., 2003). Previous U-Pb zircon studies (McLelland et al., 1988; Valentino et al., in press) confirm that the granitic precursors were intruded into the region by at least 1155 Ma and likely earlier (ca. 1200 Ma). The work of Heumann et al. (2006) has suggested that highly deformed paragneisses in contact with the Piseco Lake granitoids underwent anatexis and zircon and monazite growth from 1160-1180 Ma. If the Piseco Lake granitoids provided some, or all, of the heat that facilitated the melting, a significant volume of melt/heat must have been present by 1180 Ma. This is consistent with our preliminary zircon studies previously published (Valentino et al., in press), indicating intrusion prior to AMCG plutonism at ca. 1155-1165 Ma and likely at ca. 1200 Ma.

The fabric in the Piseco Lake granitoids and surrounding mylonitic pssamitic/pelitic gneisses, and hence the gross structure of the shear zone itself, can be tied to zircon and monazite growth during high-grade ductile deformation and melting during the waning phase of the Shawinigan Orogeny (Heumann et al, 2006). This is emphasized by the lack of younger “Ottawan” zircons in many of the paragneiss localities in the northwestern, central, and southern Adirondacks studied by Heumann et al. (2006) and in mylonitic gneiss analyzed for this study. Zircons younger than 1080 Ma are nearly absent, indicating minimal, if any, zircon growth during the Ottawan Orogeny along the Piseco Lake Shear Zone; but volumetrically significant metamorphic zircon growth occurred throughout the Shawinigan Orogeny. This in turn provides further evidence that the gross crustal architecture of the Central and Southern Adirondack Region and Central Adirondack Shear System was developed during the Shawinigan Orogeny. The Piseco Lake granitoids provide direct evidence of the processes at work in the deep crust and likely, upper mantle, just prior to and during, the Shawinigan Orogeny. In essence they, and their intense deformation, set the stage for the voluminous AMCG intrusions that followed (Valentino et al., in press).

The kinematic studies on all scales from microscopic to megascopic (Chiarenzelli et al., 2000; Gates et al. 2004; Valentino et al., 2008) indicate left-lateral motion focused along the shear zone and domes developed in the Piseco Lake granitoids. The focus of this intensive deformation, between two distinct terranes, has led to the conclusion that it demarcates a cryptic suture (Chiarenzelli et al., 2011; 2010a; Valentino et al., in press. Antiformal domes and the counterclockwise rotation of rigid anorthosite bodies (Chiarenzelli et al., 2000; Gates et al., 2004) suggest bulk crustal flow throughout the CASS region, as do vertical zones of L-tectonite and rocks with pronounced linear fabrics. Taken together these observations are consistent with the intrusion of a voluminous suite of suture-stitching plutons of arc affinity within a sinistral, oblique-convergent, ductile shear system during the Shawinigan orogeny.

TECTONIC MODEL

Any tectonic model proposed to explain the origin and deformation of the Piseco Lake granitoids must take into consideration their field relations, age, geochemical and isotopic trends, intense deformation, kinematics, geophysical signature, and the geologic context of the region. Various lines of evidence presented in Valentino et al. (in press) and summarized herein, suggest that the granitoids are the deformed roots of a continental batholith which developed just prior to, and during, the Shawinigan Orogeny and preceded voluminous AMCG plutonism.

Paleogeographically, the plutonic protoliths of the gneisses appear to represent the product of northward subduction of oceanic crust beneath the Southern Adirondack arc, Trans-Adirondack Back-Arc basin, and the southeast edge of Laurentia, believed to be coincident with the Black Lake Shear Zone at this time (Chiarenzelli et al., 2010b; Wong et al., 2012; and Peck et al. 2004; 2013). Telescoping of this basin during closure led to massive shortening and collapse of the basin and attendant SW-directed thrusting and nappe formation. However, the dominant fabric, which overprints early events, was left-lateral plastic deformation related to oblique collision. The imprint of this late Shawinigan event is recorded at all scales from microscopic kinematic features to “mega” porphyroclasts and elongate structural domes (Chiarenzelli et al., 2000; Gates et al., 2004; Valentino et al., 2008) to a regional subhorizontal lineation.

AMCG plutonism began during the waning stages of Shawinigan orogenesis (Valentino et al., in press). This can be seen in the Marcy anorthosite massif where large, relative intact bodies of meta-anorthosite are cut by thin (1-2 cm) garnet-pyroxene shear zones, many with an E-W orientation (Hecklau et al., 2014). Intrusive boundaries, along coarse-gabbroic pegmatites and fine-grained granitic sheets, are often zones of intense strain. This suggests the meta-anorthosite body, although rigid and resistant to ductile deformation in general, underwent ductile deformation along heterogeneities. Coronitic metagabbros, thought to be the parental magmas from which anorthosite was derived via fractional crystallization (Regan et al., 2011), may provide a lower limit on regional ductile deformation associated with the Shawinigan Orogeny. Coronites display a wide range of deformational and metamorphic features ranging from equant bodies with pristine coronitic textures to elongate, curvilinear belts of garnetiferous amphibolite. In many instances garnetiferous amphibolites retain a small core of coronitic metagabbro that survived deformation (Lagor et al., 2013). Dating of one coronitic metagabbro from Dresden Station yielded an age of 1144+/-7 Ma via U-Pb zircon methods (McLelland and Chiarenzelli, 1989) and extends the range of ages generally attributed to the AMCG suite (ca. 1155-1165 Ma) to at least 1144 Ma. This age overlaps the age of 1151+/-9 Ma for monazite growth in the pelitic gneisses intruded by the Dresden coronitic metagabbro (Grover et al., 2013), indicating the growth of high-grade minerals in compositionally appropriate rocks during intrusion of the coronites.

The transition from arc magmatism represented by the Piseco Lake granitoids to intrusion by anorthosite massifs and cogenetic granitic rocks (AMCG suite) occurred within a relative short period of time; at most several tens of millions of years (Figure 5). This spatial and temporal link suggests that intense deformation associated with the Piseco Lake granitoids was the kinematic trigger for AMCG magmatism. Most models for AMCG rocks invoke for slab detachment or delamination, but few details are known. One possible explanation presented by Valentino et al. (in press) and favored here is that highly oblique subduction and orogeny-parallel deformation may have contributed to detachment and delamination. Shear stress may have reactivated old crustal weaknesses (transform-faults) and/or created tears that propagated into the descending slab and lower plate, resulting in splitting and fragmentation of the rigid lithosphere. Catastrophic failure of this type would allow rapid ascent of asthenospheric mantle, decompressional partial melting of the asthenosphere, and subsequent melting of underlying crustal rocks. Given left-lateral kinematics documented, the progressive closure of the ocean basin the foci of asthenospheric rise and production of AMCG magmatic rocks would propagate from west to east.

An analog for this model would incorporate aspects of the Andean margin where subducting slabs of different age and density behave as independent lithospheric "tiles". These tiles are separated by oceanic fracture zones, subducting at different rates and angles beneath South America, and control the distance between the magmatic arc and trench. In combination with highly oblique convergence, the oceanic fracture zones between these "tiles" would serve as inherent zones of weakness ultimately causing catastrophic tears in the subducting lithospheric plate. A similar scenario involving tearing and propagation of a subducting slab undergoing buckling is currently occurring along the Puerto Rican trench (Meighan and Pulliam, 2013; Meighan et al., 2013).

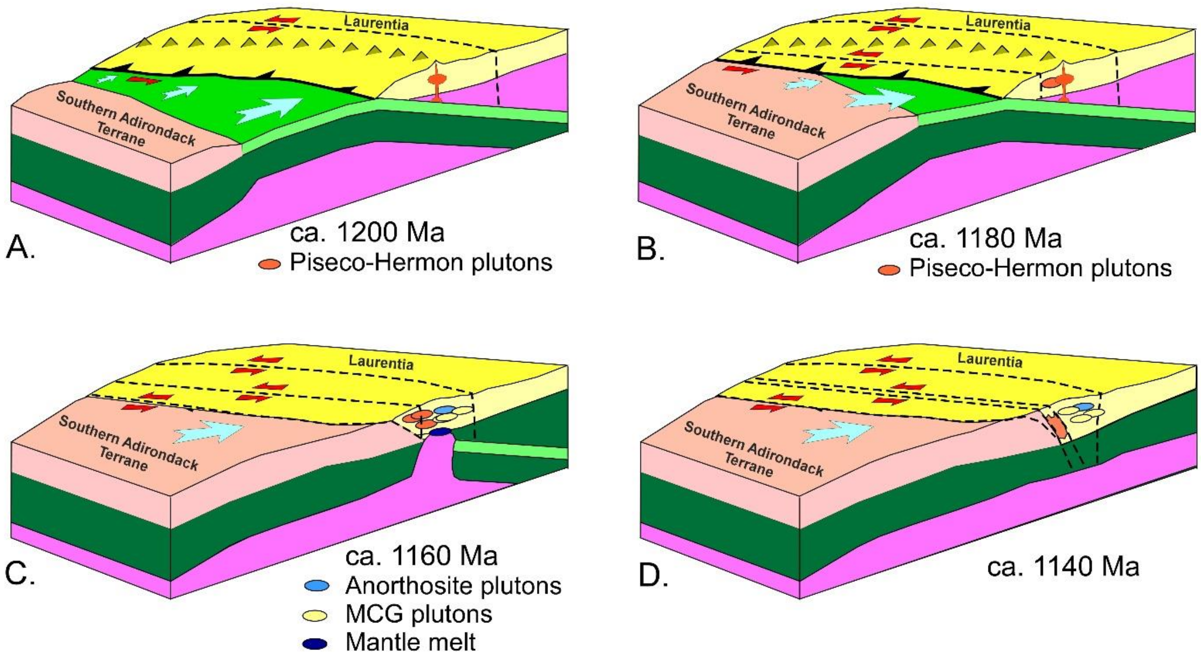
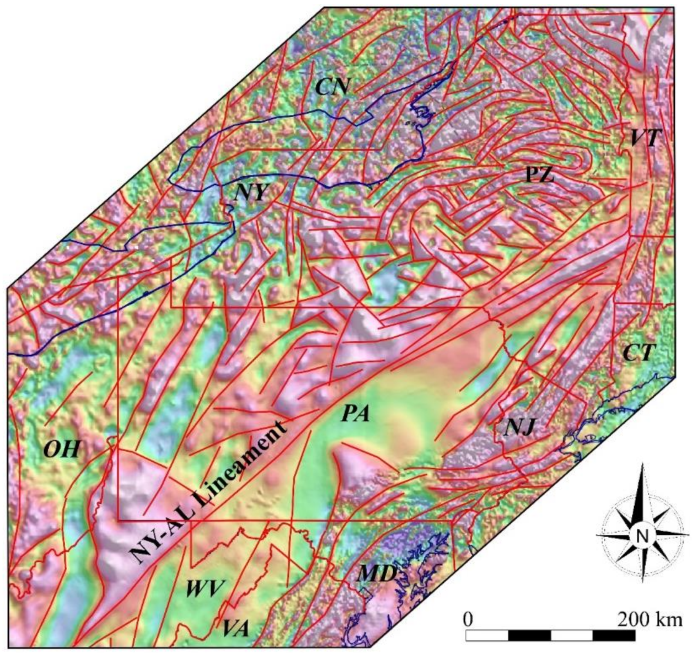


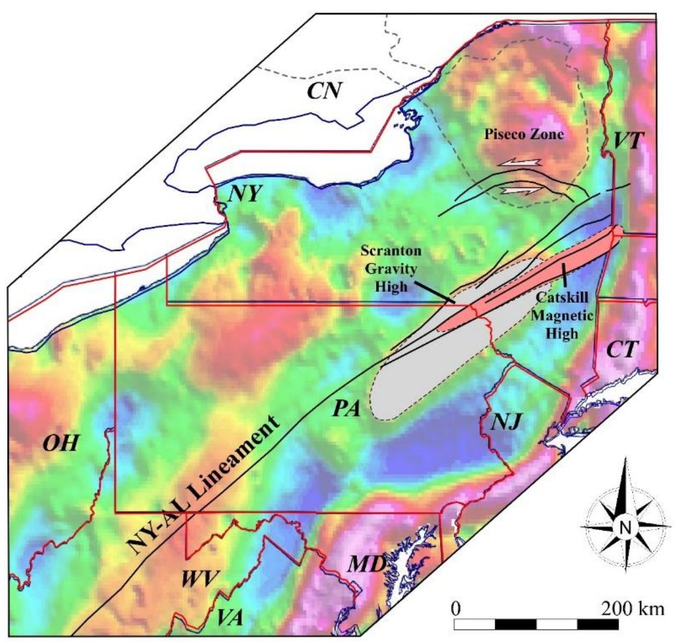
Figure 5. *Tectonic model depicting the oblique subduction, subsequent oblique collision and sinistral transform boundary that forms on the granite-stitched suture between the Southern Adirondack Terrane and the Adirondack Highlands (Laurnetia). Refer above for details.*

This tectonic model would not be complete without discussing the proximity of the Piseco shear zone to the New York-Alabama magnetic anomaly lineament (NY-AL), an anomaly that defines a major basement boundary that crosses eastern Laurentia (Steltenpohl et al., 2010). The origin of the anomaly is not definitively known, however, recent researchers have suggested that the linear nature of the anomaly is associated with an intracrustal transcurrent shear system with either dextral or sinistral displacement.

Between northwestern Georgia and northeastern Pennsylvania, the NY-AL lineament trends without deviating about 046°, a distance greater than 1000 km (Figure 5). With an easterly change in trend of 15-20°, the lineament is shown to continue northeast and include the Catskill magnetic high extending from northeastern Pennsylvania to the Vermont-Massachusetts border region (Figure 6A). This trend crosses the Scranton gravity high, proposed to be a Neoproterozoic rift basin (Rankin, 1976; Hawman and Phinney, 1992). However, if the 046° trend of the >1000 km long NY-AL lineament is projected into southern New York, it would correspond to the western margin of the Scranton gravity high (Figure 6B), in addition to the apparent truncations of a series of magnetic high anomalies, including the Piseco anomaly, and by association the granitoids and shear zone. Based on the transcurrent deformation associated with the PLz, we propose the zone to be a splay off of the major crustal boundary that is manifest as the NY-AL magnetic lineament (Steltenpohl et al., 2010).



A.

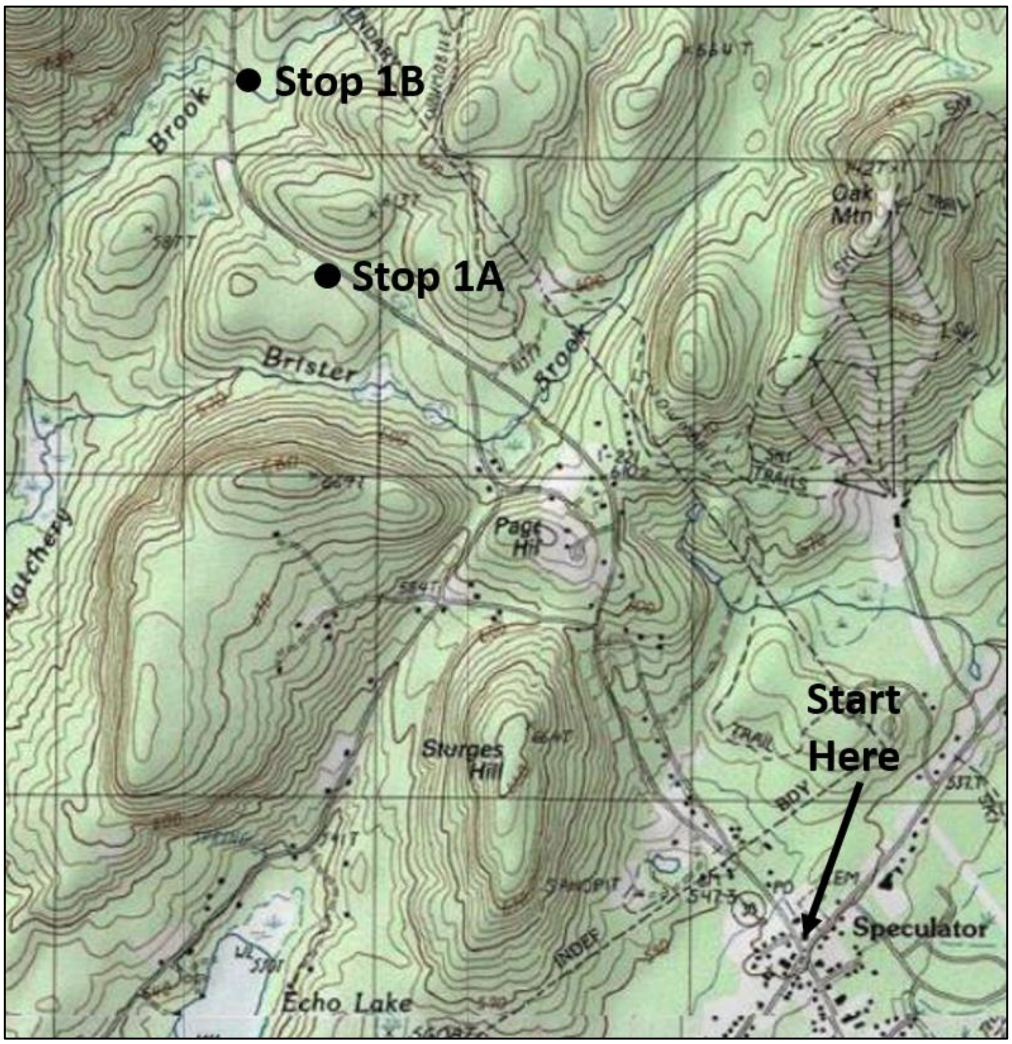


B.

Figure 6. *(A) Magnetic and (B) gravity anomaly maps with interpreted lineaments. The New York-Alabama lineament (NY-AL) is labeled along with the Scranton gravity high and the Catskill magnetic high. Outline of the Adirondacks is represented by the dashed line and the Piseco Lake shear zones (PZ) is shown with shear sense arrows (Maps from USGS respository online).*

FIELD GUIDE AND ROAD LOG

Assemble in the parking lot at the Charlie John's market located the intersection of Route 8 and Route 30 in the village of Speculator, New York (Starting Point on map below). From this point the trip heads north on Route 30 to the first stop.



Mileage:
0.0 Assembly point
1.7 Stop 1: Park on the wide shoulder on the east side of Route 30. The outcrop is on the west side of the road, the traffic is light but fast, so be very careful crossing the road.

STOP 1A: Supracrustal rocks of the Adirondack Highlands, just north of the Piseco Lake shear zone (18T 549883 m E, 4818851 m N) 43.520931 -74.382762

There are several rock types that can be observed at this outcrop, along with the contacts and primary compositional layering. The northern ¼ of the outcrop consists of garnet amphibolite, the next ~ ¼ of the outcrop is calc-silicate gneiss and the southern ½ is fine-grained quartzo-feldspathic gneiss locally interlayered with highly folded marble (Figure 7). Complex sheath folds with sub-horizontal tight, isoclinal and sheath folds of foliation and compositional layers dominate the rock, particularly in the gneiss (trend to 110^o). Excellent views of the folds are seen on top of the southern part of the outcrop. The contacts between the rock types are also folded in similar fashion. Matrix minerals define lineations that are E-W and subhoriziontal, but are variable in orientation due to folding. Overall, the diverse lithologies and style of deformation are typical for the supracrustal rocks located north of the Piseco Lake grantioids. At this point, note that primary compositional layering is preserved at this outcrop.

Figure 7. *Photograph of folded quartzo-feldspathic gneiss and marble (recessed part of outcrop). Note the stalk of grass for scale.*

STOP 1B: Piseco Lake granitoid (18T 549600 m E, 4819452 m N)

From Stop 1A, walk north along Route 30 approximately 400 meters to Hatchery Brook crossing, and then follow the foot path on the east side of Rt 30 to the small waterfall. The water falls over a small ledge of Piseco Lake granitoid, and observable outcrops occur to the right. At this location, the megacrystic granitoid is mildy deformed (Figure 8). There is a weak alignment of deformed feldspar and quarts forming shallowly dipping foliation and mineral lineation that trends approximately east-west. Note that the mineral elongation lineation is parallel to the axes of small folds and mineral lineations at Stop 1A. Also note that this occurrence of Piseco Lake granioid is north of the main igneous body that will be observed later on this trip, and that this location will be the only place where you will see Piseco Lake granitoid in a state of low strain. Walk back to vehicles and follow trip directions.

From Stop 1, proceed north to a safe place to turn around, and return to the intersection of Route 30 and Route 8 in the village of Speculator.

Mileage:

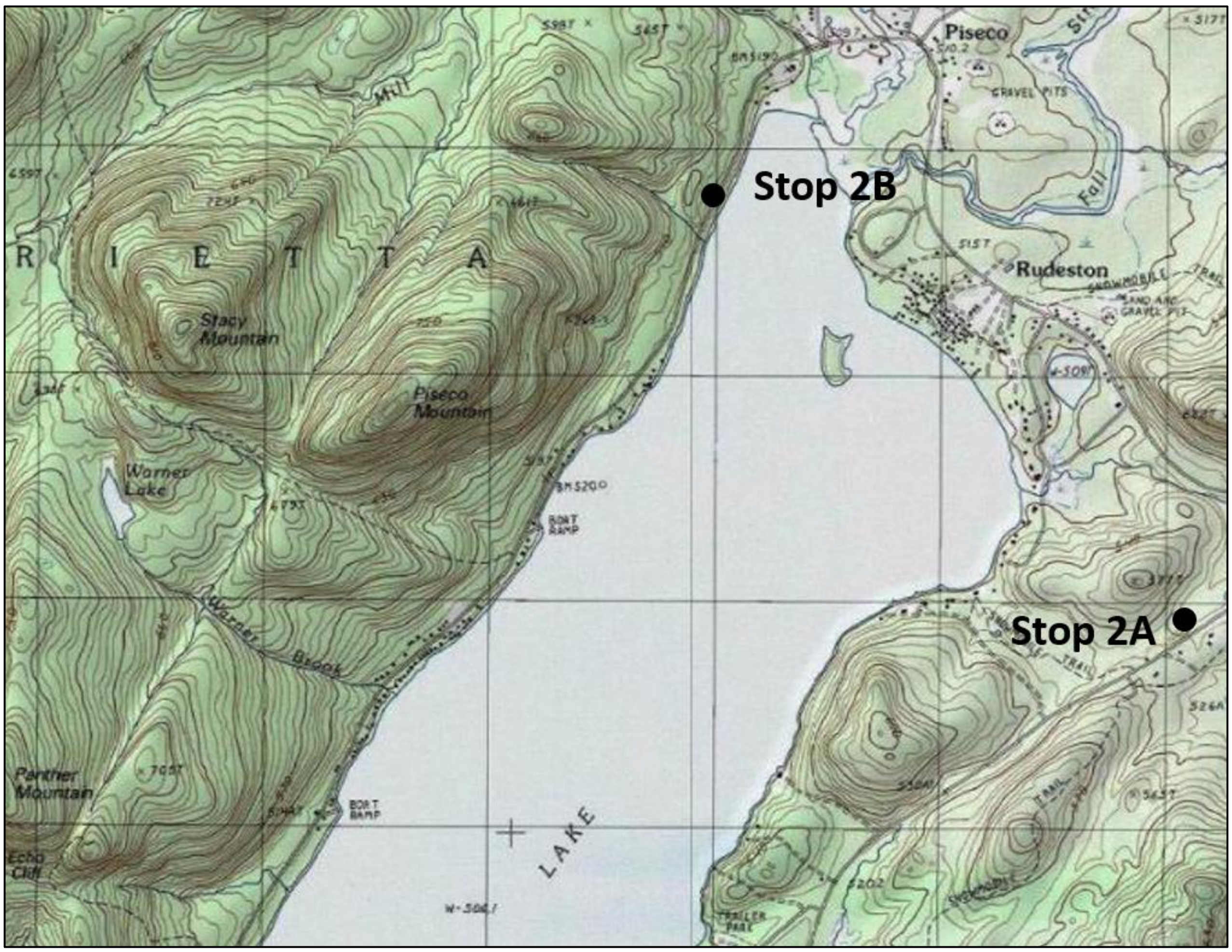
3.4	Turn west on Route 8 in the Village of Speculator.
5.4	First roadcut on right is highly lineated granitic gneiss in the margin of the PLsz.
12.2	Intersection with Old Piseco Road in Piseco, NY.
13.3	Park on north side of Route 8 at the low roadcut for Stop 2.

Figure 8. *Piseco Lake graniotid at Hatchery Brook.*

STOP 2A-D: L-S and L>S fabrics in the Piseco dome – driving traverse

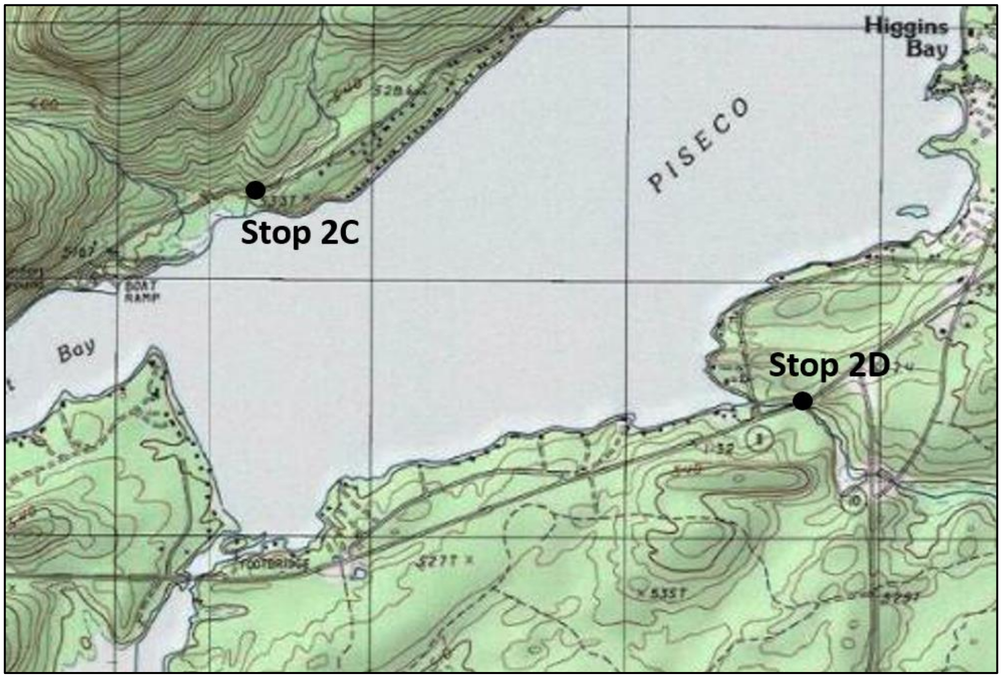
Stop 2A: 18T 540256 m E, 4808187 m N
Stop 2B: 18T 537957 m E, 4809931 m N
Stop 2C: 18T 534370 m E, 4805562 m N
Stop 2D: 18T 536919 m E, 4804838 m N

The Piseco Lake shear zone includes the northern foliation domes that merge with the steeply dipping shear zones to the south. This series of field stops shows variations in the attitude and type of fabrics that occur in the core of the dome at Piseco Lake. Stops 2A to 2D are a driving traverse around Piseco Lake, the type location of the Piseco antiform. At all of these localities, dynamically recrystallized feldspars and quartz form spectacular ribbon- and rod-shaped mineral lineations (McLelland, 1984), in addition to accessory biotite and magnetite. In many places, the alignment of ribbons forms the foliation in this outcrop. Individual quartz-ribbons have aspect ratios upward of 60:1. At Stop 2A, the foliation is weakly to moderately developed, and dips shallowly southward at the eastern part of the outcrop, but is steeply dipping at the western end of the outcrop. The transition between these different foliation attitudes is difficult to determine because the intensity of the foliation is variable. Lineations are penetrative on all scales, and consistent in attitude (Trend: 110°; Plunge: 05°). Stop 6A occurs on the southern flank of the map-scale dome portion of the Piseco antiform. At the western end of the outcrop there are rods of amphibolite (10-30 cm diameter) within the granitic gneiss.

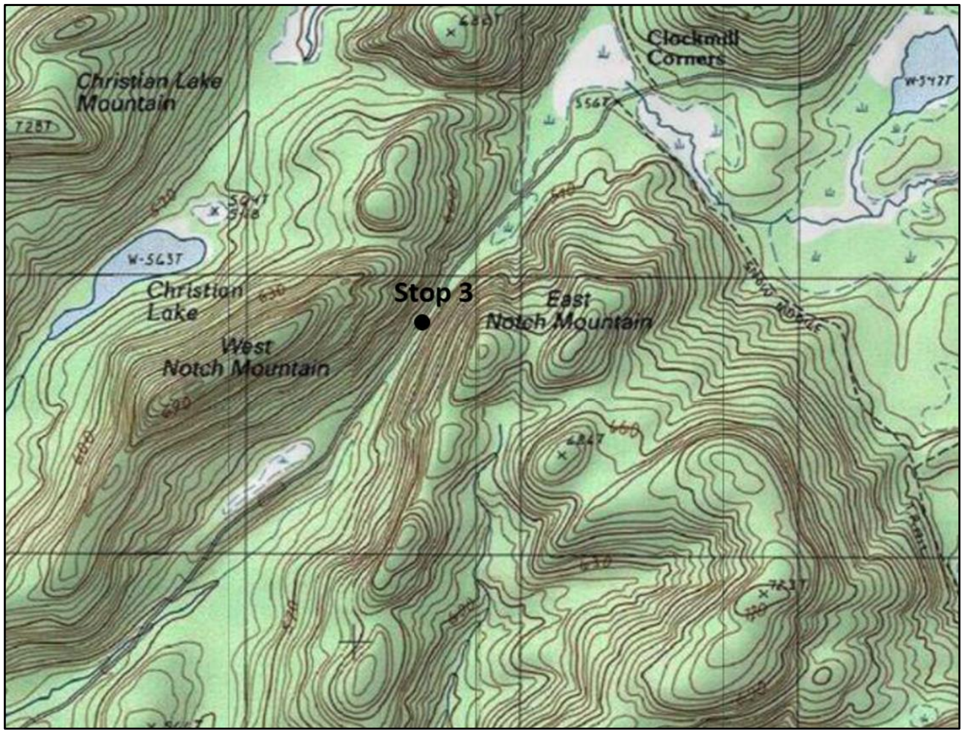


There is an apparent Mesozoic fault that traces down the western side of Piseco Lake and has locally displaced parts of the Piseco dome (Figure 3). At Stop 2B, again the foliation is weakly developed with a penetrative shallowly plunging lineation. However, the foliation, where it can be observed, dips northerly. As you drive southwest along the western side of Piseco Lake, note that the foliation gradually shallows and then dips southerly. There are several outcrops that can be observed where this transition in foliation attitude occurs. The rock fabrics at Stop 2C are similar to those at Stops 2A and 2B, but again the variably developed foliation dips toward the south. Stop 2D is located at the intersection of Rts. 8 and 10, and the folation and lineation is penetrative.

Common to all of these field stops is that biotite and sometimes chlorite blades form microscopic lineations and foliation parallel to the macroscopic structure. Rare grains of hypersthene have been found, but they always have well developed overgrowth textures that include biotite and chlorite. The biotite and chlorite are the most abundant index minerals in the granitic gneiss, and suggest the deformation was last active under low- to moderate- metamorphic conditions, although probably began at much higher conditions to account for the relict grains of hypersthene.



Mileage:
13.3 Turn around and proceed east on Rt. 8 about 0.5 mile. west on Rt. 8.
13.8 Turn north and follow the road around Piseco Lake to Stops 2B and 2C.
22.7 At the intersection with Rt. 8, turn east and proceed about 2.9 miles.
23.6 Turn south onto Rt. 10 and park for Stop 2D. Proceed south on Rt. 10 about 1.2 miles.
24.8 Turn west onto Powley Road (becomes a gravel road) and continue 4.9 miles.
29.7 Park where Powley Road traverses through the Notch.



STOP 3: Steeply dipping mylonite zone of the southern PLsz (18T 530638 m E, 4799018 m N)

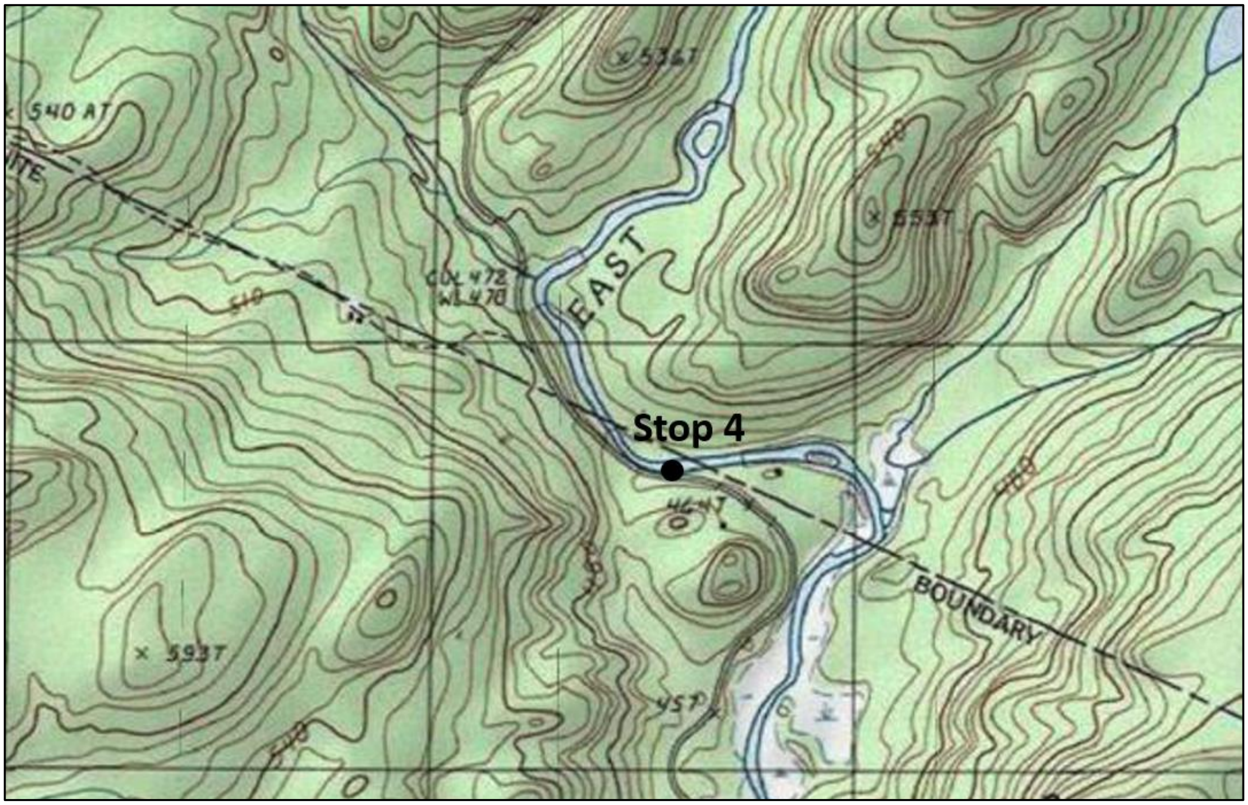
Along Powley Road, depending on the time of year and the amount of road maintenance, there are a sub-continuous series of pavement exposures located in the road bed, as well as in the gutter on the east (northeast-bound) side of the road. Due to the nature of this location, the extent of the exposed rock at this stop changes yearly, so some or all of the rock described here may be viewable depending upon the time of the season in which the stop is visited (best later in the summer). The best exposures occur along the road between West and East Notch Mountains.

All rocks in this region are very similar in mineral content, and vary only in detail with regard to mineral percentage and fabric type and intensity. The rock is dominantly granitic gneiss with intense subhorizontal to shallowly plunging mineral elongation lineation that trends on average about 095 $^{\circ}$, with steeply dipping generally east-west striking foliation. Both fabric elements are defined by ribbons of quartz, and ribbons of aggregate feldspar and quartz (generally 1-5 cm long depending upon grain size). Intensity of the fabric varies across strike at the 50 cm scale, with local layers of significantly coarser-grained fabrics (grains up to 1 cm in diameter). There are also places where the foliation intensity varies as seen at the field stops around Piseco Lake. Rare amphibolite bodies that are 10’s of cm thick occur within the granitic gneiss. Shear sense indicators are abundant in the granitic rocks and consistently show sinistral shear sense.

Mileage:

35.7 Continue south on Powley Road about 6 miles and park. Outcrops are located along the bank of East Canada Creek.

STOP 4: Southern extent of steeply dipping mylonite (18T 527500 m E, 4790961 m N)

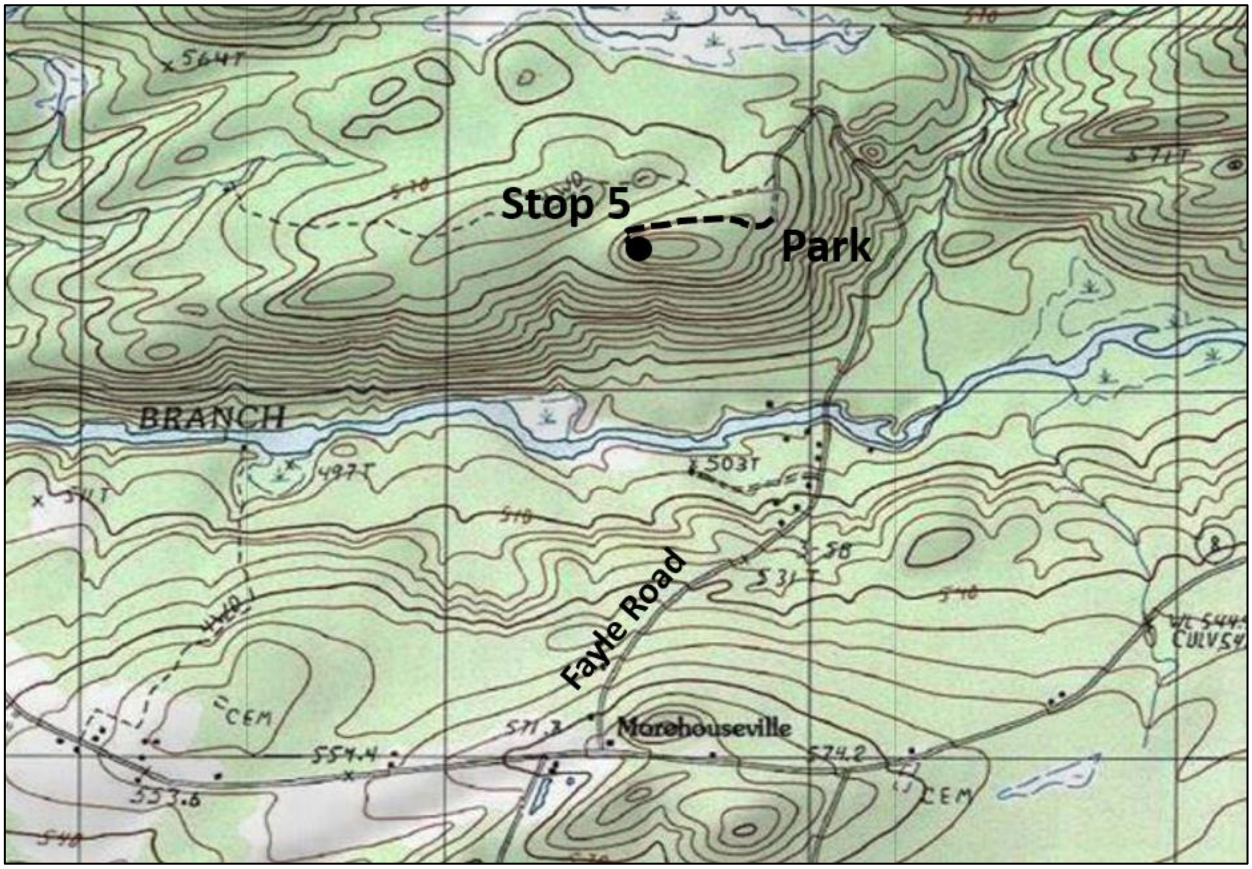


Here the granitic gneiss fabrics contain both a penetrative foliation and lineation. The foliation is steeply dipping and strikes about east-west. Mineral elongation lineation defined by linear aggregates of quartz and feldspar are subhorizontal. The extent of readily available bedrock exposure diminishes south of this location, so this may be the southern-most exposure of the Piseco Lake shear zone. Note that this location is about 21 kilometers across strike from the northern side of the Piseco dome where the pronounced lineation occurs.

Mileage:

46.6 Turn around and proceed back to the intersection of Powley Road and Rt. 10, about 10.9 miles.
47.8 Continue north on Rt. 10 about 1.2 miles to the intersection with Rt. 8.
59.9 Turn west onto Rt. 8 and drive 12.1 miles to Moorehouseville.
61.5 Turn north onto Fayle Road, proceed north. Cross a one lane wood bridge and drive to an opening in the trees at the end of Fayle Road. Park and hike to the west about 350 meters to the west end of the small linear hill.

STOP 5: L>>S and L-tectonite in the core of the Piseco antiform (18T 518563 m E, 4805631 m N)



Excellent outcrops on the northwestern side of a small hill just west of the parking area. Follow the dirt road to a path through the woods, and then head up hill to the south to the outcrops. This outcrop of granitic gneiss contains domains of L>S and L>>S. The L>S domains contain large and numerous σ-type shear sense indicators, some δ-type are present but are much less frequent. The porphyroclasts are large about 1-3 cm and the recrystallized porphyroclastic material is often wrapped with a quartz ribbons. The interpreted shear sense is low-angle and left lateral. The foliation strikes east-west and dips gently to the south. Return to the vehicles by following the path. At this point the trip is over. Retrace your trip to Route 8. If you are headed back to Lake George, then follow Route 8 back to Speculator, NY, then turn south on Route 30.

End of trip.

ACKNOWLEDGMENTS

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REFERENCES CITED

Baird, G. B., and MacDonald, W. D., 2004. Deformation of the Diana Syenite and Carthage-Colton Mylonite Zone: implications for timing of the Adirondack Lowlands deformation. in (Tollo, R. P., Corriveau, L., McLelland, J., and Bartholomew, M. J., eds.) Proterozoic Tectonic Evolution of the Grenville Orogen in North America: Geological Society of America Memoir 197, p. 285-297.

Cannon, R.S., Jr., 1937. Geology of the Piseco Lake Quadrangle. New York State Museum Bulletin, no. 312, 107p.

Carr, S.D., Easton, R.M., Jamieson, R.A., and Culshaw, N.G., 2000, Geologic transect across the Grenville orogen of Ontario and New York: Canadian Journal of Earth Sciences, v. 37, p. 193-216.

Chiarenzelli, J., Hudson, M., Dahl, P., and deLorraine, W. D., 2012. Constraints on deposition in the Trans-Adirondack Basin, Northern New York: Composition and origin of the Popple Hill Gneiss: Precambrian Research, v. 214-215, p. 154-171.

Chiarenzelli, J. R., Valentino, D. W., Thern, E., and Regan. S., 2011, The Piseco Lake Shear Zone: A Shawinigan Suture (abstract): Geological Association of Canada, v. 34, p. 40.

Chiarenzelli J., Regan, S., Peck, W., Selleck, B., Baird, G. and Shrady, C., 2010a. Shawinigan Magmatism in the Adirondack Lowlands as a Consequence of Closure of the Trans-Adirondack Back-Arc Basin: Geosphere, v. 6, p. 900-916.

Chiarenzelli, J., Lupulescu, M., Cousens, B., Thern, E., Coffin, L., and Regan, S., 2010b. Enriched Grenvillian Lithospheric Mantle as a Consequence of Long-Lived Subduction Beneath Laurentia: Geology, v. 38, p. 151-154.

Chiarenzelli, J., Valentino, D., and Gates, A., 2000. Sinistral transpression in the Adirondack Highlands during the Ottawan Orogeny: strike-slip faulting in the deep Grenvillian crust. Abstract presented at the Millenium Geoscience Summit GeoCanada 2000, Calgary, Alberta May 29-June 1st.

Dickin, A.P. and McNutt, R.H., 2007, The Central Metasedimentary Belt (Grenville Province) as a failed back-arc rift zone: Nd isotope evidence: Earth and Planetary Science Letters, v. 259, p. 97-106.

Fakundiny, R. H. 1986. Trans-Adirondack Mountains Structural Discontinuities. In Proceedings of the Sixth International Conference on Basement Tectonics, edited by M. J., Jr. Aldrich, and A. W. Laughlin, pp. 64-75. International Basement Tectonics Association, Salt Lake City, Utah.

Fakundiny, R. H., Yang, J., and Grant, N. K. 1994. Tectonic Subdivisions of the Mid-Proterozoic Adirondack Highlands in Northeastern New York: *Northeastern Geology*, v. 16, p. 82-93.

Gates, A., Valentino, D., Chiarenzelli, J., Solar, G., and Hamilton, M., 2004. Exhumed Himalayan-type syntaxis in the Grenville Orogen, northeastern Laurentia: Journal of Geodynamics, v. 37, p. 337-359.

Geraghty, E. P., Isachsen, Y. W., and Wright, S. F. 1981. Extent and character of the Carthage-Colton mylonite zone, northwest Adirondacks, New York. U.S. Nuclear Regulatory Commission, NUREG/CR-1865, 83 p.

Glennie, J. S., 1973. Stratigraphy, structure, and petrology of the Piseco Dome area, Piseco Lake 15' quadrangle, southern Adirondack Mountains, New York (Ph.D. thesis). Syracuse, New York, Syracuse University, 45 pp.

Hamilton, M.A., McLelland, J.M., and Selleck, B.W., 2004. SHRIMP U/Pb zircon geochronology of the anorthosite-mangerite-charnockite-granite suite, Adirondack Mountains, NY: Ages of emplacement and metamorphism, *in* Tollo, R.P., Corriveau, L., McLelland, J.M., and Bartholomew, M.J., eds., Proterozoic Tectonic Evolution of the Grenville Orogen in North America: Geological Society of America Memoir 197, p. 337–355.

Hanmer, S., Corrigan, D., Pehrsson, S., and Nadeau, L., 2000. SW Grenville Province, Canada: the case against post-1.4 Ga accretionary tectonics: Tectonophysics, v. 319, p. 33-51.

Hawman, R.B. and Phinney, R.A.,1992. Structure of the crust and upper mantle beneath the great valley Allegheny Plateau of eastern Pennsylvania, 1, Comparison of linear inversion methods for sparse wide-angle reflection data, Journal of Geophysical Research, v. 97, p. 371–391.

Hecklau, S., MacKenzie, K., and Chiarenzelli, J., 2014. Geological investigation of Bennies Brook Landslide, Lower Wolfjaw Mountain, Adirondack High Peaks Region (abstract): NE Geological Society of America Abstracts with Programs, v. 46.

Heumann, M.J., Bickford, M.E., Hill, B.M., McLelland, J.M., Selleck, B.W., and Jercinovic, M.J., 2006. Timing of anatexis in metapelites from the Adirondack lowlands and southern highlands: A manifestation of the Shawinigan orogeny and subsequent anorthosite-mangerite-charnockite-granite magmatism: Geological Society of America Bulletin, v. 118, p. 1283-1298.

McLelland, J. M., 1984. Origin of ribbon lineation within the Southern Adirondacks, U.S.A.: Journal of Structural Geology, v. 6, p. 147-157.

McLelland, J. and Chiarenzelli, J., 1989. Age of a xenolith-bearing olivine metagabbro, Eastern Adirondack Mountains, New York. Journal of Geology, v. 97, p. 373-376.

McLelland, J. M. and Chiarenzelli, J. R., 1990. Geochronological studies of the Adirondack Mountains, and the implications of a Middle Proterozoic tonalite suite, *in* Gower, C., Rivers, T., and Ryan, C., eds., Mid-Proterozoic Laurentia-Baltica: Geological Association of Canada Special Paper 38, p. 175-194.

McLelland, J. and Isachsen, Y., 1986, Geologic synthesis of the Adirondack Mountains and their tectonic setting within the Grenville of Canada, *in* Moore, J., Baer, A., and Davidson, A., eds., The Grenville Province: Geological Association of Canada Special Paper 31, p. 75–95.

McLelland, J.M., Selleck, B.W., and Bickford, M.E., 2010. Review of the Proterozoic evolution of the Grenville Province, its Adirondack outlier, and the Mesoproterozoic inliers of the Appalachians, *in* (Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., eds.) From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region: Geological Society of America Memoir 206, p. 1–29.

McLelland, J., Chiarenzelli, J., Whitney, P., and Isachsen, Y., 1988. U-Pb zircon geochronology of the Adirondack Mountains and implications for their geologic evolution: Geology, v. 16, p. 920-924.

Meighan, H. E., Pulliam, J., Brink, U., Lopez-Venegas, A. M., 2013. Seismic evidence for a slab tear at the Puerto Rico trench, Journal of Geophysical Research: Solid Earth, v. 118, iss. 6, p. 2915-2923.

Meighan, H. E. and Pulliam, J., 2013. Seismic anisotropy beneath the northeastern Caribbean: implications for subducting North American lithosphere, Bulletin de la Societe Geologique de France, v. 184, Iss. 1-2, p. 67-76.

Peck, W., Selleck, B., Wong, M., Chiarenzelli, J., Harpp, K., Hollocher, K., Lackey, J., Catalano, J., Regan, S., and Stocker, A., 2013. Orogenic to post-orogenic (1.20–1.15 Ga) magmatism in the Adirondack Lowlands and Frontenac terrane, southern Grenville Province, USA and Canada: Geosphere, v. 9, p. 1637-1663.

Peck W. H, Valley J. W., Corriveau L., Davidson A., McLelland J., and Farber, D.A ., 2004, Oxygen-isotope constraints on terrane boundaries and origin of 1.18-1.13Ga granitoids in the southern Grenville Province. *in* (Proterozoic tectonic evolution of the Grenville orogen in North America, Tollo RP, Corriveau L, McLelland J, and Bartholomew MJ, eds.) Boulder, Colorado, Geological Society of America Memoir 197, p. 163-182.

Piaschyk, D., Valentino, D.W., and Solar, G.S., 2005. Variations in L- and S-tectonite on the northern boundary of the Piseco Lake shear zone, western Adirondack mountains, New York. *in* Valentino, D.W. (ed.), Guidebook for Field Trips for the Annual Meeting of the New York State Geological Association, v. 77**,** Trip B2, 20p.

Price, R., Valentino, D., Solar, G., and Chiarenzelli, J., 2003. Greenschist facies metamorphism associated with the Piseco Lake shear zone, Central Adirondacks, New York (abstract): Northeast Geological Society of America Abstract with Programs, Halifax, Nova Scotia, v. 35, p. 22.

Rankin,D.W.,1976. Appalachian salient and recesses: late Precambrian break-up and the opening of the Iapetus Ocean, Journal of Geophysical Research, v. 81, p. 5605–5619.

Regan, S.P., Chiarenzelli, J.R., McLelland, J.M., and Cousens, B.L., 2011. Evidence for an enriched asthenospheric source for coronitic metagabbros in the Adirondack Highlands: Geosphere, v. 7, no. 3, p. 694-709.

Rivers, T., 2008. Assembly and preservation of lower, mid, and upper orogenic crust in the Grenville Province – Implications for the evolution of large hot long-duration orogens: Precambrian Research, v. 167, p. 237-259.

Rivers, T. and Corrigan, D., 2000, Convergent margin on southeastern Laurentia during the Mesoproterozoic: tectonic implications: Canadian Journal of Earth Sciences, v. 37, p. 359-383.

Roden-Tice, M. K. and Tice, S. J., 2005. Regional-scale mid-Jurassic to Late Cretaceous unroofing from the Adirondack Mountains through central New England based on apatite fission-track and (U-Th)/He thermochronology: Journal of Geology, v. 113, p. 535-552.

Selleck, B.W., McLelland, J.M., and Bickford, M.E., 2005. Granite emplacement during tectonic exhumation: The Adirondack example: Journal of Geology, v. 33, p. 781-784.

Steltenpohl, M. G., Zitez, I., Horton, J. W., Jr., and Daniels, D. L., 2010. New York-Alabama lineament: A buried right-slip fault bordering the Appalachains and mid-continent North America, Geology, v. p. 38, no. 6, p. 571-574.

United State Geological Survey, 2017. Magnetic Anomaly Maps and Data for North America, Mineral

Resources Online Spatial Data, https://mrdata.usgs.gov/magnetic/

Valentino, D., Chiarenzelli, J., Hewitt, E., and Valentino, J., 2012. Applications of water-based magnetic gradiometry to assess the geometry and displacement for concealed faults in the southern Adirondacks Mountains, New York, U.S.A.: Journal of Applied Geophysics, v. 76, p. 109-126.

Valentino, D., Chiarenzelli, J., Paaschyk, D., Williams, L., and Peterson, R., 2008. The Southern Adirondack Sinistral Transpressive Shear System: *in* Friends of the Grenville (FOG) Field Trip 2008, D. Valentino and J. Chiarenzelli (eds.), September 28th, Indian Lake, New York, 56 p.

Valentino, D. W., Chiarenzelli, J. R. and Regan, S., 2019. Spatial and temporal links between Shawinigan accretionary orogenesis and massif anorthosite intrusion, southern Grenville province, New York, U.S.A., Journal of Geodynamics, v. 129, p. 80-97.

Wiener, R.W., 1983. Adirondack Highlands-Northwest Lowlands ‘boundary’: A multiply folded intrusive contact with associated mylonitization: Geological Society of America Bulletin, v. 94, p. 1081-1108.

Wong, M. S., Peck, W. H., Selleck, B. W., Catalano, J. P., Hochman, S. D., and Maurer, J. T., 2011, The Black Lake shear zone: A boundary between terranes in the Adirondack Lowlands, Grenville Province, Precambrian Research, v. 188, p. 57-72.

Trip A-4

OUTDOOR EARTH SCIENCE- A GEOLOGIAL/ECOLOGICAL NATURE TRAIL FOR STUDENTS OF ALL AGES

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ABSTRACT

The field of geosciences addresses a wide range of Earth processes and materials. Individuals of all ages tend to learn about these topics in much the same way, regardless of their grade or level of education. These concepts are best understood when learners can physically see and interact with these processes and materials in their natural environments. We have taken this approach to develop an immersive, geoscience-themed learning experience for visitors to Rice Creek Field Station (RCFS) in Oswego, NY. Participants will hike an approximately 2-mile long ‘geology trail’ that explores various sites of geological interest, giving them an opportunity to see and interact with the environment both literally and virtually. At each stop, participants will be able to see and feel the varied textures of different types of rock, walk the change in topography left behind by glaciers, watch the movement and flow of water eroding sediment, and see how the land has been shaped by natural and human processes over thousands of years. Maps and placards at each stop will provide information and help guide participants navigate the trail; scannable QR codes linked to websites support a more in-depth understanding of fundamental concepts, and student research taking place at each location. This supports a more complete learning experience and provides greater context of the science for learners of all levels.

RCFS is used primarily by students and faculty of Oswego State University, however, many other local groups and residents regularly make use of the trails and facilities. This presents a valuable opportunity to engage with the public and provide science outreach for adults and students in the area. This project has been developed and maintained by undergraduates who are able to use this as a forum to preserve and communicate their research on local geology to individuals who frequent the park, while educating them about fundamental scientific principles. It also provides K-20 Earth science teachers a destination to share outdoor field experiences with their classes, while reinforcing objectives of state and national science curricula. We provide a simple model for integrating science outreach and education in an outdoor setting that can be easily adapted to many different local settings.

INTRODUCTION

Rice Creek Field Station (RCFS) is a public space dedicated to research and education in the natural sciences to individuals of all ages, which is of particular interest and use to individuals in the Oswego area. Many examples of geological materials and processes that define Central New York are on display at Rice Creek, we will explore some of these during the walk around the property. Glacial and fluvial deposits dominate the surface materials in Central New York (fig. 1). Glacial till is the major surface material, in addition to kame and glacial sands. Lacustrine sands and clays are also minor components of the surface in the area due to the formation of proglacial lakes and streams. In addition to the abundant materials indicating the influence of glacial activity, there is also striking geomorphological evidence of glacial advance preserved in the topography of the region (fig. 2). Small hills, or drumlins dot the landscape south of Lake Ontario across a large swath of Central New York. RCFS encompasses one of these drumlins, allowing park goers to observe the scale of sediment deposits formed by the movement of ice during the last glacial maximum. RCFS has an extensive trail system that traverses the entire park (fig. 3), allowing visitors to explore various types of geologic materials (rocks, sediment, water) and features (hills, streams, water bodies) formed by natural and human interactions.

While observing these features at RCFS is an effective and educational experience- this area is far from unique, and these features and materials are not quite “world class”, however they still provide an ideal environment for learning about basic geologic concepts that often can befuddle students from junior high school through a freshman year in college. Often, there is a great deal of overlap between the content of course work for students aged from 10 to 21, making immersive experiences like this beneficial to a large swath of the population. Further, non-students that are part of the general population who frequent locations such as RCFS can also benefit from outreach programs that employ this hands-on, outdoor, educational approach. It is the goal of this field trip to not simply educate, but encourage educators to develop similar experiences for students of all ages in areas (parks, preserves, campuses) convenient and relevant to your population.

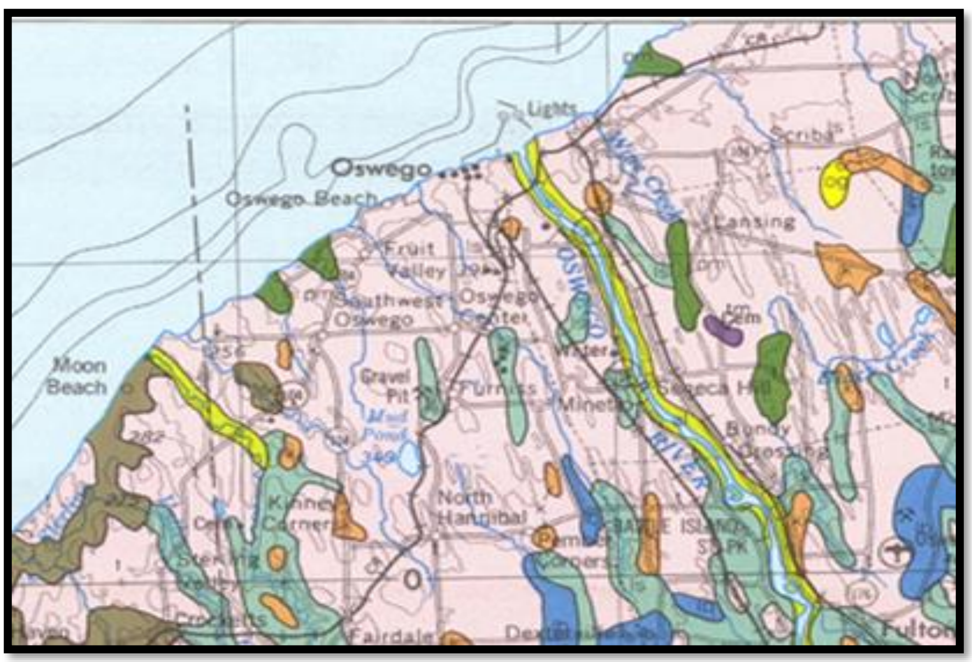

Figure 1. *Geologic surface map (NYS) showing the types of sedimentary deposits found in the region surrounding RCFS. Most abundant is glacial till (pink) and lacustrine sands and silts (teal/brown). After Muller et al. (1985).*

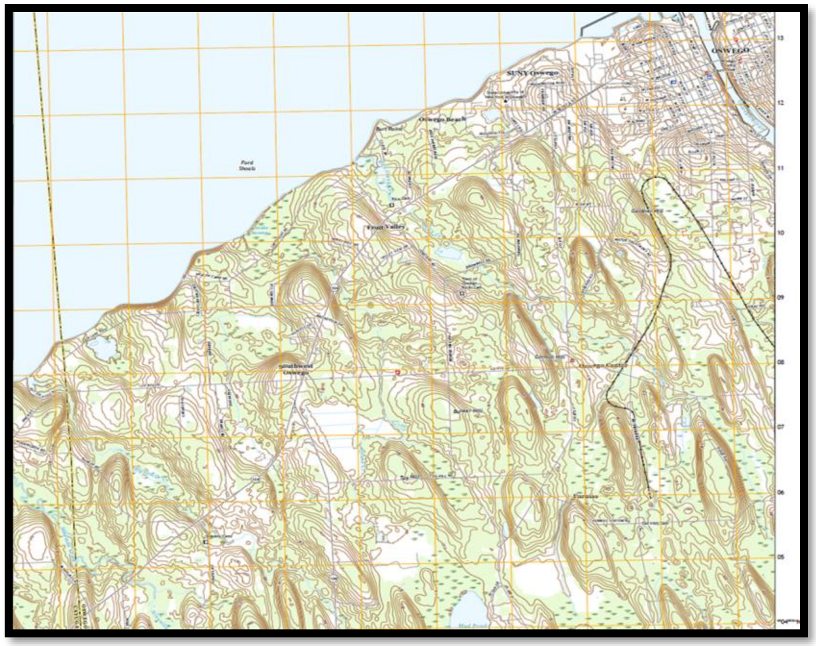

Figure 2. *Topographic map of the area, showing abundant drumlins. Oswego West, USGS (1997)*

Figure 3. *Trail map of Rice Creek Field Station with nature trail stops noted. Used from DiFrancesco (2022).*

GEOLOGIC NATURE TRAIL AROUND RICE CREEK FIELD STATION

The following is an approximately 1.75 mile moderately strenuous walk around the RCFS in Oswego, NY. Parking is free, but there are limited spaces. Overflow parking is available along the roadside. Bathroom facilities are also available in the Field Station building.

WALKING LOG

The group will meet at the pavilion outside of the Rice Creek Field Station main building (adjacent to the traffic circle). This is within sight of the trailhead for the Red Trail- where we will begin our walk.

STOP 1: Bottom of the drumlin

Proceeding north along the red trail, away from the pavilion the group will walk approximately a quarter mile until reaching the intersection with the orange trail. At this point, we will bear to the right and walk up the orange trail. Here, we can notice the trail steepen- this is the edge of the drumlin. When you look up the trail you can see that you are looking up the slope of it (Figure 4).

Figure 4. *Standing at the bottom of the drumlin, the trail takes a noticeable climb up the slope of the hill.*

STOP 2: Meta/igneous erratic

Here just off the trail, there is an erratic made of granite. We will take the opportunity to discuss the fact that an erratic is a rock or boulder that differs from the surrounding rock and is believed to have been brought from a distance by glacial action. We will discuss the differences between the sedimentary rock that makes up the bedrock of most of Central NY (fig. 5), and contrast that with the magma that formed this rock. It is notable that this is different from the regular glacial sediment, as these erratics are by definition quite large and stick out, as this one does. In addition, we can discuss the composition- the minerals in it are, potassium feldspar (pink), quartz (white/clear), and biotite mica and hornblende (black).

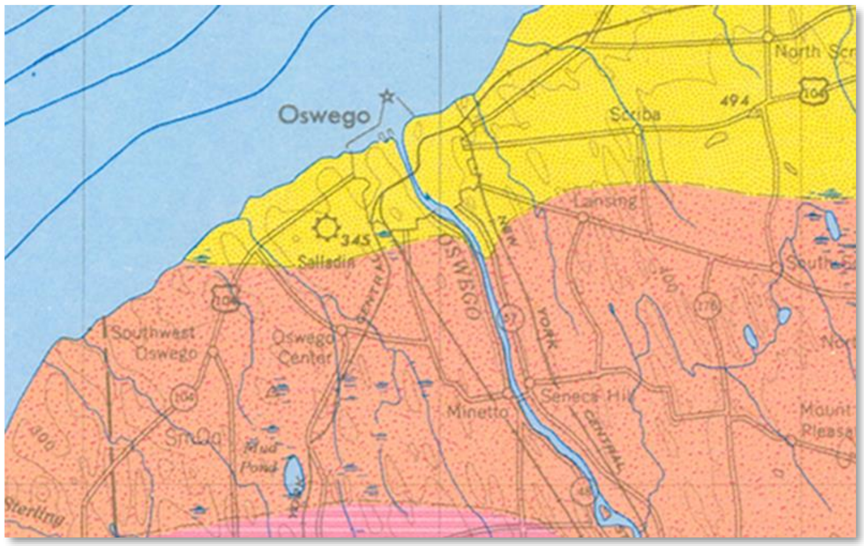

Figure 5. *Geologic bedrock map of the area surrounding RCFS. Note the abundance of sandstone, siltstone and shale- all sedimentary rocks. Rickard et al. (1970)*

STOP 3: Granitic erratic

This erratic is a granite, which is an igneous rock. The minerals in it are potassium feldspar (pink), quartz (white/clear), and biotite mica (black). Note the difference in color and texture from the previous stop. It is clear that these erratics are from different sources, in addition to being distinct from the sedimentary rock forming the bedrock in this area.

STOP 4: Large intermediate meta-igneous erratic

Here we have another erratic. This one is significantly bigger than the other two and is what's known as a diorite gneiss. The sodium plagioclase (white), and hornblende and biotite (black). The minerals in this rock are substantially darker than the other granitic rocks we’ve observed. This demonstrates the variability in composition of many of the erratics that are deposited in RCFS. We can also note the variable texture of this rock (pegmatite veins), suggesting interaction with fluids (Figure 6).

Figure 6. *Student examining the large glacial erratic at stop 5.*

STOP 5: Top of the drumlin

This stop is the top of the drumlin. At this point this surface was directly in contact with the glacier. At that point you would be standing under nearly 2 miles of ice. After the glacier ablated, the melt water flooded the area, creating glacial Lake Iroquois. This lake eroded and flattened the tops of many of the drumlins closest to Lake Ontario today.

STOP 6: Seasonal stream

Walking down the side of the drumlin, we see evidence of groundwater infiltration as well as surface runoff. Seasonal streams such as these running along and beside the trail help to gradually erode and wear down the surface. This is a visual representation of a large scale process operating at a very small scale. Also observable is particle sorting- small particles of clay are carried further than larger particles of silt, and sand. The effect is more dramatic directly after heavy rainfall, but it is useful to illustrate these processes.

STOP 7: Bog

A bog is a type of wetland that is too soft to support a heavy body. Walking off the trial, you will sink a little bit. A bog is also known for the accumulation of peat. Peat is a deposit of dead plant material, and another major constituent of the local surface geology (fig. 2 – peat and muck). It can be discussed that this peat gets buried and lithified (compressed into a rock) it forms coal.

STOP 8: Rock Wall

Proceeding down the orange trail to the intersection with the blue trail, we come across one of many rock walls that are constructed throughout RCFS (fig. 7). Notice, the majority of the rock forming the wall is flat, squared off sandstone/siltstone- this is the Oswego formation, and makes up the majority of rock in the area.

In contrast, there are occasional rounded stones that are made of granite, gneiss, or some other igneous/metamorphic rock. These rocks are very different from the sandstone, and they each have very different formational histories.

Figure 7. *Rock wall along the blue trail. Note the abundance of flat sedimentary rocks, with only a few igneous/metamorphic rock that has been more rounded.*

STOP 9: Bridge over stream and vernal pool

This section of the blue trail is damp, requiring a bridge likely because the surface is quite close to the water table. This area is also home to a vernal pool (or ephemeral pool) that forms during the rainy season. These are temporary bodies of surface water that provide sanctuary to different species of plants, insects, and animals. It should be understood that these pools form as a consequence of the local geology and environmental conditions which can vary throughout the year.

STOP 10: Meandering stream

Another stream draining off the drumlin, and flowing along the flat lowland along the base of the hill. This stream clearly shows a meander- it is a single channel, but we can observe that it winds back and forth, following the route of least resistance (fig. 8). Closer examination of the stream channel, we can see sorting of sediment, grading from coarser to finer moving closer to the shoreline. Point bars can be identified as well, forming as a consequence of the stream flow.

Figure 8. *Small stream running through RCFS showing a characteristic meander as it flows onto more level terrain.*

STOP 11: Rice Creek

Here we follow the blue trail down to the deep, wide channel of Rice Creek for which the field station is named. Walking upstream, we see the flowing water has eroded much of the loose, finer, sediment from its channel, exposing more of the cobble to boulder-sized till. These rocks are continually being mechanically and physically weathered by the movement of the creek. This water flowing out of Rice Creek will eventually enter into Lake Ontario as a minor tributary. Walking along the creek, it may be possible to visit the artificial dam that forms Rice Pond.

STOP 12: Rice Pond

Rice Pond is artificially formed by the dam at the northern edge of the pond. Rice Pond conforms to the local topography and geology of the area as Rice Creek has flooded the lowest points of the valley. Proceeding further south, the pond transitions into another large swamp/bog. Note that the flow has slowed considerably, but not stopped. This allows for deposition of sediment that has been transported from further upstream. Walking along the shoreline of the pond will lead back to the traffic circle and the start of the walk.

Figure 9. *Rice pond, partially frozen over during the winter of 2021. Here we see the center of the pond is not frozen, suggesting there is still some gradual flow through the channel.*

End of trip.

ACKNOWLEDGMENTS

The geology walking tour of Rice Creek has been developed and revised over several years by many undergraduate researchers and teaching assistants for use with various lab classes as well as youth, school, and community groups with the goal of educating the public about the natural history of their community. I am grateful for the input and participation of all of these individuals. I am also grateful to Rice Creek and Rice Creek Associates for their continued support of this venture and other associated research.

REFERENCES CITED

DiFrancesco, N., Neglia, A., Ramparsad, J Bratage, K. (2022) A Path for Geoscience Outreach: A Geology Walking Trail for Teaching Local Natural History to Students of all ages. Geological Society of America Abstracts with Programs. Vol. 54, No. 3

Muller, E.H. and Cadwell, D.H., 1985. Finger Lakes Sheet. Surficial Geologic Map of New York.

Rickard, L.V. and Fisher, D.W., 1970. Finger Lakes Sheet. Bedrock Geologic Map of New York.

Trip B -1

DETRITAL ZIRCON GEOCHRONOLOGY OF THE ADIRONDACK LOWLANDS

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ABSTRACT

During this day-long trip we will visit and discuss several sites in the Adirondack Lowlands for which detrital zircon geochronology is available. Stops will include quartz-rich protoliths from throughout the upper Amphibolite facies stratigraphic succession including the turbiditic sequence at Pyrites, NY, sandy and tourmaline-rich intervals in the Lower marble, sandy, calc-silicate interlayers within the Popple Hill gneiss, and base, middle, and top of the Upper Marble. The data collected to date has important implications for the age of metasedimentary rocks deposited before Grenville orogenesis, shifting provenance in response to tectonism, the events leading to the formation of the world-class Pb-Zn sedimentary exhalative deposits, the timing and intensity of metamorphism in the Lowlands, and potential basement rocks. Much of this information is summarized from Chiarenzelli et al. (2015a; 2017); some is a summary of on-going work.

GEOLOGICAL SETTING

The contiguous Grenville Province forms an extensive part of the Canadian Shield added to the southern and eastern margin of Laurentia (present coordinates; Figure 1) during a sequence of orogenic events occurring during the Mesoproterozoic (ca. 1600-1000 Ma). McLelland et al., (1996) summarize the orogenic event effecting the Adirondack Region. As one of the last major continent building events during the Precambrian, associated tectonism reworked, and incorporates, rocks of significantly older parts of the Canadian Shield including the Archean Superior Province and Paleoproterozoic terranes (Dickin et al., 2016; Figure 1). In a global context, the end of the Mesoproterozoic was a time of world-wide orogenesis resulting in the formation of the supercontinent Rodinia; hence orogenic rocks of this age occur globally.

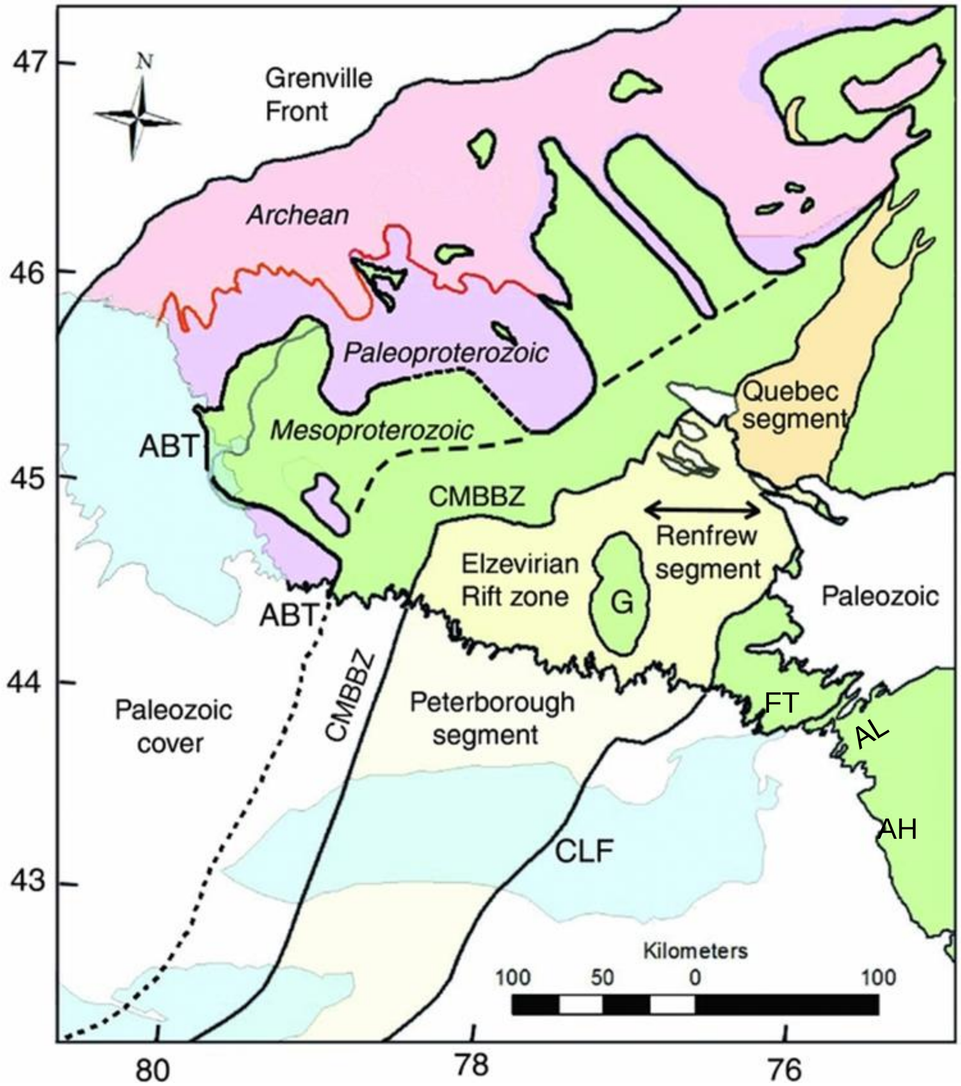


Figure 1. *Regional map of the southwest Grenville Province (from Dickin et al. 2016). Abbreviations are as follows. ABT – Allochthon Boundary Thrust; AH – Adirondack Highlands; AL – Adirondack Lowlands; CMBBZ – Central Metasedimentary Belt boundary zone; CLF – Clarendon–Linden fault; FT – Frontenac Terrane; G – Grimsthorpe domain.*

The Adirondack Region is physically connected to the remainder of the Grenville Province via the Frontenac Axis or Arch (for summary of the geology see Chiarenzelli and Selleck, 2016). Several fundamental lithostructural and tectonic terranes occur in the Adirondacks including, from northwest to south: the Frontenac Terrane, the Adirondack Lowlands, the Adirondack Highlands, and the Southern Adirondack Terrane (Figure 2; Valentino et al., 2019). The metasedimentary sequence in the Adirondack Lowlands is dominated by calcitic and dolomitic marbles and calc-silicate gneisses, and includes meta-evaporitic units and metamorphosed clastic rocks consisting of quartzites, pelitic gneisses, meta-arkoses, meta-greywackes and meta-siltstones. As part of orogenic activity between 1250-1150 Ma (Elzevirian and Shawinigan orogenies) in the southern Grenville Province, these metasedimentary rocks are deformed and metamorphosed to mid-upper Amphibolite facies. They are part of a larger region of the southern Grenville Province, characterized by a thick, marble-rich supracrustal succession, named the Central Metasedimentary Belt (Wynne-Edwards, 1972). Similar metasedimentary rocks can be found in the Adirondack and New York-New Jersey Highlands (e.g. Peck et al., 2009) and adjacent areas. However, in contrast to most other areas in the Adirondacks and elsewhere in the Grenville Province, the Lowlands stratigraphic succession, while deformed and intruded, is relatively intact and well described due to the world-class zinc deposits it hosts (Figure 3). The stratigraphy of the Upper Marble, in particular, is well documented from thousands of diamond drill cores obtained from the Sylvia Lake syncline and other nearby zinc deposits hosted in the Upper Marble (deLorraine, 2001; deLorraine and Sangster, 1997).

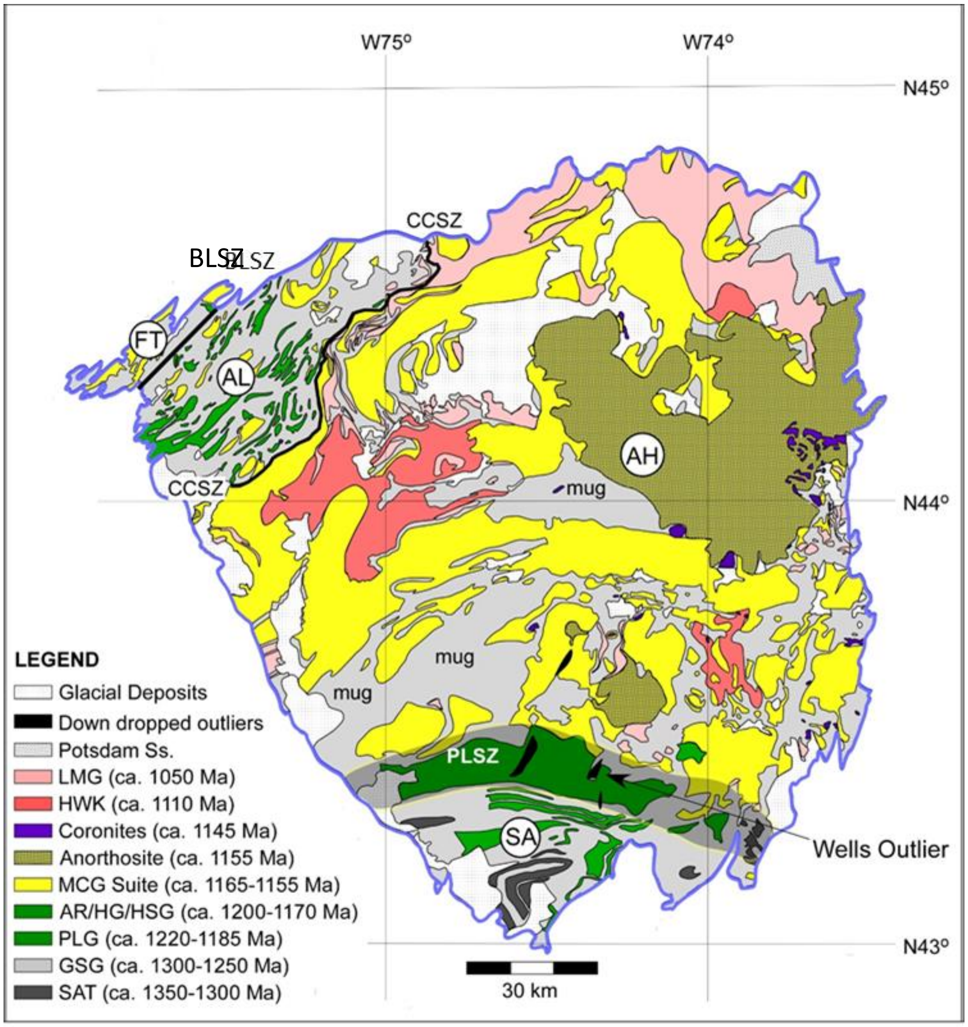


Figure 2. *Simplified geological map of the Adirondack Region after Chiarenzelli and Selleck (2006). Abbreviations of lithologic units: AR/HG/HSG – Antwerp Rossie, Hermon Granite, Hyde School Gneiss; GSG – Grenville Supergroup; HWK – Hawkeye Suite; LMG – Lyon Mountain Granite; MCG – Mangerite, Charnockite, Granite Suite; PLG – Pisceo Lake Granite; SAT – Southern Adirondack Tonalites; mug – supracrustal rocks extensively intruded, migmatized, and disrupted. Abbreviations of Terranes and Terrane Boundaries: AH – Adirondack Highlands; AL – Adirondack Lowlands; BLSZ – Black Lake Shear Zone; CCsz – Carthage-Colton Shear Zone ; FT – Frontenac Terrane; PLSZ – Piseco Lake Shear Zone; SA – Southern Adirondacks.*

With one likely caveat, to be discussed later, the oldest, known rocks in the Lowlands are metasedimentary rocks of what was known regionally as the Grenville Supergroup. In the Lowlands they are intruded by several metaigneous suites ranging in age from 1200-1150 Ma and thus must have been deposited before 1200 Ma (Peck et al; 2012); the time associated with the beginning of the Shawinigan Orogen (Corrigan, 1995; ca. 1200-1140 Ma). It is also possible that metasedimentary rocks in the Adirondack Lowlands where deformed during the Elzevirian Orogeny (ca. 1245-1220 Ma), as field relations in the Lowlands indicate isoclinal folding occurred before intrusion of the oldest Shawinigan intrusive suite – the ca. 1200 Ma Antwerp-Rossie granitoids (Chiarenzelli et al., 2010b). If so, this would indicate the deformational effects and metamorphism associated with the Elzevirian Orogeny, defined in the Central Metasedimentary Belt of Ontario, may extend farther to the south than previously confirmed.

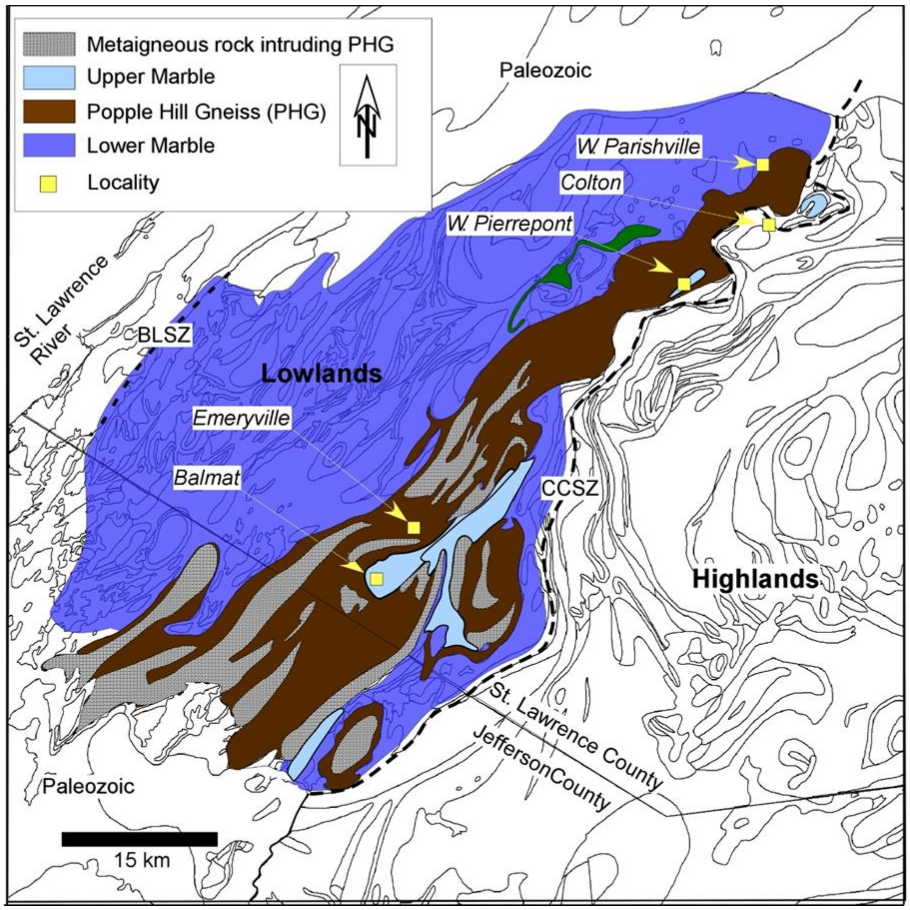


Figure 3. *General geologic map of the Adirondack Lowlands, St. Lawrence County, New York.*

STRATIGRAPHY OF THE ADIRONDACK LOWLANDS

Several previous versions of stratigraphy of the Adirondack Lowlands and, in some cases, Highlands, have been proposed in the literature and summarized in Wiener et al. (1984). Most are based on regional and local field studies completed before the widespread availability of geochronological data and constraints derived therefrom, modern structural interpretations, and information gleaned from recent zinc exploration. While many follow the general sequence noted below, some, such as that proposed by Wiener et al. (1984) have significant departures.

Modern analysis suggests that the Lowlands stratigraphic package is allochthonous as basement rocks are currently unknown or unrecognized. From bottom to top, the package includes the Lower Marble, Popple Hill Gneiss, and Upper Marble, which has a preserved thickness on the order of 2-3 kilometers (Figure 4). The Upper Marble, host to the Pb-Zn sedimentary exhalative deposits, has been subdivided into 16 units, largely from insight from the thousands of diamond drill cores penetrating the Sylvia Lake Synform near Balmat, New York in search of ore (deLorraine, 2001; deLorraine and Sangster, 1997).

The Upper Marble is the most spatially restricted of the three major stratigraphic units in the Lowlands, occurring in a discontinuous belt of rock extending from Balmat to Seven Springs, northeast of Colton. It also has the best preserved stratigraphic relations with earlier workers recognizing natural gas and evaporitic units similar to those found in much younger petroleum settings (Brown, 1932). Zinc mining in the Lowlands has occurred only in this unit and along most of the belt from Balmat to Edwards to West Pierrepont. It should be noted that zinc has not been mined in the Lower Marble or in marble units the Adirondack Highlands. The rocks within the Upper Marble consist primarily of dolomitic marble, siliceous marbles, and calc-silicate rocks with subordinate pelitic, evaporitic, and quartz-rich lithologies. Compilation of thousands of diamond drill cores in the Sylvia Lake synform, a ca. 10 km-scale over turned fold, has led to the recognition of 16 units and three evaporitic-ore cycles. Evaporites consist primarily of thick accumulations of anhydrite, presumably once gypsum, but halite is also found. A wide range of minerals with unique or different chemistry are found throughout the Balmat area making it a prime location for the identification of new or rare minerals, and mineral collection in general (Chamberlain et al., 2018).

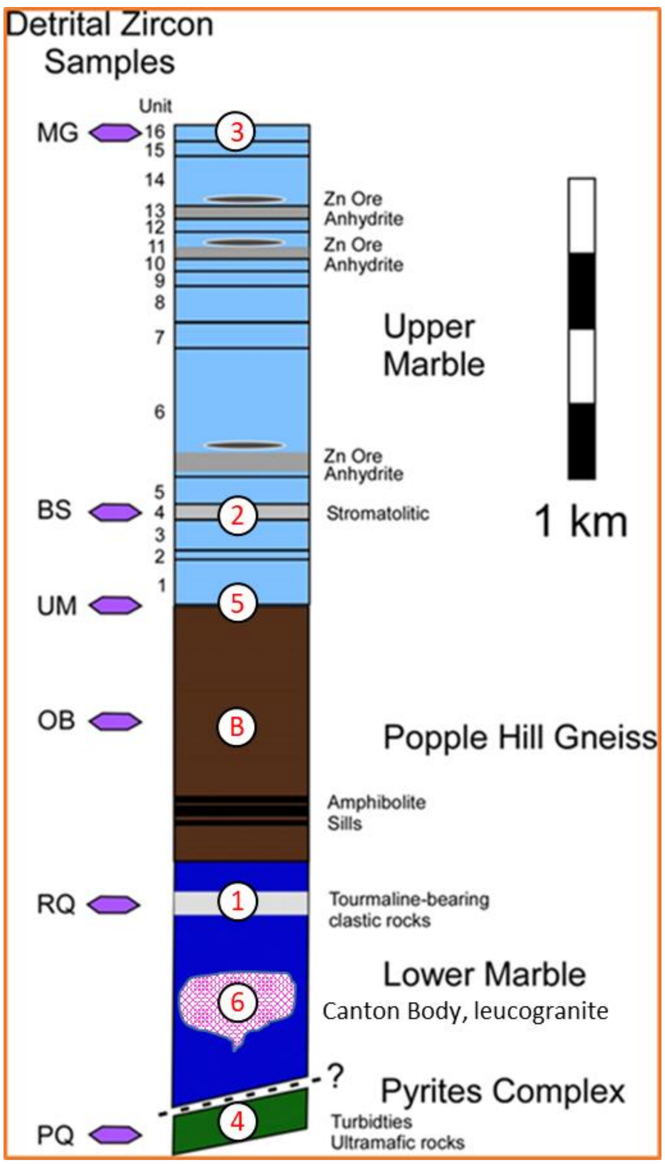


Figure 4. *Simplified diagram of the Lowlands stratigraphy after Chiarenzelli et al. (2017), deLorraine (2001), and deLorraine and Sangster (1997). Circles with red numbers/ letter correspond to field trip stops. Abbreviations: B – Bonus Stop (if time permits); BS – Balmat stromatolites; MG – Median gneiss; OB – O'Brien Road Popple Hill gneiss; PQ – Pyrites quartzite; RQ- Richville quartzite; UM – Upper Marble. Leucogranite of the Canton Body (STOP 6) shown schematically intruding Lower*

The Popple Hill Gneiss is primarily composed of medium-grained biotite-plagioclase-quartz gneisses (Chiarenzelli et al., 2012) occasionally containing garnet and/or sillimanite, particularly where highly strained. While various interpretations have been made regarding its ultimate protolith including volcanic rocks (e.g. Carl et al., 1988), the most likely scenario appears to be sandstones, siltstones, and mudstones deposited in a deep water basin. The base of the unit is often marked by regionally extensive pyritic gneisses once mined for sulfur. The bottom third or so of the unit is extensively intruded by amphibolite, gabbro, and mafic-ultramafic sills, while the upper portion generally lacks them. Towards the top of the unit, mm-cm-scale quartzite and dcm thick marble interlayers occur as the Upper Marble is approached. In contrast to the Lower Marble, intruded largely by the Antwerp-Rossie suite, the Upper Marble is extensive intruded by the coarse-grained Hermon Granite gneiss and has undergone extensive migmatization during the Shawinigan event (Heumann et al., 2006). It is uncertain if the Popple Hill gneiss is in sedimentary and/or structural contact with the Lower Marble but could have been, at least in part, synchronously deposited in a deeper parts of the same basin.

The Lower Marble consists of calcitic and dolomitic marbles, calc-silicate rocks dominated by tremolite or diopside, schists and gneisses (including talc-tremolite, tourmaline, and graphitic and schists and pyrite+/-graphite-rich pelitic gneisses), and minor quartzite and meta-arkosic interlayers. Zinc mineralization and evaporitic lithologies are generally absent, although minerals such as scapolite and tourmaline, likely associated with evaporitic conditions, are frequently found in specific stratigraphic intervals. Wiener et al (1984) suggested that a single marble unit (Gouverneur Marble) was present in the Lowlands and subdivided it into five units (A through E), but variable lithologies, regional variation, extensive ductile deformation and structural cutouts, disruption by later plutonic suites, and less detailed exploration / investigation has thus far limited stratigraphic resolution.

The contact between metasedimentary rocks of the Adirondack Lowlands and ultramafic rocks of the Pyrites Complex (Figure 5) are exposed along the Grasse River at Pyrites, New York (Chiarenzelli et al., 2010a; 2011). Here regionally extensive pyrite and graphite-rich gneisses overlie ultramafic rocks consisting of hydrothermally altered periodite, pyroxenite, and rare lamprophyre dikes. Lying immediately above the peridotite and below the pyrite and graphite-rich gneisses are inclined layers of metamorphosed turbiditic rocks with cm-scale sand-silt-mud packages, now quartzite to garnet-biotite-sillimanite gneiss folded about isoclinal hinges. The nature of their contact with the ultramafic rocks is ambiguous the supracrustal rocks becoming more mafic over several meters as the peridotite is approached, perhaps suggesting a structurally modified, depositional contact. The ultramafic rocks from Pyrites lie on a Sm-Nd isochron of 1442±120 Ma (Chiarenzelli et al., 2010a), indicating they are older than all other rocks thus far identified in the Lowlands.

Although small in aerial extent (1-2 km^2), a gravity survey indicates significant subsurface extension of the Pyrites Complex to the southwest (Revetta and McDermott, 2003). The exposure of the ultramafic rocks occurs within the core of a folded layer of amphibolite aligned parallel with the regional northeastern lithological and structural trend. This is one of many elongate linear, discontinuous amphibolite-dominated lithologic bands associated with pyritic gneisses throughout the Lowlands (Prucha, 1957), extending more than 50 km within close proximity, and parallel to the Carthage-Colton shear zone. The association of metamorphosed mafic, hydrothermally altered ultramafic, and chemogenic metasedimentary rocks with shallow water metasedimentary rocks of the Lowlands sequence, including vast volumes of marble, suggests tectonic interleaving of a suprasubduction or back-arc ophiolite complex during obduction associated with the Shawinigan Orogeny (Chiarenzelli et al., 2010a; 2011).

Figure 5. *Left. Peridotite (above) in contact with knobby weathering pyroxenite, exposed along the Grasse River at Pyrites. Hammer spans the contact. Right. Tectonic blocks of dark green ultramafic blocks within mustard yellow pyrite-rich gneisses along the Grasse River in Pyrites, New York. Two poorly balanced geologists for scale.*

DETRITAL ZIRCON GEOCHRONOLOGY

Our study of detrital zircons targeted quartz-rich units in each of the major metasedimentary units in the Lowlands. Previous work by Heumann et al. (2006) revealed that migmatitic rocks of the Popple Hill gneiss contain a bimodal population of zircons, with one population, derived primarily from leucosomal portions of the rock, yielding ages of ca. 1160-1180 Ma. This age overlaps with numerous metaigneous rock suites in the Lowlands and Highlands and falls within the recognized time frame of the peak metamorphic conditions associated with the Shawinigan Orogeny. Thus, these zircons were interpreted as anatectic in origin, in contrast to those yielding older ages, ca. 1300-1400 Ma, interpreted as detrital in origin. A similar distribution of zircon ages was found by Bickford et al. (2008) in the eastern Adirondack, but there, 1050 Ma anatectic zircons dominate the histogram (Figure 6).

In our samples, many detrital zircons showed thin rims (ca. 10-20 microns) or outer areas of recrystallization. Most of these were of limited width and volume precluding direct analysis; however, they may have likely played a role in some of the “hybrid” ages found, adding some uncertainty to the results and conclusions reached. However, in comparison to zircons from quartzites subjected to granulite facies in the southern Adirondacks (Peck et al., 2010), recrystallization appears to be relatively minimal (Figure 7).

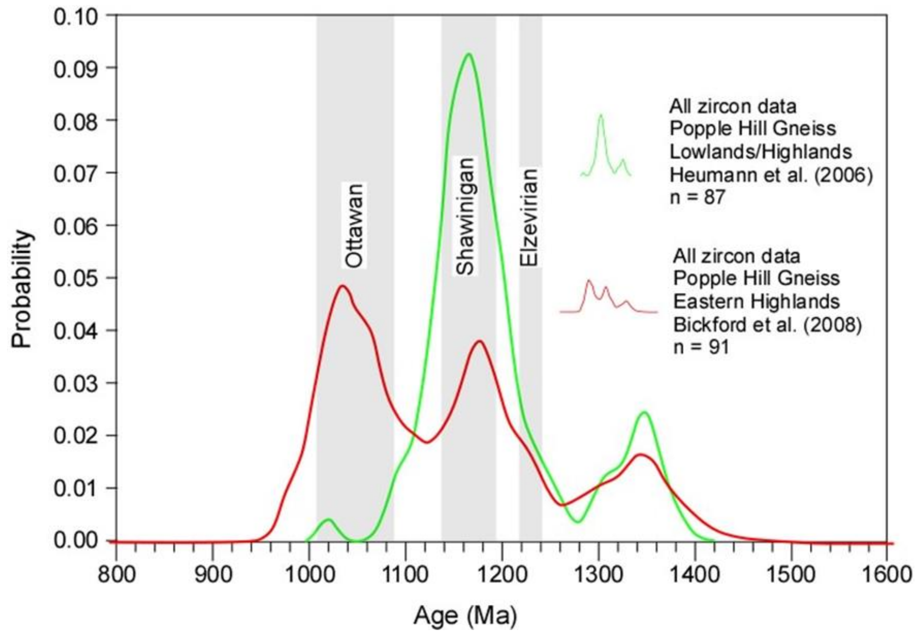


Figure 6. *Relative probability histogram utilizing the data of Heumann et al. (2006) and Bickford et al. (2008) for the Popple Hill Gneiss and inferred Highlands equivalents (from Chiarenzelli et al., 2012).*

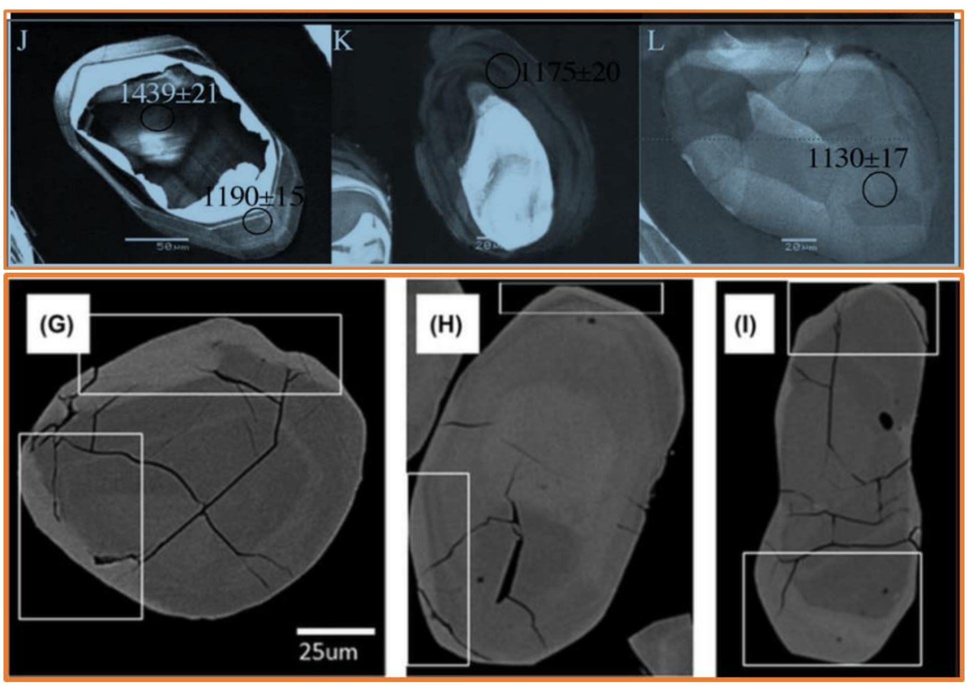
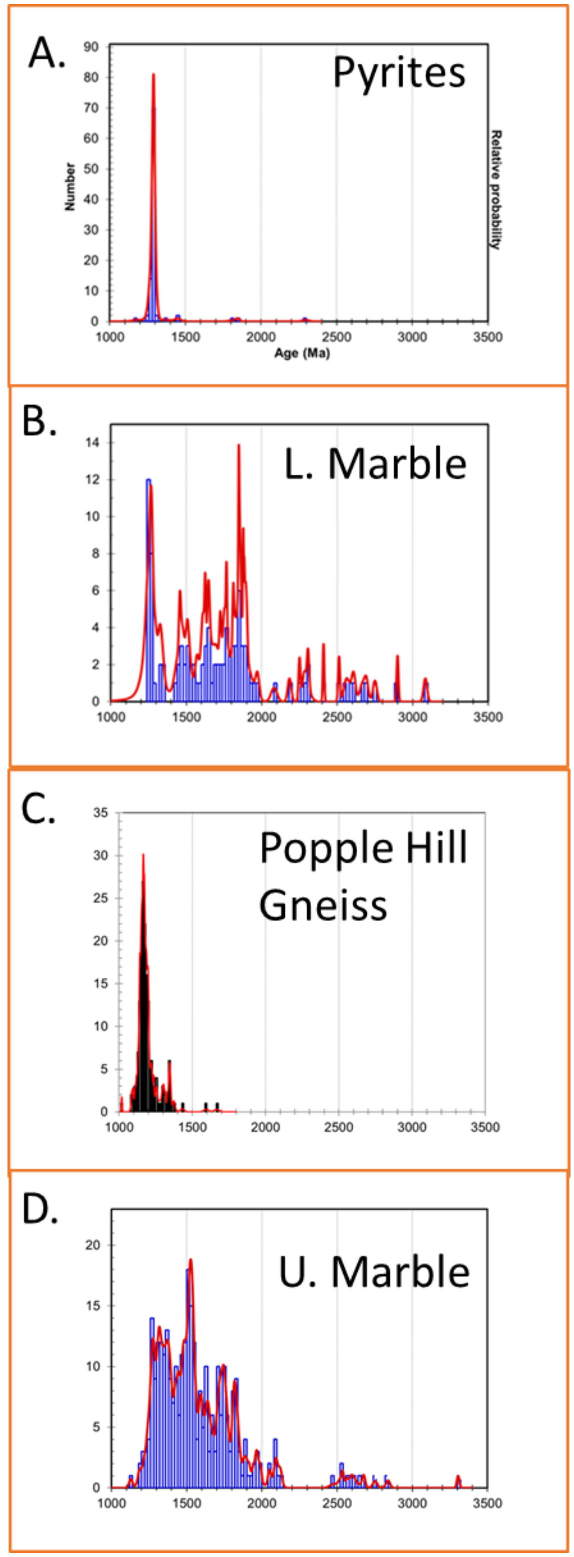


Figure 7. *Scanning electron microscope images of zircons from Adirondacks quartzites. Top row: Granulite facies zircons imaged by CL from the southern Adirondack Irving Pond quartzites (from Peck et al., 2010). Bottom row: Upper Amphibolite facies zircons images by BSE from Lowlands quartzites (from Chiarenzelli et al., 2017). Note the difference in outer rim development, presumed to be related to recrystallization during metamorphism. Thickest rims highlighted in white rectangular outlines.*

Figure 8. *On left schematic cartoon showing the development and fill of the Trans-Adirondack Back-arc Basin. On the right, relative probability histograms for zircon populations from each stage of basin evolution. See text for detailed explanation. Note that in contrast to the quartzites from Pyrites and the Lower and Upper Marbles, the Popple Hill Gneiss is dominated by anatectic or metamorphic zircon formed during the height of Shawinigan metamorphism (ca. 1170 Ma). Detrital zircons yield ages mostly between 1300-1400 Ma (Heumann et al., 2008).*



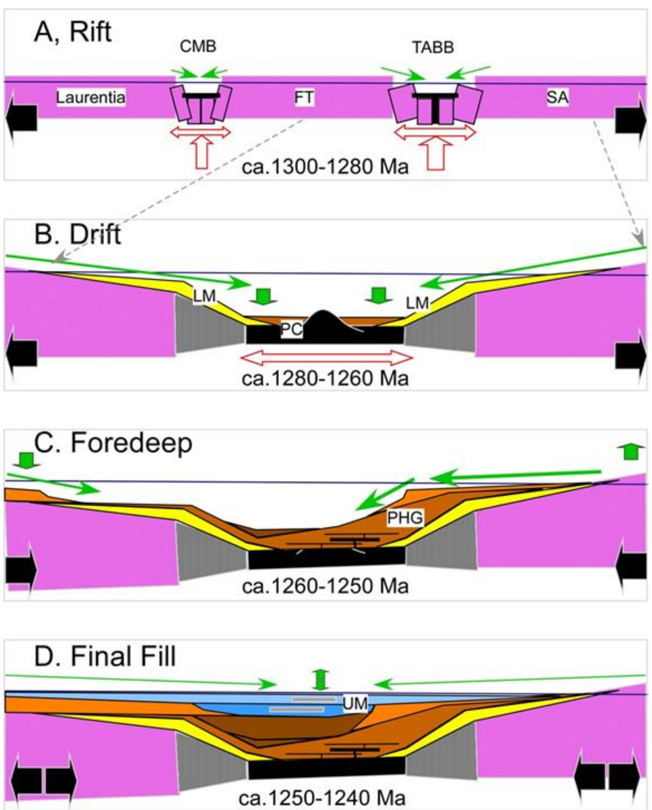


Figure 8 summarizes the results of the detrital zircon study. Our findings suggest:

1) The metasedimentary rocks in the Adirondack Lowlands were deposited in a relatively short interval of time between about 1260-1280 Ma.
2) The age of metamorphic overgrowths indicate both Elzeviran (ca. 1225-1235 Ma) and Shawinigan effects (ca. 1170-1185 Ma), with little to no secondary growth / recrystallization of zircon in the most quartz-rich units. Later Grenville events appear to have little to any impact on zircons in the Lowlands.
3) Several drastic changes in provenance occur throughout the sequence, with samples from the Lower and Upper marble containing detrital zircons of a wide age range and as old as ca. 3300 Ma; while those of the Popple Hill gneiss were derived predominantly from a terrane composed of rocks whose age is between ca. 1300-1400 Ma.
4) Detrital zircons from turbiditic metasedimentary rocks overlying the Pyrites Ultramafic Complex are largely restricted to a unimodal source ranging in age from ca. 1280-1300 Ma, likely derived from rift volcanism during birth of the Trans-Adirondack Back-arc Basin (Chiarenzelli et al., 2013).

IMPLICATIONS FOR THE TRANS-ADIRONDACK BACK-ARC BASIN

From their study of Nd T_{DM} model ages in the Central Metasedimentary Belt of Ontario and Quebec, Dickin and McNutt (2007) proposed a back-arc failed rift environment for supracrustal rocks of the Central Metasedimentary Belt (labeled Elzevirian rift zone in Figure 1). In concert with their work, Chiarenzelli et al. (2010b) used the term Trans-Adirondack Basin (TAB) to describe the depositional setting of metasedimentary rocks of the Adirondack Region and suggested that a series of back-arc basins developed along the southeastern margin of Laurentia prior to the Elzevirian Orogeny (ca. 1220-1245 Ma). Ultramafic rocks, occur in both the Central Metasedimentary Belt and Adirondack Region, indicating sufficient spreading occurred to generate oceanic crust (Chiarenzelli et al., 2010; Smith and Harris, 1996). The presence of both shallow and deeper water sedimentary protoliths in the Lowlands represent deposition in various depositional environments and during various stages of basin evolution (Figure 8).

The documentation of shifting provenance from detrital zircon study of rocks of the Lowlands metasedimentary sequence is thought to reflect drastic changes in the sediment supply in response to tectonic events (Figure 8). Initially, a deep basin formed in which turbiditic rocks, rich in organic material (graphite) and chemogenic components (pyrite) was derived from a restricted source yielding ages between ca. 1280-1300 Ma, likely a thin veneer of rift related volcanic rocks. Quartz-rich units in the carbonate-dominated Lower Marble, in this case, a biotite and tourmaline-bearing quartzite in the Lower Marble sequence, yielded a much wider range of ages compatible with derivation from a vast area within the interior of Laurentian, including zircons as old as 3300 Ma.

Rapid deepening of the Trans-Adirondack Basin, perhaps related the initial impingement of the Southern Adirondack Terrane during the Elzevirian Orogeny, resulted in a shift of provenance as the basin received detritus from the south, largely restricted in age to ca. 1300-1400 Ma. A foredeep, similar to the one developed during the Taconic Orogeny in eastern New York, is envisioned. Filling of the basin after deposition of the Popple Hill Gneiss led to shallow water deposition of the Upper Marble. Cessation or relaxation of tectonic forces and erosion led to rearrangement of drainages and shifting provenance. A broader, shallower basin, perhaps yoked to the Central Metasedimentary Belt. incorporated a source region once again incorporating eastern and central Laurentia. Figure 8 provides a schematic of the proposed sequence of events.

IMPLICATIONS FOR ZN-PB ORES

The metasedimentary sequence in the Adirondack Lowlands has long been of interest for numerous reasons, especially the occurrence of Zn-Pb ore within the stratigraphic sequence of Upper Marble. Dominated by sphalerite (Figure 9), these ores have been interpreted as sedimentary exhalative deposits related to the venting of basinal brines similar to younger unmetamorphosed and undeformed examples (deLorraine, 2001; deLorraine and Sangster, 1997). A key observation, made by former principal mine geologist William deLorraine, is that the ore horizons are restricted to three distinct stratigraphic intervals and each is underlain by evaporitic intervals, largely composed of anhydrite (deLorraine, 2001). Subsequent metamorphism and ductile deformation has resulted in the remobilization of the ore along fault planes cutting across the stratigraphy of the Sylvia Lake synform (Matt et al., 2019; 2020) thus explaining the occurrence of ore throughout much of the sequence and its occurrence within later cross-cutting faults.

The association of Zn-Pb sedex deposits with underlying evaporitic horizons is intriguing. Chiarenzelli et al. (2015) suggested that the stratigraphic positioning of the ores may be related compression from the approach of Southern Adirondack Terrane resulting in the beginning phases of closure of the Trans-Adirondack Back-Arc Basin. In such a scenario, communication with the open ocean may have been temporarily shut off due to compression allowing evaporative processes to dominate in a hot, dry climatic setting. Renewed subsidence upon relaxation of episodic compression allowed carbonate sedimentation to resume, punctuated by the ejection of basinal fluids along faults driven by far-field compression to the southeast. The uppermost unit of the entire sequence, the Median gneiss (Unit 16), is a fine-grained feldspathic unit, clastic in origin, with a multi-aged provenance, indicative of the Laurentian interior. It documents renewed clastic sedimentation, likely as the fringe of molassoid rocks encroaching on, cannibalizing, and covering the earlier carbonate-evaporite sequence.

Figure 9. *Coarsely crystalline sphalerite ore displaying interlocking twinned crystals, pyrite, and barite. Note the lack of brecciation and marble clasts typically found in ore associated with cross-cutting fault zones.*

BASEMENT ROCKS?

The Adirondack Lowlands has the hallmarks of a thin-skinned fold and thrust belt and also lacks recognizable basement involvement. Nonetheless, significant horizontal translation is implied due to the juxtaposing of shallow and deep water metasedimentary units and the incorporation of oceanic crust and upper mantle rocks. Evidence from Nd T_{DM} studies of various Shawinigan plutonic suites (Chiarenzelli et al., 2010b), oxygen isotopes in zircon (Peck et al., 2004; Figure 10), geographic changes in the nature of metasedimentary rocks, and structural studies, suggests the Black Lake shear zone (Wong et al., 2011), the boundary between the Frontenac and Lowlands terranes, represents a significant tectonic boundary, perhaps the reactivated northwest rifted margin of the TABB. Thus it is reasonable to expect that the basement to the Lowlands stratigraphic sequence is related to basement rocks of the Frontenac Terrane, itself rifted southwestward (present coordinates) during formation of the Central Metasedimentary Basin back arc-failed rift (Dickin and McNutt, 2007; Moretton and Dickin, 2013). On the northwest margin of the CMB boundary zone ca. 1300-1350 Ma arc plutonic rocks of the Dysart-Mt. Holly suite occur (Agustsson et al., 2013), while the Frontenanc Terrane consists of a variety of plutonic gneisses with Nd T_{DM} ages ranging from 1350-1640 Ma (Dickin and Strong, 2021) indicative of a segment of older crust reworked along the margin of Laurentia.

To better understand the possible basement rocks to the Lowlands stratigraphic sequence two samples of the Hyde School gneiss were sampled and analyzed to examine zircon xenocrysts. Dominated by xenocrystic zircon, the Hyde School gneiss, originally thought to be the basal stratigraphic unit in the Lowlands, was selected due to its ferroan leucogranitic composition compatible (Peck et al., 2013) with the preservation of considerable quantities of inherited zircon. Its age was reported as 1172 Ma (Wasteneys et al. 1999). Previously unreported, a sample of the Canton Body yielded xenocrysts of a single age (1322.5 ±2.1 Ma) and a sample from the Gouverneur Body yielded a range of xenocrysts and zircon cores from (1227-1363 Ma). Finally, xenocrystic cores from a diopside-kspar pegmatite cross-cutting breccia in the Raquette River associated with the Carthage-Colton shear zone in Colton gave an age of 1328.5 ±8.3 Ma (Chiarenzelli et al., 2019). Thus it appears likely that the basement rocks upon which the Lowlands metasedimentary sequence were deposited on are similar to those found in Dysart-Mt. Holly Suite, including those exposed in the Southern Adirondack Terrane, Green Mountains of Vermont, and the NY-NJ Highlands (Agustsson et al., 2013; McLelland and Chiarenzelli, 1990; McLelland et al., 2010).

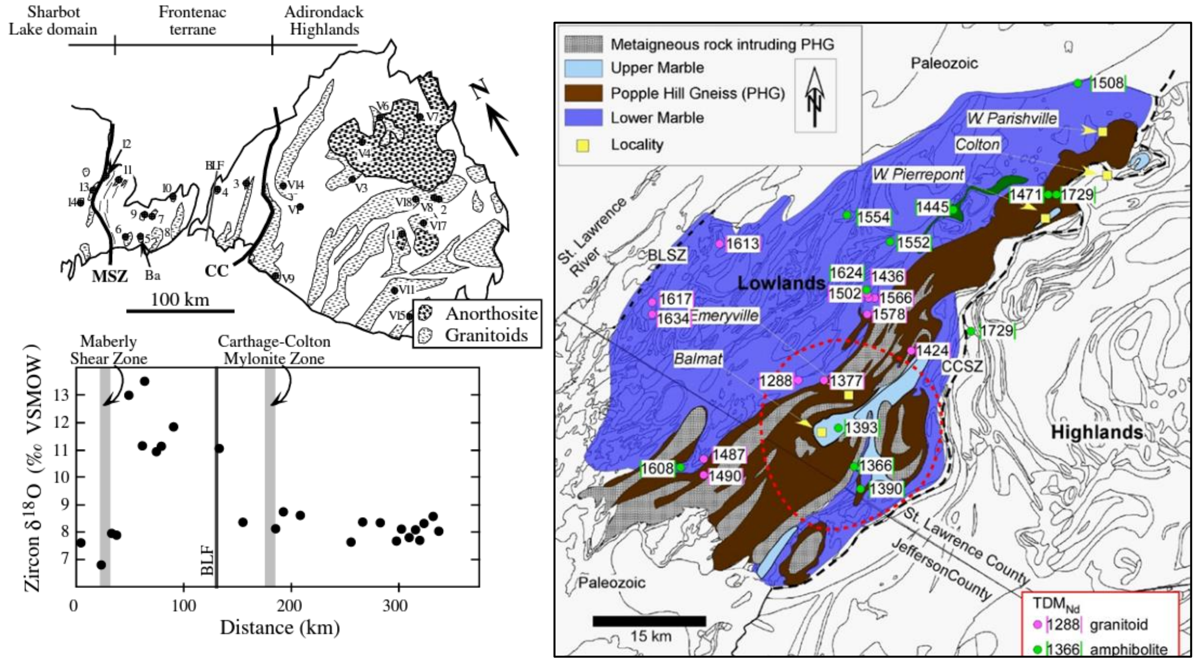


Figure 10. *Left. Oxygen isotope in zircon cross-section from the Frontenac Terrane (NW) extending across the Adirondack Highlands (SE). Note the jump in values at the Black Lake Fault (BLF). Right. Neodymium time of depleted mantle ages (T_{DM}) in the Adirondack Lowlands. Note the difference between felsic (pink) and mafic (green) igneous precursors and the older ages obtained from felsic plutons as the Black Lake Shear Zone (BLSZ) and Frontenac Terrane are approached.*

Figure 11. *Location of stops to be made during this field trip arranged in sequence. Note that the trip starts at the USPS in the village of Richville (Main Street, near Stop 1) and ends at STOP 6 west of the village of*

Canton, New York on Highway 11. Stop 5 is at Brown Hall on the campus of St. Lawrence University in Canton, NY, where we will stop to see core from the Sylvia Lake Synform and the quartzite from the Popple Hill Gneiss/Upper Marble transition.

FIELD GUIDE

Meeting Point: USPS parking lot, Main Street, Richville, New York

Meeting Point Coordinates: lat. N 44.417641° long. W -75. 391066°

Meeting Time: 9:00 AM, Sunday, September 25, 2022

Figure 12. *Starting point of field trip, USPS, Main Street, Richville, NY.*

<u>**STOP #1** – Lower Marble, Richville, Tourmaline-bearing gneisses (Figure 13)</u>

The Lower Marble includes a number of detrital metasedimentary rocks in addition to the marble and subordinate calc-silicate gneiss (Wiener et al., 1984). Of particular interest is an extensive (50 km) belt of black to reddish-brown tourmaline-rich rocks that are interlayered with dolomitic marble near the top of the sequence (Brown and Ayuso, 1985; deLorraine and Carl, 1986). Along Rt. 11, just outside of Richville, a meter-thick layer of feldspathic quartzite (arkose) within biotite, tourmaline and pyrite-rich gneiss was sampled for U-Pb zircon geochronology.

A meter-wide, shallowly inclined quartz-rich layer, whitish in color, on the north side of Rt. 11 was sampled (Figure 14). A small population (n = 300-400) of zircon grains and a large number of pyrite and tourmaline grains were separated from approximately 1 kg of sample. The zircons ranged in size from 50 to 300 µm and were predominantly rounded to oval in shape, although some are euhedral (Figure 15). Truncated oscillatory zoning and a few, thin 1-5 µm, partial, euhedral metamorphic overgrowths were observed. Uranium concentrations in zircon range from 12 to 742 ppm and average 163 ± 135 ppm. Ratios of uranium to thorium range from 0.4 to 12.4 and average 2.0 ± 1.9. The vast majority of analyses are concordant.

***Figure 13**. Sampling location of tourmaline-rich quartzose unit along Rt. 11, near Richville, New York.*

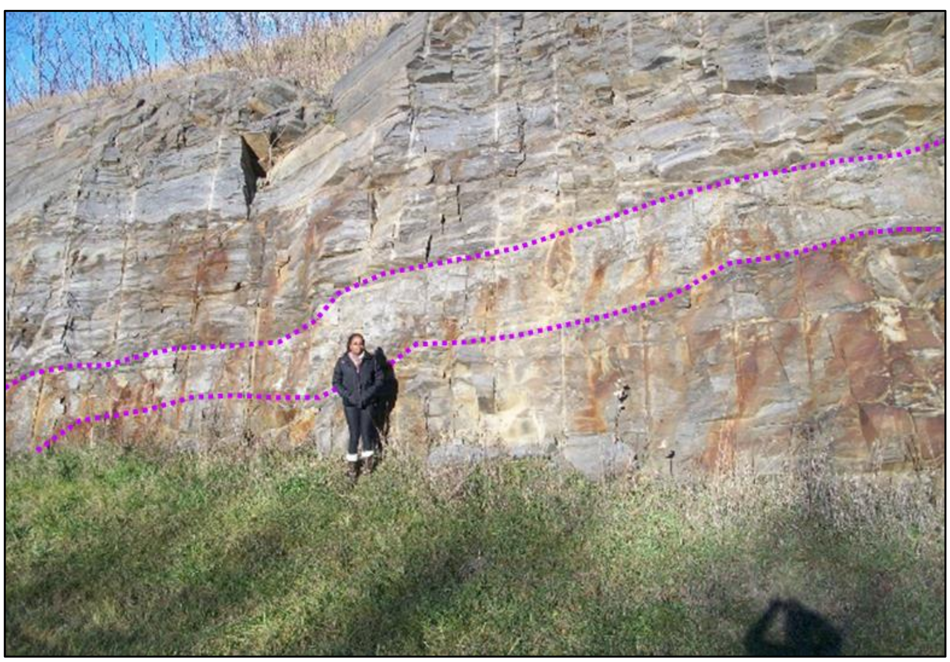


***Figure 14.** Outcrop of the tourmaline-rich unit exposed near Richville, New York. The sample was collected just to the left of Ms. Roselyne Laboso, within the whitish layer of glassy quartzite, speckled with black tourmaline grains (outlined in purple). Note abundant pyrite staining.*

One hundred and one grains are near concordant and show a wide range of ages from 1241.6 Ma to 3082.9 Ma. The youngest grain analyzed is 1241.6 ± 41.9 Ma. A cohort of the 20 youngest grains, all within analytical error of one another, gave a weighted mean of 1263.9 ± 4.3 Ma (MSWD = 1.4; PROB= 0.11), and are interpreted as the age of the youngest detrital population. Two large peaks, one at 1260.2 ± 4.7 Ma and the other at 1841 ± 2.1 Ma, are clearly defined on the probability density histogram (Figure 16) and represent the dominant ages found in this sample.

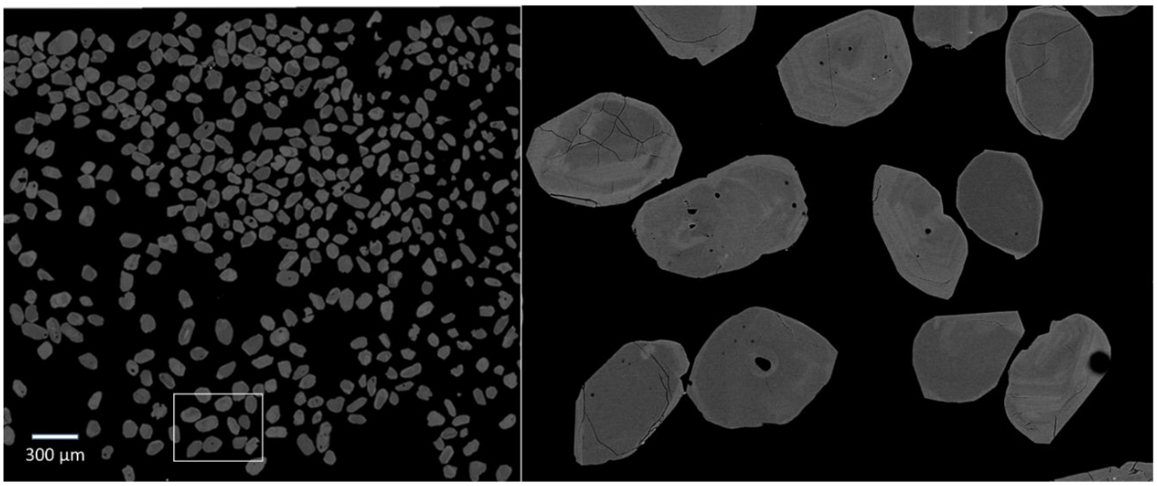


Figure 15. *Left. View of RT-11 zircon mount in backscattered electron mode of the scanning electron microscope (BSE-SEM) Right. Detail of inset outlined in white on the left side. Note the various shapes and internal features of the zircons, including zoning, truncated zoning, inclusions, and thin, discontinuous, bright rims.*

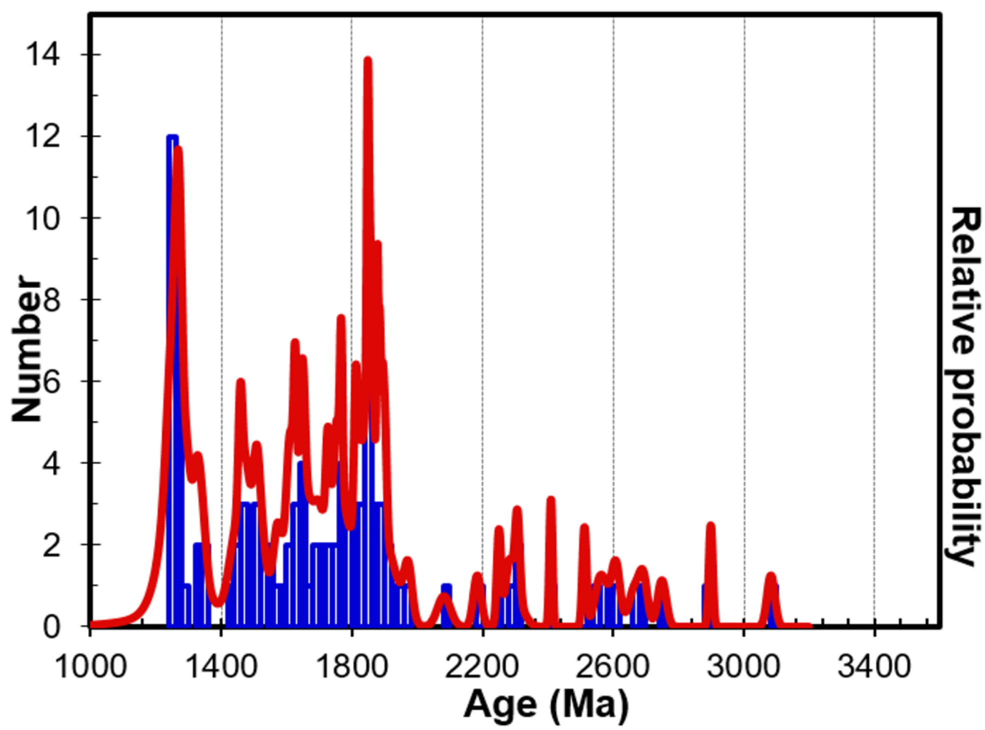


Figure 16. *Probability histogram of detrital zircon U-Pb ages from the Richville quartzite sample of the Lower Marble.*

Stop #2 – Upper Marble, Unit 4, Balmat, Stromatolites (Figure 17)

Figure 17. *Location of the Upper Marble Unit 4 (stromatolite-bearing) sampling site. The Empire State Mine is currently owned by the Titan Mining Corporation (formerly Zinc Corporation of America).*

This classic outcrop contains putative stromatolites originally noted by Yngvar Isachsen (Isachsen and Landing, 1983). Located directly across from the Titan Corporation Mine, here the contact between Units 4 and 5 of the Upper Marble can be seen. Structurally we are located on the overturned limb of Sylvia Lake Synform, which is consistent with the downward concave shape of the stromatolites.

The outcrop directly across from the entrance to the former Zinc Corporation of America headquarters near Sylvia Lake was sampled for detrital zircon geochronology (Figure 18). The rock sampled is from Unit 4 and comprised mostly of dolomite and diopside. It exhibits silicified layering interpreted as remnant stromatolites (Isachsen and Landing, 1983). Here the matrix between the sparse, upside-down stromatolite domes (located on the overturned limb of the Sylvia Lake Synform) was sampled and consisted of quartz, dolomite, serpentine, titanite, and gray diopside (no stromatolites were harmed or disfigured during the procurement of this sample). A sparse yield of approximately 200 silt-sized (20-100 µm) zircons of rounded to angular shape was obtained (Figure 19).

The U-content of zircons ranges from 95 to 2331 ppm and averages 947 ± 431 ppm, consid-erably higher than all other samples. Ratios of U /Th range from 1 to 113 and averages 12 ± 18. Because of the small size of the zircon and high U-content, little in the way of internal features can be discerned as the BSE signal is very homogeneous (Figure 19). Every grain in the mount larger than the 30 µm analytical spot size was analyzed, yielding 92 data points. Nearly all grains are within± 3% of Concordia.

The youngest zircon grain analyzed gave an age of 994.7 ± 29.1 Ma, which falls within the time frame noted for the Rigolet pulse of the Grenvillian Orogeny. A cohort of 66 grains yielded an age of 1172.7 ± 3.0 Ma (MSWD = 1.09; PROB = 0.30). Thirteen grains yield ages ranging from 1214.5 to 2607.5 Ma (Figure 20). Two grains gave a weighted mean of 1224 ± 13 Ma (MSWD = 0.34; PROB = 0.80), interpreted as the timing of Elzevirian metamorphism or analysis ablation pits that sample across age domain boundaries.

Figure 18. *Field photograph of the Balmat stromatolites, Unit 4, opposite to the entrance to the zinc mine. Lens cap for scale.*

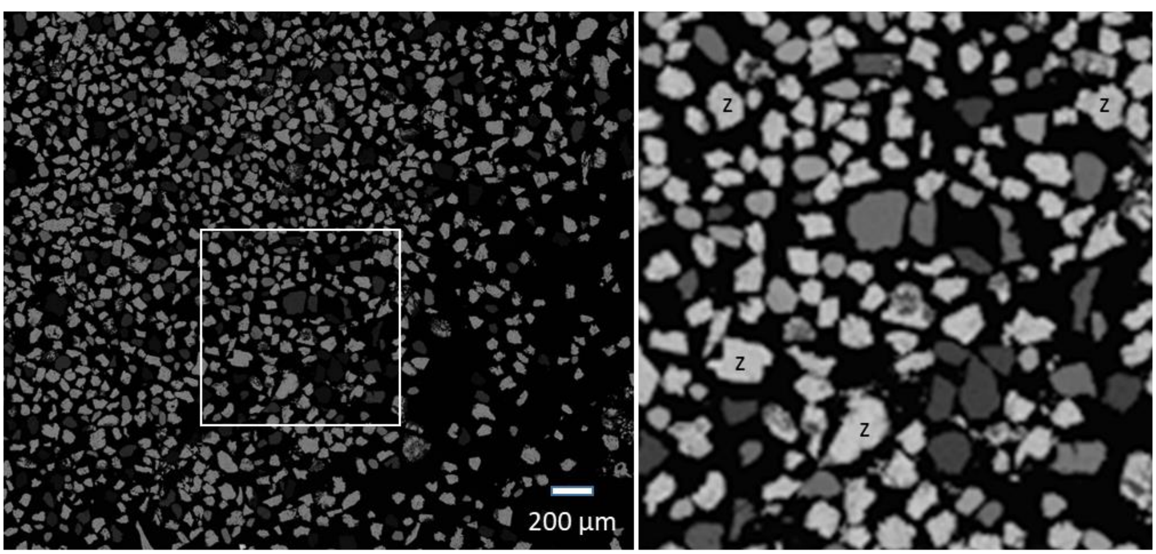


Figure 19. *Left. Population of zircons from sample of Unit 4, stromatolitic-bearing Upper Marble. Right. Blow up of inset on left showing the relatively bright SEM-BSE response of zircon (z) compared to accessory minerals including titanite.*

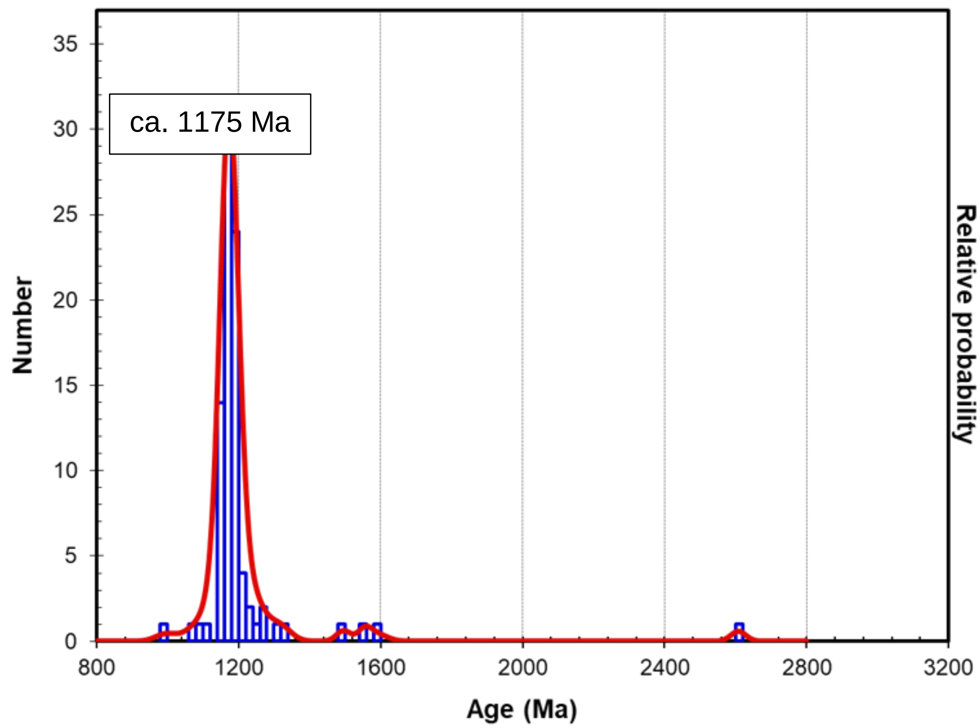


Figure 20. *Relative probability histogram for the Balmat stromatolite matrix sample. Note the vast majority of the analyses give ages of ca. 1175 Ma and are interpreted as metamorphic in origin.*

STOP #3 – Upper Marble, Unit 16, Talcville, Median Gneiss (uppermost unit; Figure 21)

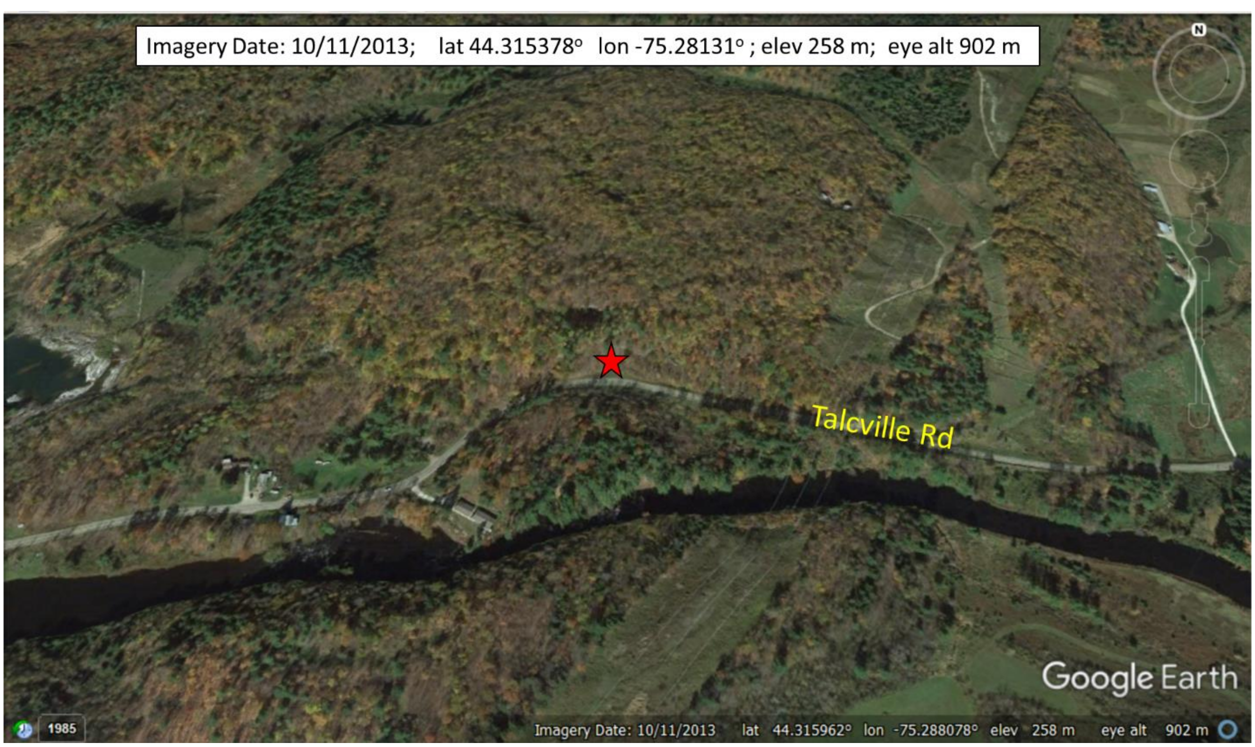


Figure 21. *Location of Median Gneiss sample near Talcville, west of Edwards, New York.*

The Median Gneiss (Unit 16 of the Upper Marble) is the youngest member of the stratigraphic succession of the Grenville Supergroup in the Lowlands. It is primarily a pink, strongly layered, quartzofeldspathic rock with a small percentage of other minerals such as diopside, micas, tourmaline, hornblende, and scapolite. Foliation-parallel layers of leucosome and feldspar porphyroclasts attest to the unit's metamorphic grade and deformation (Figure 22). The protolith of the Median Gneiss is not known, although its composition is granitic or arkosic. A 1 kg-sized sample was collected from a small outcrop exposed near Talcville, New York and it yielded thousands of zircon grains.

The zircon grains range in size from 30 to 100 µm in diameter; most grains are about 50 µm and rounded, but small populations of angular and elongate grains were also noted (Figure 23). The average U-content is 842 ± 548 ppm and ranges from 52 to 2892 ppm. Ratios of U /Th range from 0.63 to 27.78 and average 3.28 ± 3.23.

Figure 22. *Photograph of the sawn surface of the Median Gneiss sample from Talcville. Note the compositional layering, deformed leucosomal layers, and pink, feldspar porphyroclasts. Black crystals are tourmaline.*

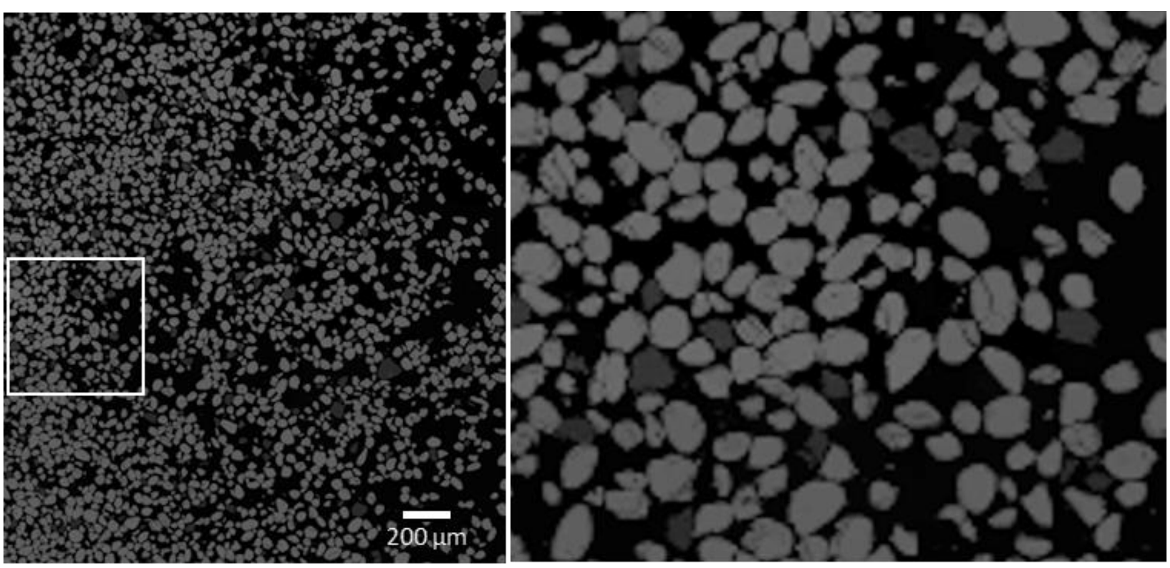


Figure 23. *Left. Zircon population recovered from the Median Gneiss, the uppermost unit of the Upper Marble. Right. Blow up of inset on left showing the SEM-BSE response of zircon (here bright) and tourmaline (dark).*

Zircon U-Pb analyses yielded ages from 1185.4 to 3303.3 Ma. Over 300 grains were analyzed; however, those falling outside the range of 95-105% concordant were filtered out of the data set leaving 158 analyses to be plotted (Figure 24). Of these, the largest peak on the probability density histogram is 1524.8 Ma. Two grains yield an age of 1234 ± 12 (MSWD = 0.29; PROB = 0.59), falling within the range of timing of Elzevirian orogenesis. The next five oldest grains form a coherent group with an age of 1253 ± 17 (MSWD = 3.7; PROB= 0.005).

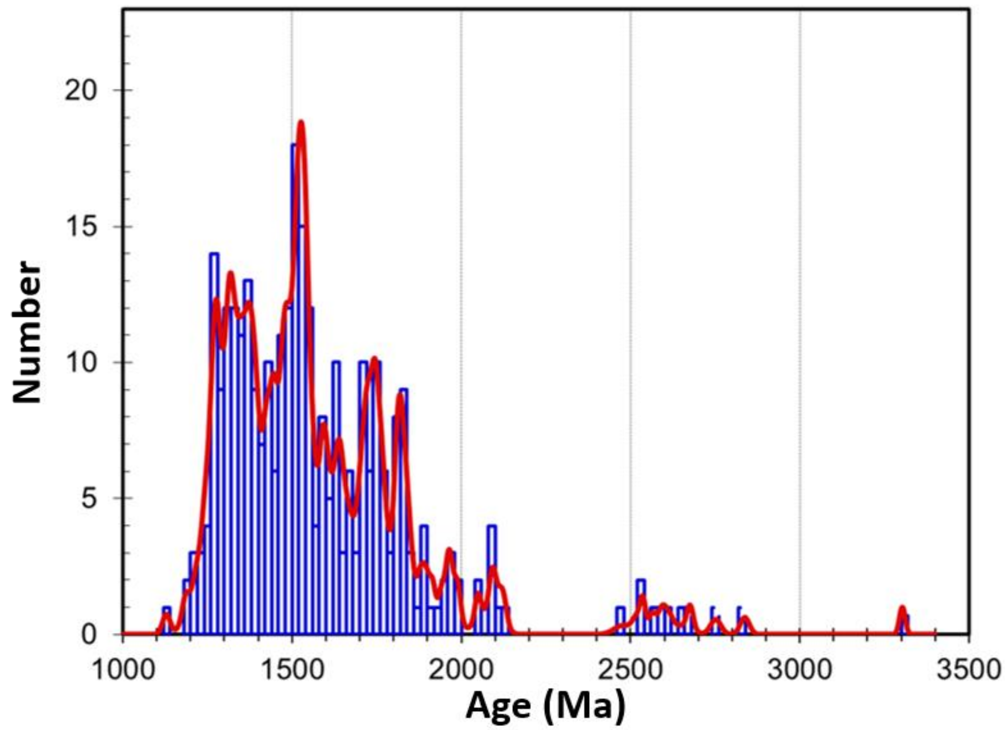


***Figure 24**. Relative probability histogram from U-Pb analyses of zircons from the Median gneiss.*

<u>STOP #4</u> <u>–Pyrites Complex, Pyrites, Turbiditic Basal Unit, (Figure 25)</u>

***Figure 25.** Location of the Pyrites turbidite sampling site. Note gossaneous weathering cap at sampling site along the Grasse River.*

Along the Grasse River near the adits to the old pyrite mine at Pyrites, New York, small samples of quartz-rich (~ 85% SiO_2), cm-scale interlayers within a gamet-sillimanite pelitic gneiss were removed utilizing a chisel and processed for zircon. The rock shows isoclinal folding of interbedded quartzite and metapelitic layers (Figure 26). Centimeter-scale layers are interpreted as the alternation of sand, silt, and mud formed within a turbidite sequence (Chiarenzelli et al., 2015). Three meters structurally below the sample site, a coarse-grained, green, hydrothermally altered peridotite is exposed. These rocks, along with more extensive gabbroic and amphibolitic units, have been named the Pyrites Complex and interpreted as a highly disrupted ophiolite suite (Chiarenzelli et al., 2011a). Continuous exposure, gradation in the composition of the metasedimentary rocks, and the occurrence of chromite (Tiedt and Kelson, 2008) in the pelitic gneiss near the contact suggests the metasedimentary sequence overlies the ultramafic in apparent depositional contact.

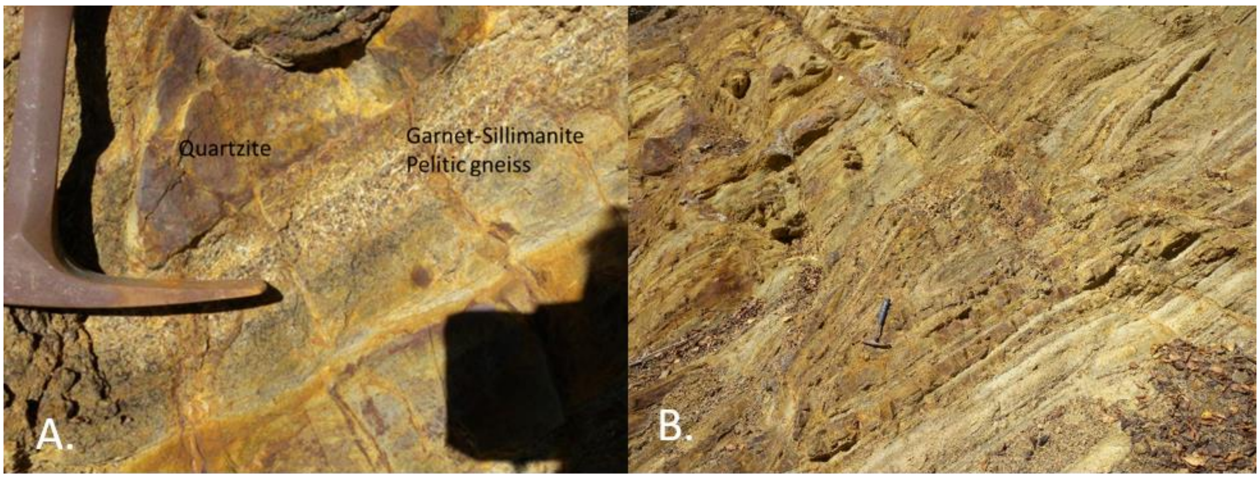


Figure 26. *(A) Hinge area of isoclinal fold at Pyrites showing putative sand-silt-mud turbidites, now quartzite and pelitic gneiss. Note iron staining from abundant pyrite. Geological pick for scale. (B) Expanded view of same photograph showing extent of folding. Note the cross-cutting set of mineralized fractures, mostly quartz and pyrite, dipping shallowly upstream (to the right).*

A small separate (n = several hundred grains) of zircon was obtained from a kilogram of sample quartzose, sand to silt-sized portions of the aforementioned outcrop. Zircons recovered are relatively small (<100 µm), oscillatory zoned, and have shapes ranging from stubby dipyramids to grains with slightly rounded boundaries (Figure 27), thought modified by erosion. The zircons from this sample are noteworthy for their homogeneity. The U-content of the zircons analyzed in this sample averaged 243 ± 93 ppm and their U /Th ratio is 1.6 ± 0.4. Most grains are concordant with a range of 104.0-94.7%.

Ninety-seven near-concordant grains are plotted (Figure 28) and a group of 86 yielded a weighted average of 1289.7 ± 1.1 Ma. One grain, of smaller size and lacking visible zoning, gave an age of 1176.2 ± 24.1 Ma, in excellent agreement with the timing of Shawinigan orogenesis, and is considered to be metamorphic. Another grain yields an age of 1237.5 ± 24.1 Ma, the timing of Elzevirian orogenesis in the Grenville Province and is also interpreted to be of metamorphic origin. A group of five analyses that are statistically indistinguishable, gave a weighted mean age of 1258.3 ± 7.7 Ma (Mean Square Weighted Distribution (MSWD) = 1.4; Probability (PROB) = 0.22) and are interpreted to be the age of the youngest detrital population. Six other detrital grains range in age between 1372.9 ± 21.2 Ma and 2294.9 ± 21.7 Ma.

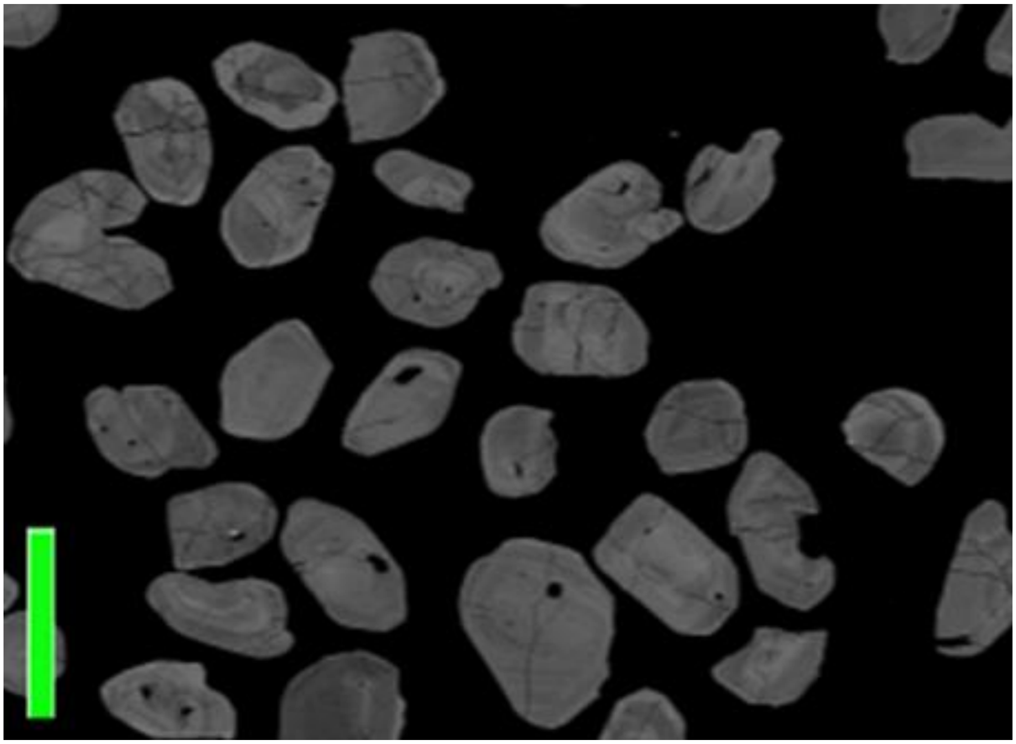


Figure 27. *SEM – BSE image of zircon population from the Pyrite quartzite layers. Scale bar is 100 microns. Zircons from this sampling site were remarkably homogeneous in size, shape, and appearance and unimodal in their age, suggesting a restricted source.*

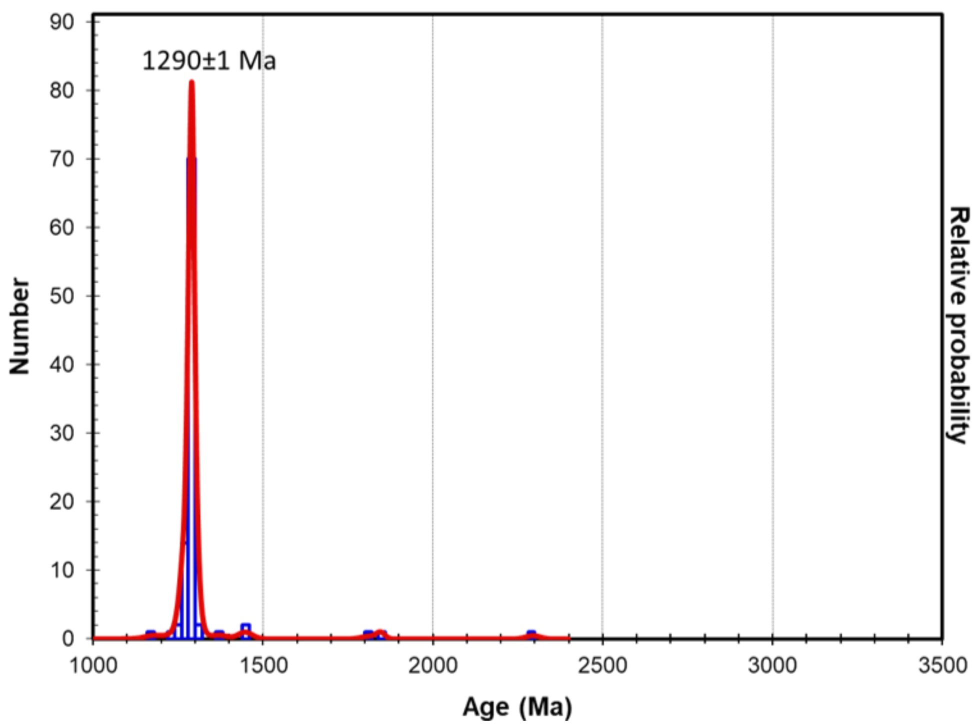


Figure 28. *Relative probability histogram from U-Pb analyses of zircons from the Pyrites quartzite.*

<u>**Stop #5** – Contact of Upper Marble and Popple Hill Gneiss, Sylvia Lake Synform, Glassy quartzite intervals (Figure 29)</u>

Good time for a break. Core will be set out in the Department of Geology, 141-142 Brown Hall, St. Lawrence University

Figure 29. *Location of Brown Hall on the St. Lawrence University campus. Here we will have a look at the glassy quartzite recovered from a drill hole penetrating the Popple Hill gneiss/Upper Marble contact within the Sylvia Lake synform. Park in the lot directly south of the red star and proceed to Rm. 141-142 Brown Hall.*

A section of drill core from an overturned limb of the Sylvia Lake Syncline near Balmat, New York (Figure 30) penetrated the upper portion of the Popple Hill Gneiss and the lower portions of the Upper Marble (Chiarenzelli et al., 2012). The transition is considered to be conformable as the percentage of quartz in Popple Hill Gneiss gradually increases upward into ~ 30 m of glassy quartzite (Figure 31), then transitions into schist, before the first marble interval is penetrated. Portions of split drill core from several meters of section composed of the glassy quartzite were sampled, crushed, and zircon grains separated.

An excellent yield of several thousand grains was obtained from 1 kg of sample. Grains range from highly rounded to angular (Figure 32). Their average size is about 100 µm, with large rounded grains up to 300 µm. Smaller angular and euhedral grains are also present. Many zircon grains show truncation of oscillatory zoning, but distinct overgrowths are few, thin, and incomplete. The average uranium content of zircons from the sample was 106 ± 75 ppm and U /Th range from 0.4 to 13.6 and average 1.5 ± 1.5.

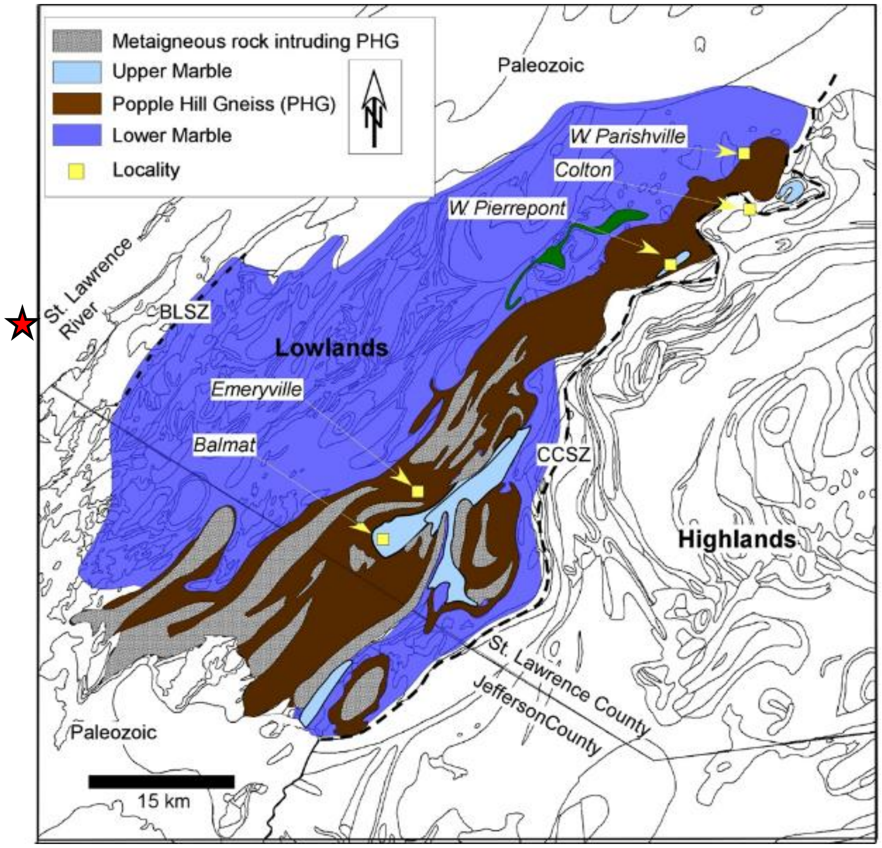


Figure 30. *Sample location for the drill core sample (red star), just north of the square labeled Balmat, with the Sylvia Lake Synform.*

Figure 31. *Glassy quartzite interval recovered from the contact between the Upper Marble just above the Popple Hill Gneiss.*

One hundred and seven near-concordant grains were plotted and showed a wide range of ages from 1270.8 Ma to 3388.3 Ma. The youngest grain analyzed gave an age of 1270.8 ± 113 Ma (note the large error). A cohort of 5 youngest grains, all within analytical error of one another, gave a weighted mean of 1277.9 ± 13 Ma (MSWD = 0.09; PROB = 0.994) and are interpreted as the age of the youngest detrital population. Two large populations, one at 1446 ± 7.8 Ma and 1650.3 ± 6.2 Ma, occur on the probability density histogram (Figure 33).

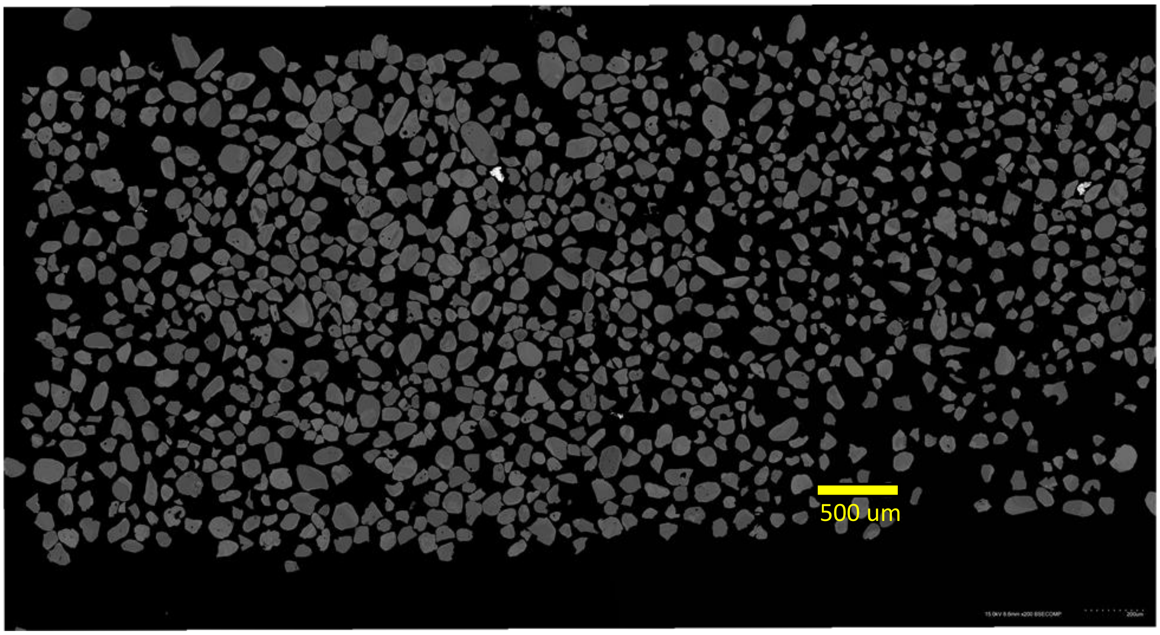


Figure 32. *Zircons recovered from the glassy quartzite drill core interval, Sylvia Lake Synform. Note rounded shapes.*

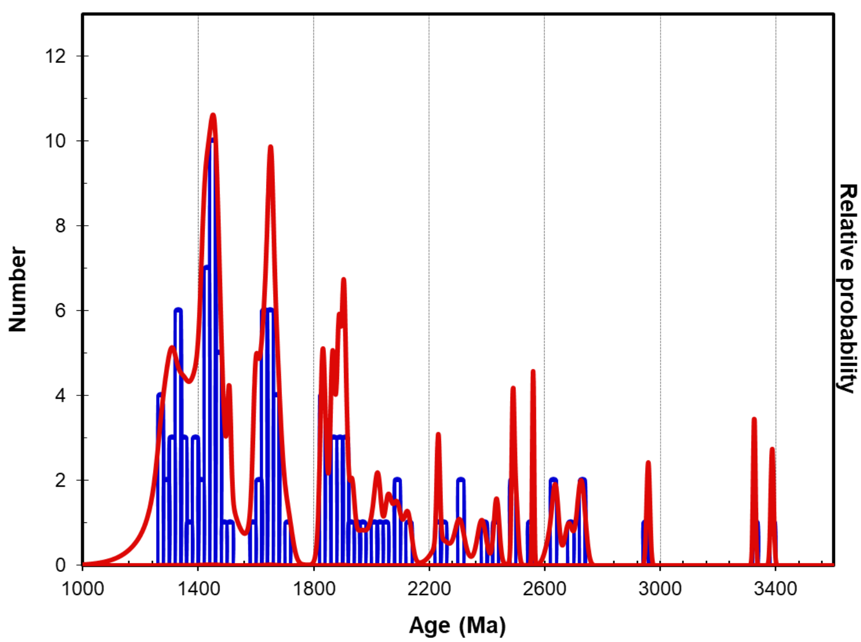


Figure 33. *Relative probability histogram from U-Pb analyses of zircon separated from the glassy quartzite core interval between the top of the Popple Hill Gneiss and base of the Upper Marble.*

<u>Stop #7</u> <u>– Canton Alaskite Body, Canton, Leucogranite and amphibolite (Figure 34)</u>

Figure 34. *Location of the sample collected from the Canton leucogranite body just southwest of the Canton railroad overpass of State Highway 11. The Grasse River is visible to the right.*

About a dozen, kilometer-scale, domical leucogranite bodies, known as alaskites occur throughout the Adirondack Lowlands. Originally interpreted as volcanic rocks at the base of the stratigraphic sequence (Carl and van Diver, 1975; Carl et al., 1990; deLorraine and Carl, 1986), subsequent work (McLelland et al. 1992; Peck et al., 2013; Wasteneys et al., 1999) has reinterpreted the suite as of intrusive origin. Known as the Hyde School gneiss these bodies range in composition, contain supracrustal xenolithic screens, and are often intruded by dismembered mafic dikes/sills or layers (now amphibolite). A large portion of these bodies consist of medium-grained interlocking crystals of alkali feldspar, plagioclase, and quartz. Nearly all mafic minerals are magnetite. Foliation is often difficult to see, however, the rock is gneissic in nature particularly near the margins of the bodies. A kilogram sized sample was collected from the pink leucogranite layer above the mafic enclaves in Figure 35.

Zircons from the leucogranite bodies investigated thus far are dominated by large crystals yielding core ages of ca. 1325 Ma. Some show dissolution, presumably from interaction with magma, during the formation of their igneous precursors. Many zircons have thin, clear, rims which yields an U-Pb age of ca. 1170 Ma. A small population of euhedral, smaller crystals lacking cores, are often observed. Earlier attempts (1985) to date the zircons using multi-grain samples yielded hybrid ages (ca. 1230 Ma; McLelland et al., 1992). Subsequent attempts at dating using refined techniques (single grain analyses of coreless crystals) yielded an age of 1172 Ma (Wasteneys et al., 1999). This data indicates substantial inheritance from the source terrane(s).

Figure 35. *Photograph of sampling site of the Canton leucogranite body west of Canton along Rt. 11. Three main lithologies occur, leucogranite, boudinaged amphibolite layers, and pegmatite. Here the leucogranite layer was sampled. Hammer for scale.*

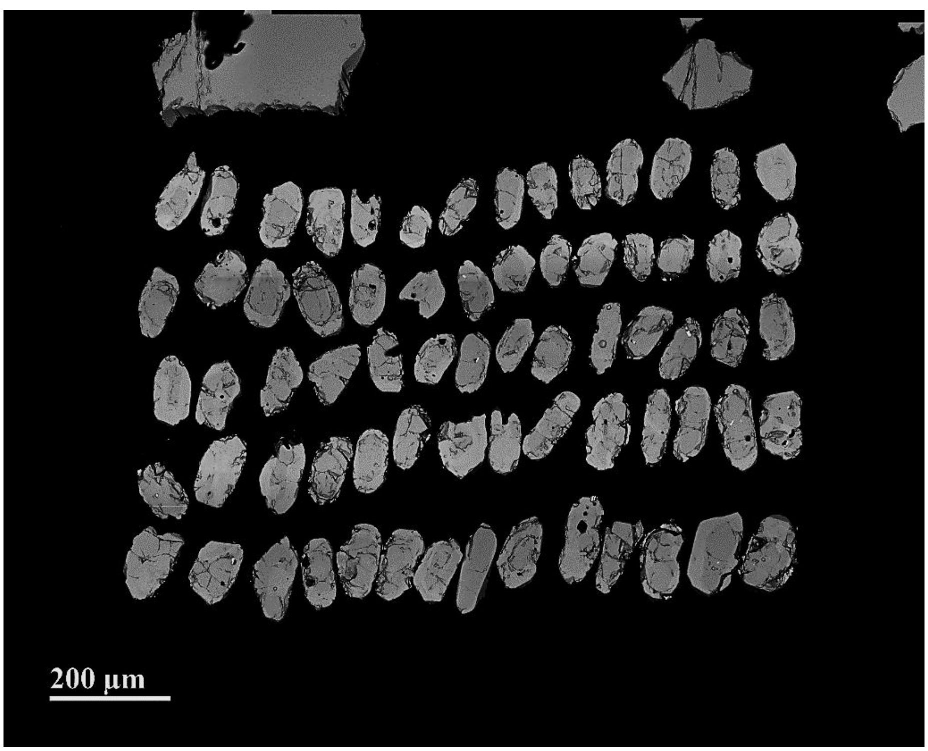


Figure 36. *Back scattered electron SEM image of zircons separated from the Canton leucogranite body. Note the shapes, internal features, and core-rim relationships.*

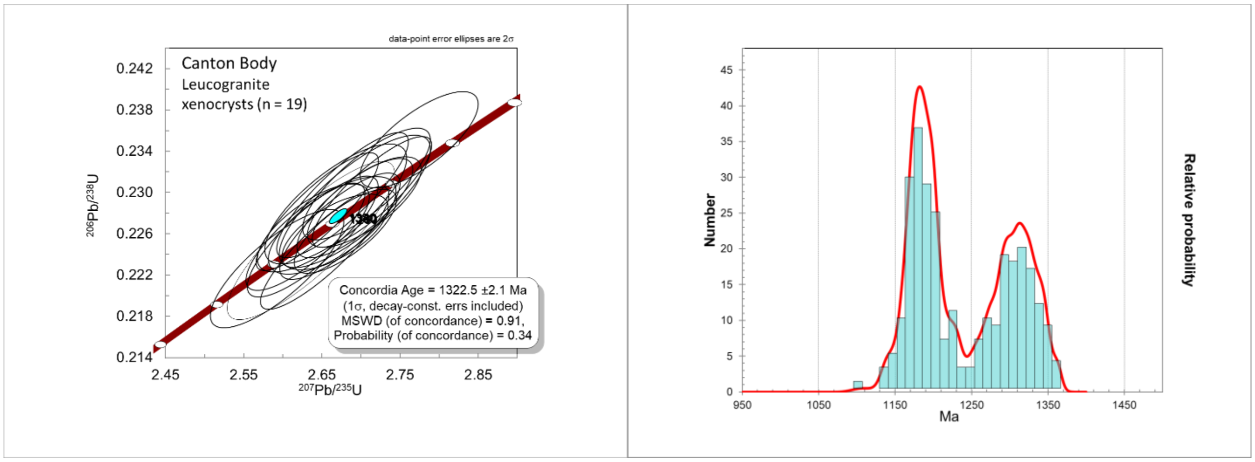


Figure 37. *Left. Concordia diagram of cores analyzed from zircon separated from the Canton leucogranite body. Right. Relative probability frequency diagram from the Hyde School Gneiss, Gouverneur Body, showing a bimodal distribution of 1175 and 1315 Ma (Unpublished data from Erkan Toraman).*

Bonus Stop – Popple Hill Gneiss, Pierrepont, Calc-silicate interlayer (Figure 38)

Figure 38. *Location of the sampling site of a calc-silicate interlayer in the Popple Hill gneiss.*

A stop on private property at O'Brien Rd. in Canton, New York will be made if time and interest permits.

A rusty, granular, calc-silicate gneiss was sampled from a 10 cm thick interval in garnet-bearing pelitic portion of the Popple Hill Gneiss near O'Brien Road, near West Pierrepont, New York (Figure 39). The gneiss is interlayered and cut by numerous concordant to discordant leucogranitic gneissic sheets. The Popple Hill Gneiss consists of a thick (several kilometers?) sequence of metamorphosed mud, silt, and sand, that is, at least in part, turbiditic (Chiarenzelli et al., 2012). Samples from partially melted pelitic to psammitic portions of similar gneisses have been investigated for U-Pb zircon geochronology by Heumann et al. (2006) and

Bickford et al. (2008) in the Adirondack Lowlands and Highlands and have been shown to have zircons of both detrital and igneous-anatectic origin.

A small population (~500 grains) of zircons was separated from a kilogram-sized sample of rusty calc-silicate gneiss (Figure 40). Zircons range in size from 20 to 80 µm and display a wide range of shapes from nearly rounded and oval to subhedral and faceted. A number of angular grains also occur.

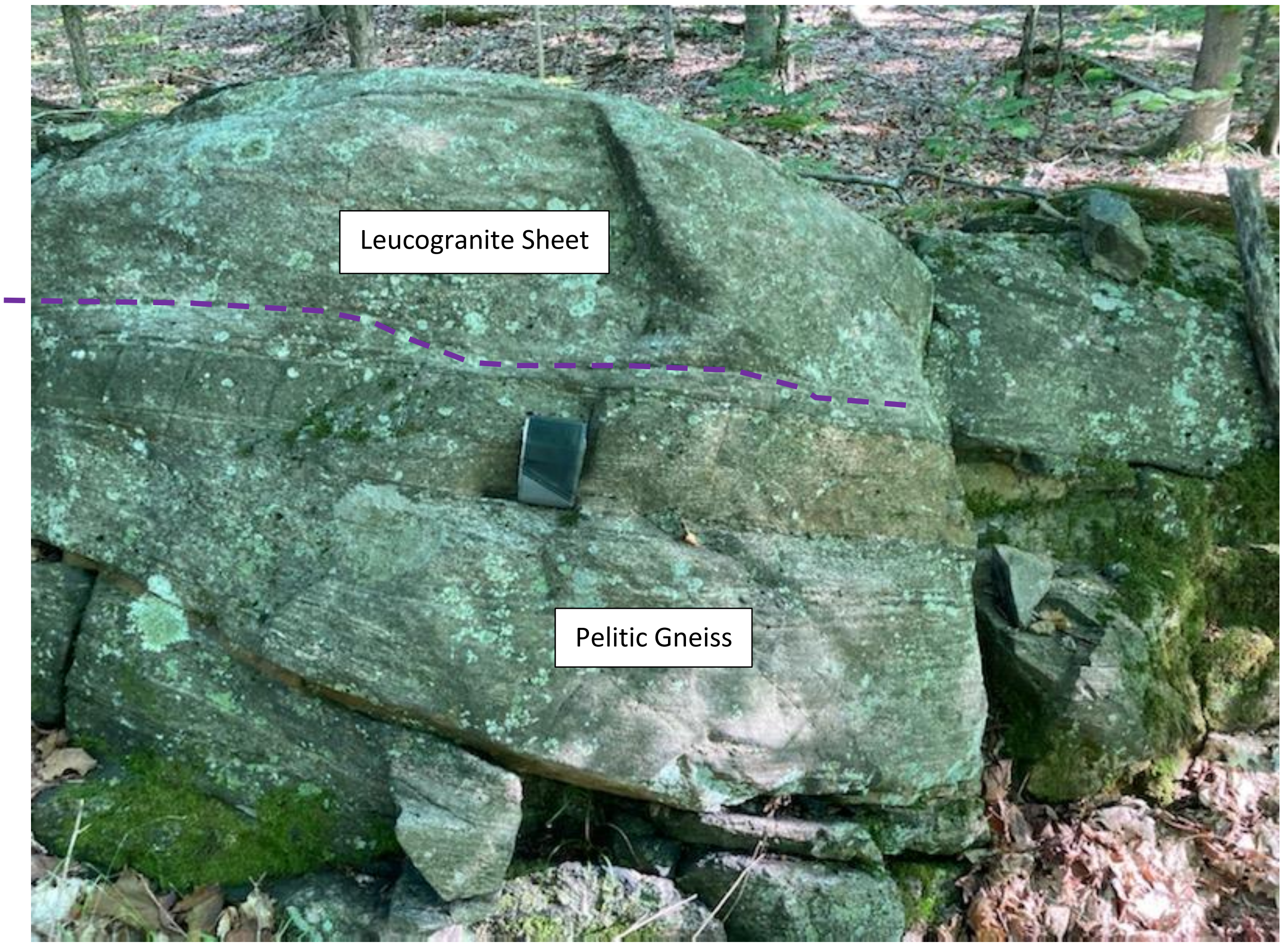


Figure 39. *Ten cm-thick interlayer of calc-silicate in the Popple Hill Gneiss, sampled for zircon U-Pb analysis. Wallet for scale. Dashed purple line shows contact between leucogranite and pelitic gneiss. Note calc-silicate unit pinches out towards the right side of the photograph and forms boudins up to a meter or more in width in other nearby localities.*

Seventy-nine grains are near concordant and yielded a range of ages between 1127.0 Ma and 1667.4 Ma (Figure 41). A population of 61 grains yielded an age of 1170.6 ± 4.0 Ma (MSWD = 1.5; PROB = 0.010), which falls in between the range of 1180-1160 Ma interpreted by Heumann et al. (2006) as the time of anataxis in nearby samples of this unit. Two grains, interpreted as the youngest detrital population, give an average age of 1260 ± 23 Ma (MWSD = 0.024; PROB = 0.88). Four grains gave an age of 1223 ± 8 Ma and are interpreted as metamorphic (Elzevirian) or hybrid in origin (MSWD = 0.57; PROB = 0.64).

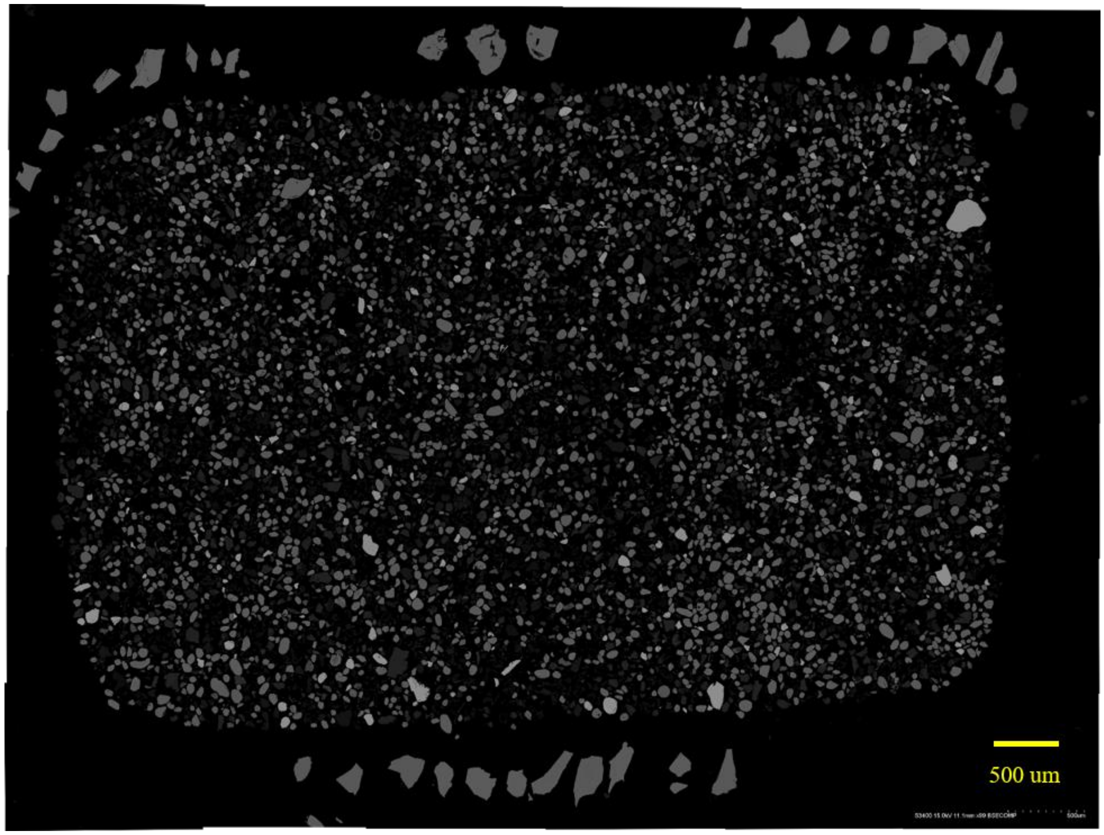


Figure 40. *View of the entire mount prepared for the O'Brien road calc-silicate sample, showing zircons recovered and mounted standards. Zircons from the sample are rounded and mid-grey in BSE, similar to the larger zircon standards surrounding the sample population. Darker and brighter grains are other accessory minerals.*

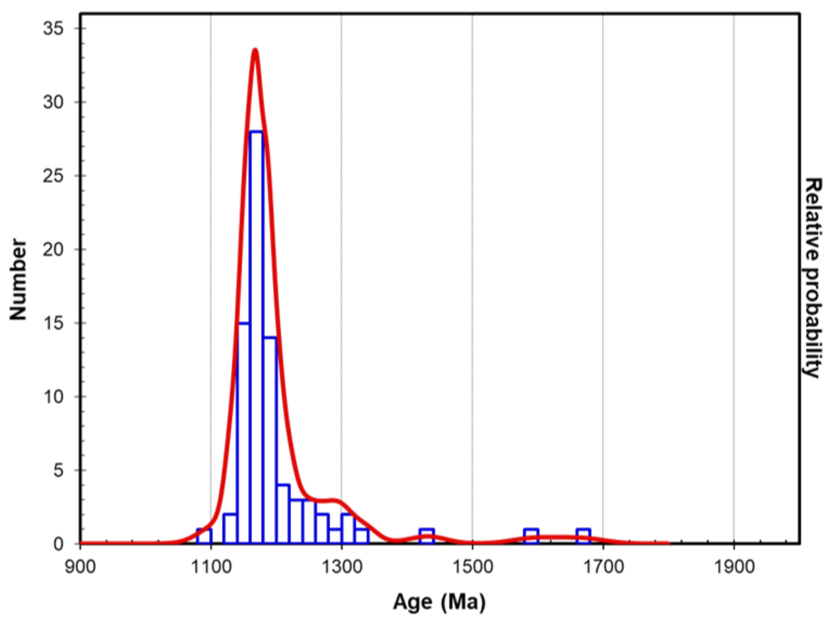


Figure 41. *Relative probability histogram from U-Pb analyses of zircon separated from calc-silicate layers in the Popple Hill Gneiss. Note the major peak occurs about 1170 Ma; peak timing of Shawinigan metamorphism in the Adirondack Lowlands.*

REFERENCES CITED

Agustsson, K., Johnston, S., Peck, W., and Kylander-Clark, Andrew, 2013, Geochemistry and geochronology of the 1.3 Ga metatonalites from the Central Metasedimentary Belt boundary thrust zone in southern Ontario, Grenville Province, Canada: Geosphere, v. 9, p 853-863, doi: 10.1130/GES00868.1.

Bickford, M.E., McLelland, J.M., Selleck, B.W., Hill, B.M., and Heumann, M.J., 2008, Timing of anatexis in the eastern Adirondack Highlands; implications for tectonic evolution during ca. 1050 Ma Ottawan Orogenesis: Geological Society of America Bulletin, v. 120, p. 950-961, doi: 10.1130/B26309.1.

Brown, J. S., 1932, Natural gas, salt, and gypsum in Precambrian rocks at Edwards, New York: Bulletin of the Am. Assoc. of Petroleum Geologists, v. 16, p. 727-735.

Brown, C.E. and Ayuso, R.A., 1985, Significance of tourmaline-rich rocks in the Grenville Complex of St. Lawrence County, New York. U.S. Geol. Surv. Bull. 1626-C, p. 1-33.

Carl, J. D., 1988, Popple Hill Gneiss as dacite volcanics: A geochemical study of mesosome and leucosome, northwest Adirondacks, New York: Geological Society of America Bulletin, v. 100, p. 841–849.

Carl, J.D., and VanDiver, B.B., 1975, Precambrian Grenville alaskite bodies as ash-flow tuffs northwest Adirondacks, New York: Geological Society of America Bulletin v. 86, p. 1691-1707.

Carl, J.D., deLorraine, W.F., Mose, D.G., and Shieh, Y-N, 1990, Geochemical evidence for a revised Precambrian sequence in the northwest Adirondacks, New York: Geological Society of America Bulletin, v. 102, p. 182-192.

Chamberlain, S., Lupulescu, M., Bailey, D., deLorraine, W., and Robinson, G., 2018, Collector's Guide to the Balmat Mining District, St. Lawrence County, New York. Schiffer Publishing Ltd., ISBN: 978-0-7643-5466-3.

Chiarenzelli, J., Lupulescu, M., Robinson, G., Bailey, D., and Singer, J., 2019. Age and Origin of Silicocarbonate Pegmatites of the Adirondack Region: Minerals, v. 9, p. 508. https://doi.org/10.3390/min9090508

Chiarenzelli, J. R. and Selleck, B. W., 2016, Bedrock Geology of the Adirondack Region: Adirondack Journal of Environmental Studies, v. 21, Article 5.

Chiarenzelli, J., Kratzmann, D., Selleck, B., and de Lorraine, W., 2015, Age and provenance of Grenville Supergroup rocks, Trans-Adirondack Basin, constrained by detrital zircons: Geology, v. 43, p. 183–186. https://doi.org/10.1130/G36372.1

Chiarenzelli, J. R., Kratzmann, D., Selleck, B., Christoffersen, P., and Durham, A., 2013. Constraining the depositional history and evolution of the Trans-Adirondack Back-arc Basin using detrital zircon geochronology: Geological Society of America *Abstracts with Programs* Vol. 45, No. 7, p. 739.

Chiarenzelli, J., Hudson, M., Dahl, P., and deLorraine, W. D., 2012. Constraints on deposition in the Trans-Adirondack Basin, Northern New York: Composition and origin of the Popple Hill Gneiss: Precambrian Research, v. 214-215, p. 154-171. doi.org/10.1016/j.precamres.2011.10.024

Chiarenzelli, J., Lupulescu, M., Cousens, B., Thern, E., Coffin, L., and Regan, S., 2010a. Enriched Grenvillian lithospheric mantle as a consequence of long-lived subduction beneath Laurentia: *Geology*, v. 38, p. 151-154.

Chiarenzelli J., Regan, S., Peck, W., Selleck, B., Baird, G. and Shrady, C., 2010b, Shawinigan Magmatism in the Adirondack Lowlands as a Consequence of Closure of the Trans-Adirondack Back-Arc Basin: Geosphere, v. 6, p. 900-916.

Chiarenzelli, J., Lupulescu, M., Thern, E., and Cousens, B., 2011a, Tectonic Implications of the Discovery of a Shawinigan Ophiolite (Pyrites Complex) in the Adirondack Lowlands: Geosphere, v. 7, p. 333-356.

Corrigan, D., 1995, Mesoproterozoic evolution of the south-central Grenville orogen: Structural, metamorphic, and geochronological constraints from the Maurice tran-sect [Ph.D. thesis]: Ottawa, Carleton University, 308 p.

de Lorraine, W.F., 2001, Metamorphism, polydeformation, and extensive re-mobilization of the Balmat zinc orebodies, northwest Adirondacks, New York, in: Part I: Proterozoic Iron and Zinc Deposits of the Adirondack Mountains of New York and the New Jersey Highlands Part II: Environmental Geochemistry and Mining History of Massive Sulfide Deposits in the Vermont Copper Belt. , Society of Economic Geologists Guidebook Series. pp. 25–54.

de Lorraine, W.F, and Carl, J.D., 1993, Trip A-I Precambrian Geology of the northwest Adirondack Lowlands: a Stratigraphic Viewpoint. New York State Geological Association annual field conference.

de Lorraine, W.F, and Carl, J.D., 1986, Precambrian stratigraphy of the Northwest Adirondacks, New York: Friends of the Grenville guidebook for field trip, 37p.

de Lorraine, W. F. and Sangster, A. L., 1997, Geology of the Balmat Mine, New York: Field Trip A5, Geological Association of Canada/Mineralogical Association of Canada Joint Annual Meeting, Ottawa, 43p.

Dickin, A. P. and Strong, J. W., 2021, Nd isotope mapping of the Frontenac Terrane in southwestern Grenville Province. Canadian Journal of Earth Sciences, v. 58, p. 659-669. https://doi.org/10.1139/cjes-2020-0063

Dickin, A.P. and McNutt, R.H., 2007, The Central Metasedimentary Belt (Grenville Province) as a failed back-arc rift zone: Nd isotope evidence: Earth and Planetary Science Letters, v. 259, p. 97-106. doi: 10.1016/j.epsl.2007.04.031

Dickin, A., Hynes, E., Strong, J., and Wisbborg, M., 2016, Testing a back-arc aulacogen model for the Central Metasedimentary Belt of the Grenville Province. Geological Magazine, v. 153, p. 681-695, doi:10.1017/S0016756815000874.

Heumann, M.J., Bickford, M.E., Hill, B.M., McLelland, J.M., Selleck, B.W., and Jercinovic, M.J., 2006, Timing of anatexis in metapelites from the Adirondack lowlands and southern highlands: A manifestation of the Shawinigan orogeny and subsequent anorthosite-mangerite-charnockite-granite magmatism: Geological Society of America Bulletin, v. 118, p. 1283-1298, doi: 10.1130/B25927.1

Isachsen, Y. W. and Fisher, D., 1970, Adirondack Sheet. New York State Museum and Science Service Map and Chart Series, University of the State of New York, Albany, NY, 1:250,000 scale,

Isachsen, Y.W., and Landing, Ed, 1983, First Proterozoic stromatolites from the Adirondack massif-Stratigraphic, structural, and depositional implications [abs.]: Geological Society of America Abstracts with Programs, v. 15, no. 6, p. 601.

Matt, P., Powell, W., Mathur, R., and deLorraine, W., 2020. Zn-isotopic evidence for fluid-assisted ore remobilization at the Balmat Zinc Mine, NY, Ore Geology Reviews, v. 116, ISSN 0169-1368.

Matt, P., Powell, W., deLorraine, W. F., and Chiarenzelli, J., 2019, Sulfide and silicate anatexis in the Balmat zinc deposit (New York, USA) and its implications for ore remobilization: Ore Geology Reviews, v. 107, p. 392-401. https://doi.org/10.1016/j.oregeorev.2019.03.002.

McLelland, J., Selleck, B. and Bickford, M.E., 2010, Review of the Proterozoic evolution of the Grenville Provi, its Adirondack inlier, and the Mesoproterozoic inliers of the Appalachians,, *in* Tollo, R.P., Bartholmew, J.P., Hibbard, J.P., and Karabinos, P.M., *eds*., From Rodinia to Panges: The Lithotectonic Evolution of the Appalachian Region: Geological Society of America Memoir, v. 206, p. 21-49.

McLelland, J. M., Daly, J. S., and McLelland, J., 1996, The Grenville Orogenic Cycle (ca. 1350-1000 Ma): an Adirondack Perspective: Tectonophysics, v. 265, p. 1-28.

McLelland, J., Chiarenzelli, J., and Perham, A., 1992, Age, field, and petrological relationships of the Hyde School Gneiss, Adirondack Lowlands, New York: Criteria for an intrusive igneous origin: Journal of Geology, v. 100, p. 69–90, doi: 10.1086/629572.

McLelland, J. M. and Chiarenzelli, J. R., 1990, Geochronological studies of the Adirondack Mountains, and the implications of a Middle Proterozoic tonalite suite, in Gower, C., Rivers, T., and Ryan, C., eds., Mid-Proterozoic Laurentia-Baltica: Geological Association of Canada Special Paper 38, p. 175-194.

Moretton, K. and Dickin, A. P., 2013, Nd isotope mapping of the Dysart gneiss complex: Evidence for a rifted block within the Central Metasedimentary Belt of the Grenville Province: Precambrian Research, v. 228, p. 223-232.

Peck, W., Selleck, B., Wong, M., Chiarenzelli, J., Harpp, K., Hollocher, K., Lackey, J., Catalano, J., Regan, S., and Stocker, A., 2013, Orogenic to postorogenic (1.20–1.15 Ga) magmatism in the Adirondack Lowlands and Frontenac terrane, southern Grenville Province, USA and Canada: Geosphere, v. 9, p. 1637-1663.

Peck, W. H., Bickford, M. E., McLelland, J. M., Nagle, A. N., and Swarr, G. J., 2010, Mechanism of metamorphic zircon growth in a granulite-facies quartzite, Adirondack Highlands, Grenville Province, New York: American Mineralogist, v. 95, p. 1796-1806.

Peck, W. H., Volkert, R. A., Mansur, A. T., and Doverspike, B. A., 2009, Stable Isotope and Petrologic Evidence for the Origin of Regional Marble-Hosted Magnetite Deposits and the Zinc Deposits at Franklin and Sterling Hill, New Jersey Highlands, United States: Economic Geology, v. 104, p. 1037–1054. doi: https://doi.org/10.2113/econgeo.104.7.1037

Peck, W.H., Valley, J.W., Corriveau, L., Davidson, A., McLelland, J., Farber, D.A., 2004. Oxygen-isotope constraints on terrane boundaries and origin of 1.18–1.13 Ga granitoids in the southern Grenville Province, in: Tollo, RP, Corriveau, L, McLelland, J, and Bartholomew, MJ, (Eds.), Proterozoic tectonic evolution of the Grenville orogen in North America: Boulder, Colorado, Geological Society of America Memoir 197, p. 163-182.

Prucha, J., 1957, Pyrite deposits of St. Lawrence and Jefferson Counties, New York: Albany, University of the State of New York, 87p.

Revetta, F.A., and McDermott, A. M., 2003, The compilation and preparation of high-resolution gravity data for petroleum exploration in New York State and adjoining regions: New York State Energy Research Development Authority (NYSERDA) PON 715-02.

Smith, T. E. and Harris, M. J., 1996, The Queensborough mafic-ultramafic complex: a fragment of a Meso-Proterozoic ophiolite? Grenville Province, Canada: Tectonophysics, v. 265, p. 53-82.

Strong, J. and Dickin, A., 2021, Neodymium isotope mapping a polygenetic TTG batholith: Failed back-arc rifting in the Central Metasedimentary Belt, SW Grenville Province. Canadian Journal of Earth Sciences, v. 59, p. 71-86. 10.1139/cjes-2021-0061.

Tiedt, K., and Kelson, C., 2008, Geochemical and mineralogical characterization of ore from the Pyrites and Stella pyrite deposits, St. Lawrence County, New York: Geological Society of America Abstracts with Programs, v. 40, p. 6.

Valentino, D. W., Chiarenzelli, J. R., and Regan, S. P., 2019, Spatial and Temporal Links Between Shawinigan Accretionary Orogenesis and Massif Anorthosite Intrusion, southern Grenville Province, New York, U.S.A.: Journal of Geodynamics, v. 129, p. 80-97, https://doi.org/10.1016/j.jog.2018.04.001.

Wasteneys, H., McLelland, J.M., and Lumbers, S., 1999, Precise zircon geochronology in the Adirondack Lowlands and implications for revising plate-tectonic models of the Central Metasedimentary belt and Adirondack Mountains, Grenville Province, Ontario and New York: Canadian Journal of Earth Sciences, v. 36, p. 967–984, doi: 10.1139/cjes-36-6-967.

Wiener, R.W., McLelland, J.M., Isachsen, Y.W., and Hall, L.M., 1984, Stratigraphy and structural geology of theAdirondack Mountains, New York, in Bartholomew,M., ed., The Grenville Event in the Appalachians andRelated Topics, Review and Synthesis: Geological Society of America Special Paper 194, p. 1–55.

Wong, M. S., Peck, W.H., Selleck, B.W., Catalano, J.P., Hochman, S.D., and Maurer, J.T., 2011, The Black Lake shear zone: A boundary between terranes in the Adirondack Lowlands, Grenville Province: Precambrian Research, v. 188, 57-72, doi:10.1016/j.precamres.2011.04.001

Wynne-Edwards, H.R., 1972, ‘The Grenville Province.’ In: Variations in Tectonic Styles in Canada, R.A. Price and RJ.W. Douglas, (eds.), *Geological Association of Canada, Special Paper*, no. 11, p. 263–334.

Trip B-2

EASTERN LAKE ONTARIO DUNE-WETLAND COMPLEX FROM THE PLEISTOCENE TO TODAY: LAURENTIDE ICE SHEET DEGLACIATION, SHORELINE EVOLUTION AND MODEN GEOECOLOGY

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INTRODUCTION

The Eastern Lake Ontario Dune Complex and associated landforms, coastal morphology and wetlands have survived through constant change since the Late Pleistocene. While the general location of the Lake Ontario shoreline is the result of deglaciation of the Laurentide Ice Sheet (16,000-12,000 YBP) and drainage of Glacial Lake Iroquois, the natural processes acting upon the 27 km stretch of dunes is modified further by complexities of human interaction with the landscape since European settlement post-colonization. Today, natural processes such as longshore transport of sediment, climatic influence on lake level, erosion at the headwaters of Lake Ontario tributaries, and human intervention such as lake level controls and shoreline erosion management are in discordance with one another. The natural climatic and geologic processes in the context of human settlement along the lake creates a balancing act in which human settlements attempt to remain resilient while conserving and protecting delicate coastal ecosystems.

GEOLOGIC SETTING AND CHRONOLOGY

Glacial Lake Iroquois – Modern Lake Ontario

The previous stages of Glacial Lake Iroquois (proto-Lake Ontario, 16,000-13,000 cal YBP), sourced by glacial meltwater, exceeded the modern elevation of today's lakes by 60-75m (Bird and Kozlowski, 2016). The multitude of shoreline elevation changes in the late Quaternary have been attributed to repeated glacial lake meltwater events, periodically draining these lakes through the Iro-Mohawk River system (14.8 +/- 1.3 ka, Porecca, et. al 2018). Potholes from turbulent meltwater flow along Moss Island in Central New York have been attributed to these drainage events.

With additional northward retreat and a subsequent drop in the level of Glacial Lake Iroquois to that of modern Lake Ontario, a lower-elevation opening of the St. Lawrence Seaway eliminated the Iro-Mohawk River drainage, re-routing instead to the Northern Atlantic through the St. Lawrence River, which has continued into the present interglacial period.

Figure 1. *Map view with North oriented right of Eastern Lake Ontario Dune Complex, showing the entire Eastern Lake Ontario Dune Complex and field trip stops (orange.)*

<u>*Development of Dunes*</u>

As Glacial Lake Iroquois transitioned to modern Lake Ontario (13,000-present), remnant lakeshores remained across the New York and Ontario landscape, with similar processes and timing taking place for Glacial Lake Warren (Young et al., 2021). Woody debris found in situ along these remnant shorelines or buried in ice-contact lakes, coupled with lake cores depicting mineralogical/organic transitions have been used in constraining the ages of these events (Lewis and Anderson, 2020; Young et al., 2021). With the significant drop in elevation from the glacial lakes to the modern lakes, sand and silt composing the lake bottoms was exposed to the prevailing winds and dry conditions of the Late Pleistocene and Early Holocene in New York, resulting in the aeolian deposition of sand along the eastern Lake Ontario shoreline. The resulting complex includes young foredunes (beach-adjacent), interior dunes (middle aged, vegetated with young shrubs and trees) and older backdunes (capable of supporting large trees such as red oak and large cottonwood.) Additional dating of these sand dunes through OSL (optically-stimulated luminescence) could aid in determining the ages and lifecycles of these dunes along Eastern Lake Ontario. Additional geochronology could be pursued by constraining the mineralogical-biologic transition of Lake Ontario-adjacent wetlands and ponds.

Figure 2. *Young cottonwoods mixed in with dune grass populate the interior dune at Black Pond WMA.*

Sediments Sources

In addition to offshore sand bars in Lake Ontario, the shoreline of the lake and upland areas contribute to the sediments sources for Lake Ontario's eastern shore dunes. The southern shoreline tributaries of the Genesee and Oswego Rivers drain thousands of acres of farmland in New York State alone, adding to the sediment sources within the lakes, although much of this sediment is captured in shallow-basin bays along southern and eastern Lake Ontario, often protected from longshore currents by barrier bars. With the added erosion of drumlins (elongated, often teardrop-shaped subglacial landforms), there remains a constant sediment source for maintaining the eastern Lake Ontario dune complex, as well as the coastal features (barrier bars and spits) notorious throughout Wayne County and the spit complex at North Sandy Pond in Oswego County. More detailed studies on these processes are being proposed and are underway.

Figure 3. *Looking southeast towards the Southern Spit at North Sandy Pond, December 2020. Constant sand transport towards eastern Lake Ontario results in seasonal deposition within the only active inlet to the pond, which is dredged annually by the Town of Sandy Creek and other entities. Dredged material is used to re-nourish breaches within the spit and in dune restoration projects.*

Dunes along eastern Lake Ontario remain in a constant state of change while balancing lake levels, sediment sources, environmental and climatic changes while under the influence of human populations. During recent flooding events in 2017 and 2019, many of these delicate dunes were significantly eroded, while the spit at North Sandy Pond experienced multiple dune blowouts and breaches, with other coastal formations such as the Blind Sodus Bay barrier bar completely eroding away, opening previously-closed bays to the forces of Lake Ontario and further, significant erosion and flooding of inland shorelines.

While growth and destruction of incipient foredunes is common and expected, destruction and rebuilding of backdunes is not. When erosion of these backdunes occurs, whether through deforestation, blowouts or other processes, the dune is often lost for good as the sediment supply is cut off or too far for Aeolian transport to remain viable. Subsequent stabilization of the dunes is possible through the return of vegetation to the system and dune fencing.

Dune-building Processes on Eastern Lake Ontario

With the available sand source from Glacial Lake Iroquois and the offshore sand bars in the newly-established Lake Ontario, winds across the basin transport sand and silt to the shoreline, while waves move more sediment to the nearshore and beaches. These sediments, when exposed along the shoreline, dry and become entrained with the wind and begin traveling up the shoreline to begin the dune forming process. When encountering an object such as vegetation, woody debris or rocks, the winds slow down and lose their capacity to carry sediment, and sand grains are deposited. Primary dune vegetation such as Champlain Beach Grass and creepers begin the stabilization process by trapping sediment and sending roots into it, forming the *incipient dune* or *foredune.* Secondary vegetation in the form of fast-growing shrubs and trees (sand cherry, dune willow, shrub willows, cottonwood) begin to thrive in the *foredune* environment. As these trees and shrubs enter mid-life and later, the *hind dune* or *backdune* is fully stabilized, beginning the creation of shallow forest soils and growth of larger trees and a more diverse (oak species, maple, etc.) forest. Storm surge, lake levels and wind events will often erode the fronting beach dunes, leaving the face of the dune eroded to the angle of repose (with slight influence based on vegetation). This erosion will continue until the dune face has stabilized or wind-blown sediment is returned to the system, often in the form of a drop in lake level or artificial nourishment.

Figure 3. *Dune grass and fencing at Southwick Beach State Park, part of a dune restoration project.*

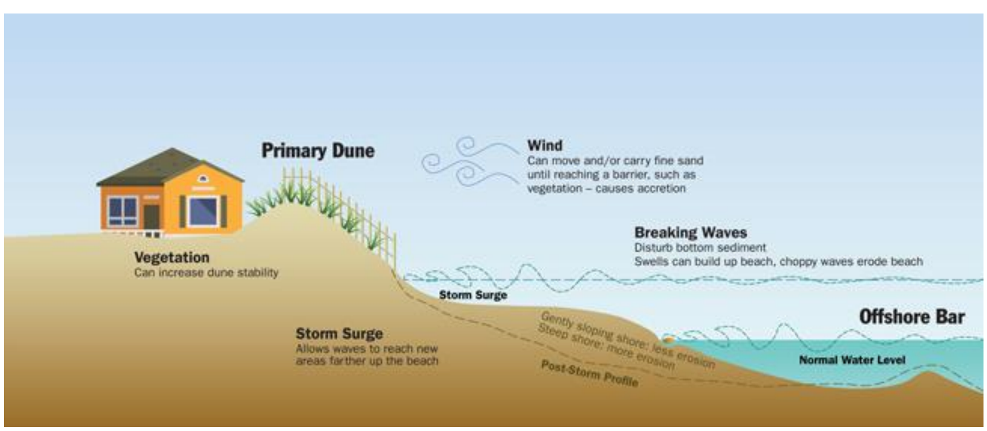


Figure 4. *Basic Dune-forming processes adapted from Long Island (from Fallon, 2022.)*

Dune Blowouts

Across the dune complex, dune blowouts may occur. Blowouts are depressed features within previously-vegetated dunes, where wind erosion has become accelerated and cuts into the dune perpendicular to the shore. In many cases this process is naturally occurring, a result of funneled winds along the shoreline. However, in Oswego County in particular, human activity along the dunes (walking over and destroying dune grasses, motorized vehicle transport, or the erection of buildings along the shoreline) has caused significant blowout events. Overuse of all-terrain vehicles in the 1980s resulted in significant dune blowouts, necessitating the formation of the Eastern Lake Ontario Dune Coalition to attempt to control human destruction of the dunes and to begin conservation programs locally. Remnants of the blowouts, both natural and unnatural, are still visible today and can be seen at Stops 1 and 3 (Black Pond WMA, Sandy Island Beach State Park.) In order to reduce the erosion of sand dunes through dune blowouts, multiple public and private projects have been executed to construct boardwalks or dune overwalks throughout the region, allowing dune grass to live while also securing human access to coastal areas. Additionally, dune fencing is used to slow the transport of sand grains along the shoreline. When coupled with planting of dune grass, dune fencing can help re-establish dunes within 2-5 years.

Figure 5: *Recent dune blowout at Black Pond WMA.*

DEVELOPMENT OF WETLAND COMMUNITIES AND MODERN COASTAL MORPHOLOGY

While dune complexes drain exceptionally well, wetlands are often associated with areas such as Eastern Lake Ontario as a result of multiple geological and hydrological processes. These processes have been fluctuating with climate trends since the last glacial maximum (LGM), and are further complicated by ecological processes and human intervention.

In most cases along eastern Lake Ontario, these wetland complexes have formed in areas where drainages meet an elevation impediment near the lakeshore, often the dunes themselves. This is starkly visible at Deer Creek Marsh Wildlife Management Area (WMA, Stop 2.) Deer Creek has headwaters just south of the Winona State Forest in Sandy Creek, but drains a significant portion of Oswego County before entering the WMA, a 1771 acre marsh, nestled between NYS-3 and shoreline dunes. The creek itself bends southward once meeting the dunes, and travels south for around 1 km before reaching Lake Ontario, where the creek only seldom drains into the lake directly. Reworked lake sediments often form a complex transitional area at Brennan Beach (private) impounding the creek.

With limited drainage and flushing of these coastal tributaries into Lake Ontario, wetland communities abound in the inland reaches behind back dunes. Cattail marshes are common, as well as forest and early successional grassland and shrublands. The dunes shelter these wetland areas, and frequent deposition of wind-blown sand creates new areas for pioneer species such as Champlain beach grass (*Ammophila brevigulata ssp. champlainensis),* before being populated by dune cherry (*Prunus pumila*), dune willow (*Salix cordata),* paper birch, (*Betula papyrifera*) eastern cottonwood (*Populus deltoides*), and eventually red oak (*Quercus rubra*).

Figure 6. *Open water and swamp at Deer Creek Marsh Wildlife Management Area as seen from lookout deck.*

Inland from the dunes are deep emergent and shallow emergent marshes, as well as deep water ponds and swamps (North Sandy Pond, South Pond, Lakeview Pond). In shallow emergent marshes (3”-6’ in depth), grasses and grass-like sedges dominate, along with tall wildflowers such as Joe Pye Weed and shrubs, including red osier dogwood, shrub willows and alder (Figure 7.) In the deeper emergent wetlands, cattails and pickerelweed are common, while larger fauna such as muskrats, wood ducks and other migratory waterfowl can be seen.

<u>Coastal Fens</u>

Fens are rare plant communities living above layers of peat. Composed of sedges and shrubs, fens are sometimes floating masses, and feature unique plants and animals not found in other biomes. Within the Eastern Lake Ontario Dune Complex, there are locally- and globally-rare species, including the bog buckbean (*Menyanthes trifoliate)* and the endangered bog buckmoth, (*Hemileuca spp.)* which form a symbiotic relationship. Of the ten known colonies of bog buckmoth throughout the world, six occur in Oswego County (Bonnano, 1997.) Also found within the fens of Oswego County are the insect-eating plants sundew (*Drosera sp.)* and pitcher plant (*Sarracenia purpurea)*.

Other Coastal Geologic Features of Eastern Lake Ontario

The Eastern Lake Ontario Dune Complex is defined by a 27 km stretch of coastal dunes and wetlands, but also features the North Sandy Pond Spit, a barrier complex with a pair of recurved spits (Mattheus, et al., 2016) and dynamic inlet/outlet changes well-documented throughout written history in the region (Figure 3.) Here, the ample sediment supply in eastern Lake Ontario provides sand to the eastern shore, with occasional breaches through the narrower stretches of the ~10-300m wide barrier bar. The main inlet has been kept remarkably consistent since the 1950s, and dredging of the inlet and sand shoal has maintained this inlet in recent years. Since 1950 multiple breaches have occurred which either self-healed or were repaired by means of artificial sand nourishment along the shoreline.

Further north at Black Pond-El Dorado, the dune complex is abruptly truncated by the mouth of Black Pond outlet. Here is one of the few locations along Lake Ontario where bedrock (Chaumont Limestone, Black River Group, middle Ordovician) is exposed at the surface. The outcrop, accessible from Stop 1 at Black Pond WMA, is believed to act as a dam, keeping the dunes constrained to the south of the feature.

Shoreline Management

With increasing development pressure in the region and the transition from seasonal to full-time residency along the shoreline, the dunes along with eastern shore of Lake Ontario are subjected to a myriad of shoreline management strategies, some of which have detrimental effects of the coastal processes and ecology of the region. Prior to European settlement, this was not the case. Greene Point, located within North Sandy Pond, remains one of the first known European settlements in the area, being founded in 1803, and began the cycles of deforestation and the building of permanent structures before recreation and tourism became a dominant industry by the 1870s.

Figure 7. *Limestone rip-rap and dune revegetation on display along the south spit, October 2021.*

In the human need of controlling the environment, shoreline stabilization was inevitable. The dune complex remains highly active and at times unpredictable, and stabilization of inlets and shorelines to match property boundaries was deemed necessary to maintain erosion-resilient communities along the Lake Ontario shoreline. However, due to their dynamic nature, dunes and sandy shorelines are notoriously difficult to control in a geomorphic context. Since cottage development in the 1950s, shoreline residents have attempted many solutions to solving the shifting sands: wooden plank cribbing, gabion baskets (metal mesh containers filled with stones), rock rip-rap (large boulders parallel to the shore) and nature-based restoration (mimicking the natural environment to stabilize erosion, Figure 8) remain the most common approaches, while physical movement of structures away from coastal erosion hazard areas has been discussed extensively, yet remains unviable in many situations and unpalatable to residents in others. Shoreline residents are not encouraged to move away from the shoreline, as in many cases, these residents provide a significant portion of the tax bases for many Great Lakes counties, towns and villages.

While heavily engineered solutions to shoreline erosion exist throughout the dune complex, longshore sand transport remains efficient along the eastern shore. However, further hardening of the shoreline may endanger these processes. In many situations, these hard structures along the shoreline act in opposition to the coastal processes needed to maintain the dunes and barrier spit morphology. Hard shorelines often cause scouring at the base of structures, moving sand off the shoreline and out into the lake where it becomes inaccessible to waves, failing to bring the much-needed sediment to the shoreline. Additionally, these walls have been shown to accelerate erosion along adjacent shorelines (Figure 9) and act as a deterrent to beneficial native wildlife (Gittman et al., 2015) and an attractant to nuisance species such as Canadian geese and cormorants.

Figure 8. *Headland created by artificial shoreline stabilization techniques, South Spit of Sandy Pond. Here, waves have been deflected away from the shoreline and into unprotected shoreline areas, resulting in additional, accelerated erosion along the flanks of the rip-rap wall.*

Shoreline Restoration

In recent years a trend has occurred in which shoreline restoration, rather than manipulation through shoreline hardening, has become an emergent shoreline management technique. This is evident throughout the dune complex, but is highlighted by natural resiliency projects at Southwick Beach State Park and along both private and public-owned land at the North Spit of Sandy Pond, part of Sandy Island Beach State Park.

At Southwick Beach, significant shoreline dunes were lost in the flooding and erosion as a result of historically-high lake levels in both 2017 and 2019. In order to restore these shorelines to a status worthy of presenting in a New York State park, dune fencing and significant planting of dune grass occurred between 2019-2021, with more plantings planned in the future. This minimalist approach has led to a nearly complete restoration of the previous shorelines. Additional funding and planning has also implemented stormwater basins, grassed waterways and raingardens in and around parking areas in the park to mitigate flooding events in the future.

Along the North Spit (Figure 10), significant breaches occurred in 2017 and 2019, occurring close to residential properties in the north. Local communities and organizations were mobilized to action, and received grants from the State of New York to begin a dredge-and-place project along the shoreline to repair the breaches in the barrier, install dune fencing to catch wind-blown sediment, and provide numerous spring and fall dune grass plantings with community volunteers. Dredge material is sourced from the nearby North Sandy Pond inlet, where a navigation channel is maintained (visible in Figure 3.)

Through three phases in 2020 a total of 40,000 yd^3 of dredged material was placed along the shoreline to repair the breaches. The sediment was almost immediately mobilized by the cross-lake winds, recovering the beach and barrier bar to greater widths than observed in previous years (Figure 10). The barrier is continuously monitored by a dedicated group of volunteers and organizations associated with the Eastern Lake Ontario Dune Coalition and the project is used as a demonstration project for beneficial use of dredged materials and restoration of Great Lakes shorelines through natural and nature-based methods.

Figure 9. *Aerial view of beach nourishment project, north spit of Sandy Pond, Summer 2021. Here, sand dredged from the channel (background, far left) and placed along the shorelines (in front of dune fencing) and*

allowed to blow across the narrow spit, depositing in incipient dunes and overwash deposits. Prior to 2019, the spit was wooded along its length, but was washed out completely from storm events in 2017 and 2019.

In other attempts at shoreline restoration and projection in the Oswego County, there has been less success. At Sandy Island Beach State Park (Stop 3), a combination of heavily-engineered and nature-based features were used to restore the dunes at this location. Similar to the North Spit, the shoreline here was eroded heavily in 2017. However, to maintain local infrastructure including park buildings, an access road to South Pond, and a parking lot, an engineered solution was deemed necessary. This included the installation of trap bags (geotextile bags filled with sand), toe protection through limestone rip-rap, and fill brought in from an upland source. Unfortunately in 2019, water levels exceeded previous highs witnessed in 2017, and the rip-rap was quickly overtopped by waves, eroding the filled material up to the installed trap bags, resulting in a significant loss of land (Figure 11.) Additionally, the upland sediment source used as fill was not suitable for dune-building, and was unable to be entrained within the winds enough to contribute to the success of the dune grass plantings, which have struggled. This location is now being studied extensively to develop new solutions to the local erosion issues while maintaining a resilient shoreline for the local infrastructure.

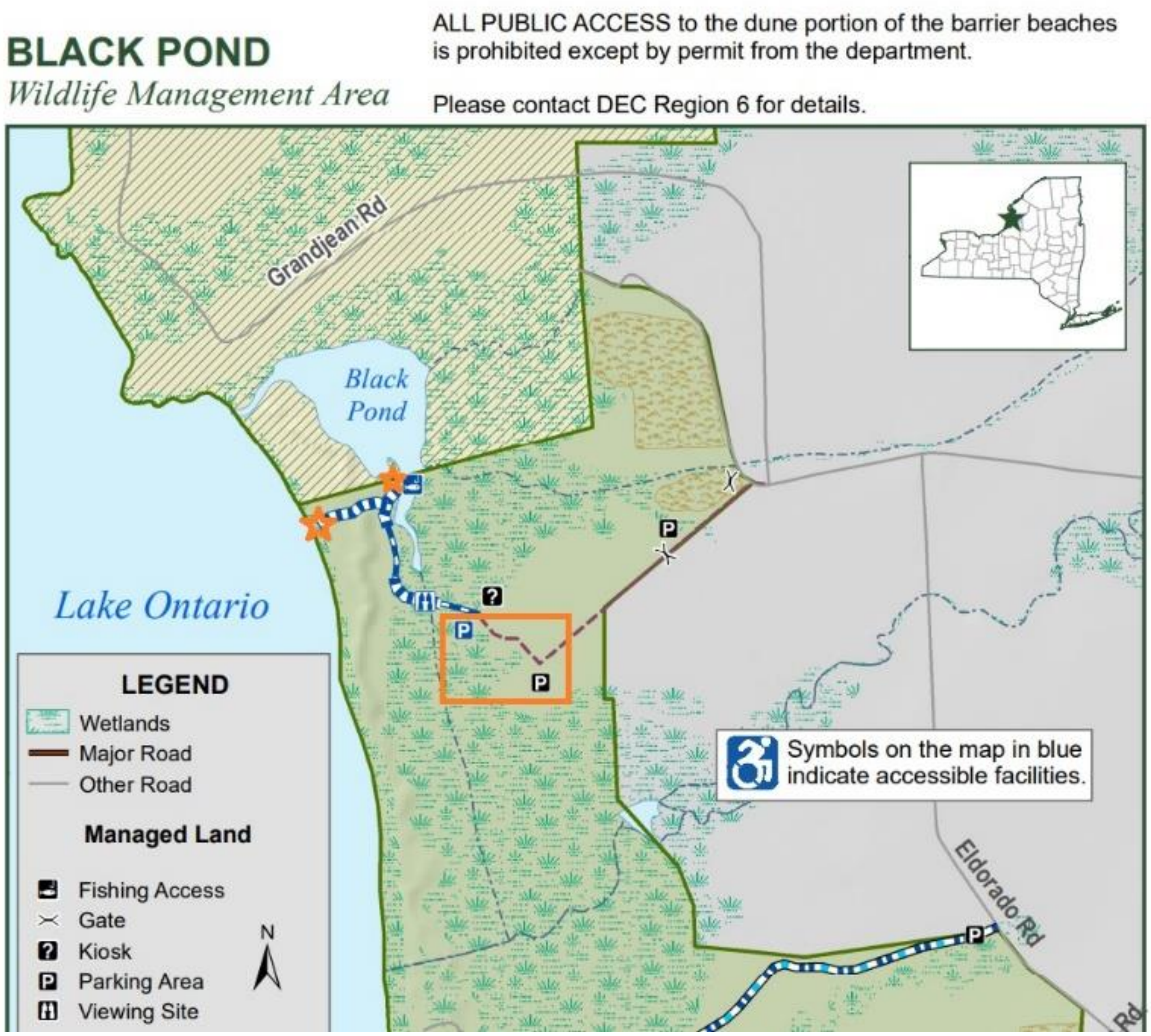
Figure 10. *Failed stabilization attempt at Sandy Island Beach, 2019. Here, large, coarse rip-rap was introduced to reduce erosion of an eroding dune face, but deflected and intensified erosion-causing waves into the inadequate upland sediment (not suitable for dune material) resulting in an immediate and excessive loss of shoreline material.*

FIELD GUIDE AND ROAD LOG

STOP 1 / Meeting Location: Black Pond Wildlife Management Area (43.802162, -76.224055)

Directions (55 minutes): From SUNY Oswego, take NY-104 E to NYS-104B E (10 miles) to the junction with NY-3 E (left-hand turn at 6 miles). Once on NY-3 E, follow for 22.4 miles to Bolton Rd (left). Follow Bolton Rd to Black Pond WMA Access Rd after about 1 mile. Take access road about .5 mile to parking lot. *Less than a mile walk over boardwalk and trail to shoreline.*

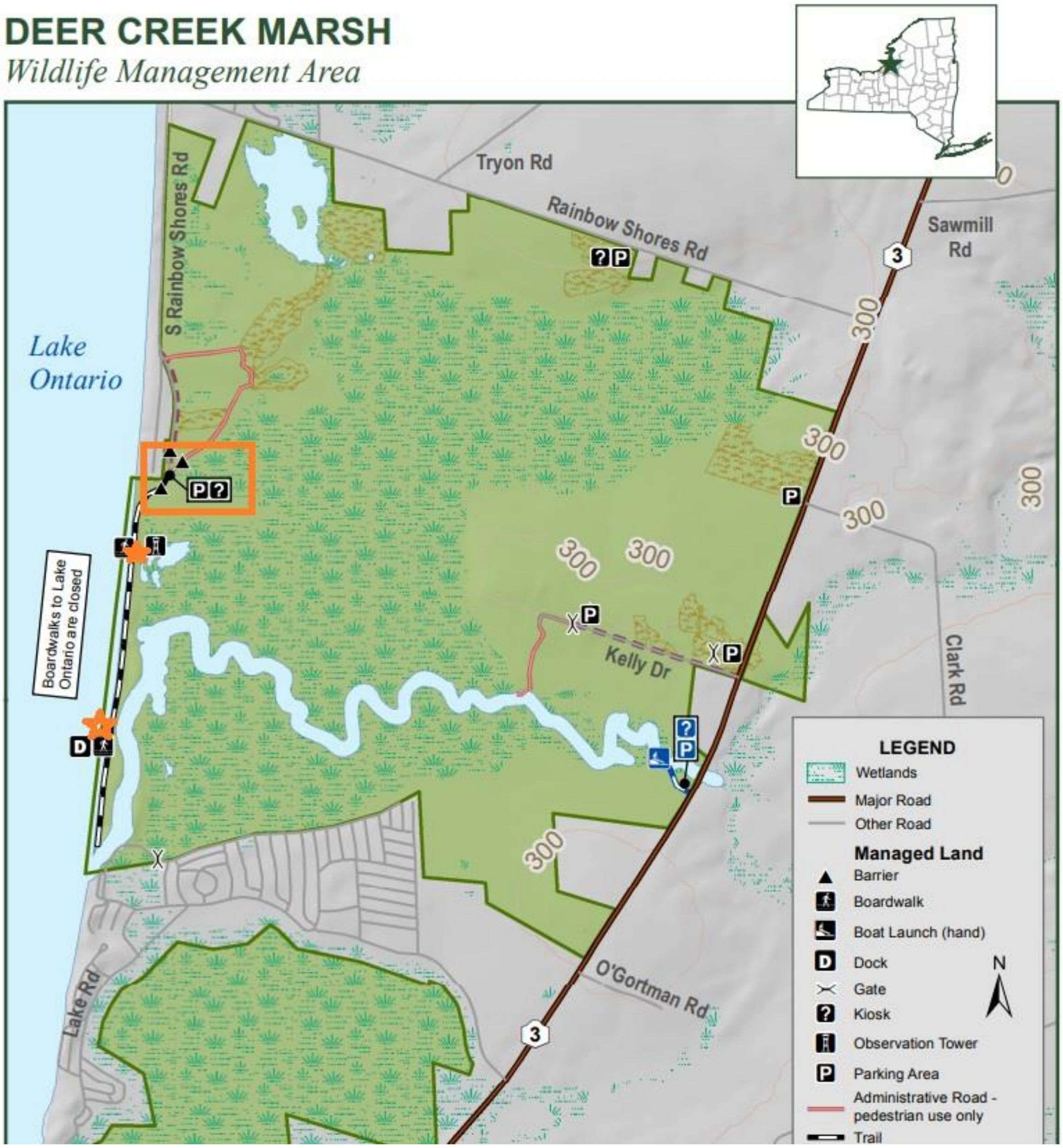
Discussion: Eastern Lake Ontario Dune Complex – Dune Formation and Modern Habitats



STOP 2
Deer Creek Marsh Wildlife Management Area - North (43.600754, -76.197299)

Directions (30 minutes): From Black Pond WMA, take access road and Bolton Rd. back to NY-3 W. After right on to NY-3 W, take for 15 miles to Rainbow Shores Rd (right.) Follow Rainbow Shores Rd. for 1.8 miles to Rainbow Shores Rd. South (left). After about .5 mile, road will diverge – take left for another .3 mile to the parking lot at Deer Creek marsh WMA. *Walking over sand and boardwalk for about 1-2 miles, depending on stops.*

Discussion: Sand dune and wetland management, dune system hydrology.



Eastern Lake Ontario Dune Complex – Coastal Geomorphology

STOP 3

Sandy Island Beach State Park – Main Parking Lot (43.630755, -76.195224)

Directions (12 minutes): From Rainbow Shores Rd., turn left on to Tryon Rd. after .7 miles. After 1.4 miles, turn left on to Ouderkirk Rd. Follow Ouderkirk Rd. to the T with County Route 15, and take a left. Sandy Island Beach State Park is .3 mile down this road, after crossing the bridge, use parking lot to the right.

Discussion: Human Intervention and Conservation in Dune Systems

STOP 4 (OPTIONAL/LIMITED PARKING)

Salmon River Bridge – DEC Launch Parking Lot

Directions (11 minutes): From Sandy Island Beach State Park, take County Route 15 to NY-3 W (2.3 miles). Take NY-3 W for 4.7 miles to fishing access/parking lot on the north side of the NY-3 Bridge over the Salmon River. NOTE: There is only space for nine vehicles here and NY-3 is very busy. This stop is optional.

Discussion: Dunes, recreation, politics and conservation.

ACKNOWLEDGEMENTS

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REFERENCES

Bird, B. and Kozlowski, A., 2016. Late Quaternary Reconstruction of Lake Iroquois in the Ontario Basin of New York. New York State Museum & Chart 80.

Bonanno, S.E. 1997. Status of studies on the bog buckmoth (*Hemileuca sp.1: Saturniidae*). Report to U.S. Fish and Wildlife Service. 7p.

Fallon, K.F., 2022. Dunes – A Strong Companion in the Face of Storms. New York Sea Grant Publication. https://seagrant.sunysb.edu/articles/t/dunes-a-strong-companion-in-the-face-of-storms-coastal-processes-hazards-news

Gabel, S.L., DeHollander, J., Venczel, R., and White, D. 2005. Monitoring Coastal Change: Eastern Lake Ontario. New York State Geological Association, 77th Annual Meeting Guidebook.

Gittman, R.K., Fodrie, F.J., Popowich, A.M., Keller, D.A., Bruno, J.F., Currin, C.A., Peterson, C.H. and Piehler, M.F., 2015. Engineering away our natural defenses: an analysis of shoreline hardening in the US. Frontiers in Ecology and the Environment 13, 301-307.

Lewis C.F.M. and Anderson, T.W., 2020. A Younger Glacial Lake Iroquois in the Lake Ontario Bsain, Ontario and New York: re-examination of pollen stratigraphy and radiocarbon dating. Canadian Journal of Earth Sciences 57: 453-463.

Mattheus, C.R., Fowler, J.K., Diggins, T.P., and Allen, D.F., 2016. Barrier-spit geomorphology and inlet dynamics in the absence of tides: evolution of the North Pond System, eastern Lake Ontario, New York State. Earth Surface Processes and Landforms, 41, 1386-1398.

Porecca, C., Briner, J.P., and Kozlowski, A. 2018. Laurentide ice sheet meltwater routing along the Iro-Mohawk River, eastern New York, USA. Geomorphology 303, pg. 155-161.

Young, R.A., Gordon, L.M., Owen, L.A., Huot, S. and Zerfas, T.D., 2021. Evidence for a late glacial advance near the beginning of the Younger Dryas in Western New York State: An event postdating the record for local Laurentide Ice Sheet recession. Geosphere 17, 271-305.

Trip B-3

MIDDLE ORDOVICIAN CLASTICS, FRACTURES, FAULTS AND WATER FALL: GEOLOGY OF THE SALMON RIVER GORGE, ALMAR, NEW YORK

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INTRODUCTION TO THE REGIONAL GEOLOGY

The Tug Hill plateau is underlain by the Lorraine Group, that consists of a number of siliciclastic rock formations from oldest to youngest, the Utica, Whetstone Gulf, Pulaski and Oswego. The Utica shale and Oswego sandstone are end-members in a stratigraphic sequence that has increasing amount of siltstone and then sandstone beds upward progressively. This sequence of rocks represent the middle Ordovician trough that filled with clastic sediments in response to the Taconian orogeny, the Taconic Clastic Wedge. The high region of Tug Hill plateau is drained by rivers and streams that flow radially from the plateau in all directions, and most of these drainages have carved deep gulfs and gorges exposing the carbonates of the Trenton Group and the Lorraine Group clastics. The Salmon River flows from the plateau interior into the Redfield Reservoir. Downstream of the reservoir, the river flows over the top of the Oswego sandstone to the Salomon River falls (Figure 1). At the falls, the river plunges 33 meters into the top of the Pulaski Formation (interbedded sandstone, siltstone and shale), and forms the Salmon River gorge downstream. The river is bound by the gorge within the geographic limits of the Oswego formation. Exiting the Oswego formation, the Salmon River flows through a generally shallow banked cut in the Pulaski Formation, until it reaches Lake Ontario at the town of Port Ontario. This field trip is a visit to the Salomon River falls to examine the transition in strata near the top of the Taconic Clastic Wedge, structures in the strata, and how these features of the bedrock have contributed to the formation of the Salmon River falls and controlled the flow direction of the river. This trip includes two locations: 1. Top of the falls; 2. Below the falls (Figure 1).

Figure 1. *Oblique aerial view of the Salmon River falls, Oswego County, New York.*

THE TUG HILL PLATEAU

Situated between the Adirondack dome to the east and Lake Ontario to the west, the Tug Hill plateau rises about 500 meters from the lake shore to the highest point (Figure 2). The plateau covers an area approximately 3500 square kilometers and is underlain by the middle Ordovician strata that sit nonconformably on the Mesoproterozoic Adirondack basement. These strata are inclined approximately 2^{o} to 5^{o} toward the southwest as the result of faulting in the basement during the uplift of the Adirondacks (Roden-Tice, 2000; Roden-Tice et al., 2009) and Tug Hill plateau (Wallach and Reault, 2010). Approximately 500 meters of strata are exposed on the flanks of the plateau through river-beds, gulfs, road outcrops, and quarries, and the interior of the plateau is covered by thick ground moraine, drumlins and wetlands. There are long linear escarpments and valleys on the southern, northern, and eastern flanks of the plateau, that are the locations of unexposed faults (Wallach and Rheault, 2010). The southern flank of the plateau is bound by the Prospect Fault (Jacobi, 2002) and the northern flank of the plateau is coincident with the southwestern extension of the Carthage-Colton shear zone (Wallach and Rheault, 2010). Finally, the eastern flank of the plateau is the steepest slope, and the location of the proposed Black River fault (Wallach and Rheault, 2010). Superficially, it appears that the general geomorphology of the Tug Hill plateau was controlled by faults within the underlying basement that extend upward into the overlying sedimentary strata. The Salomon River forms the most pronounced cut into the southwestern margin of the plateau and the linearity suggests fault control (Figure 2).

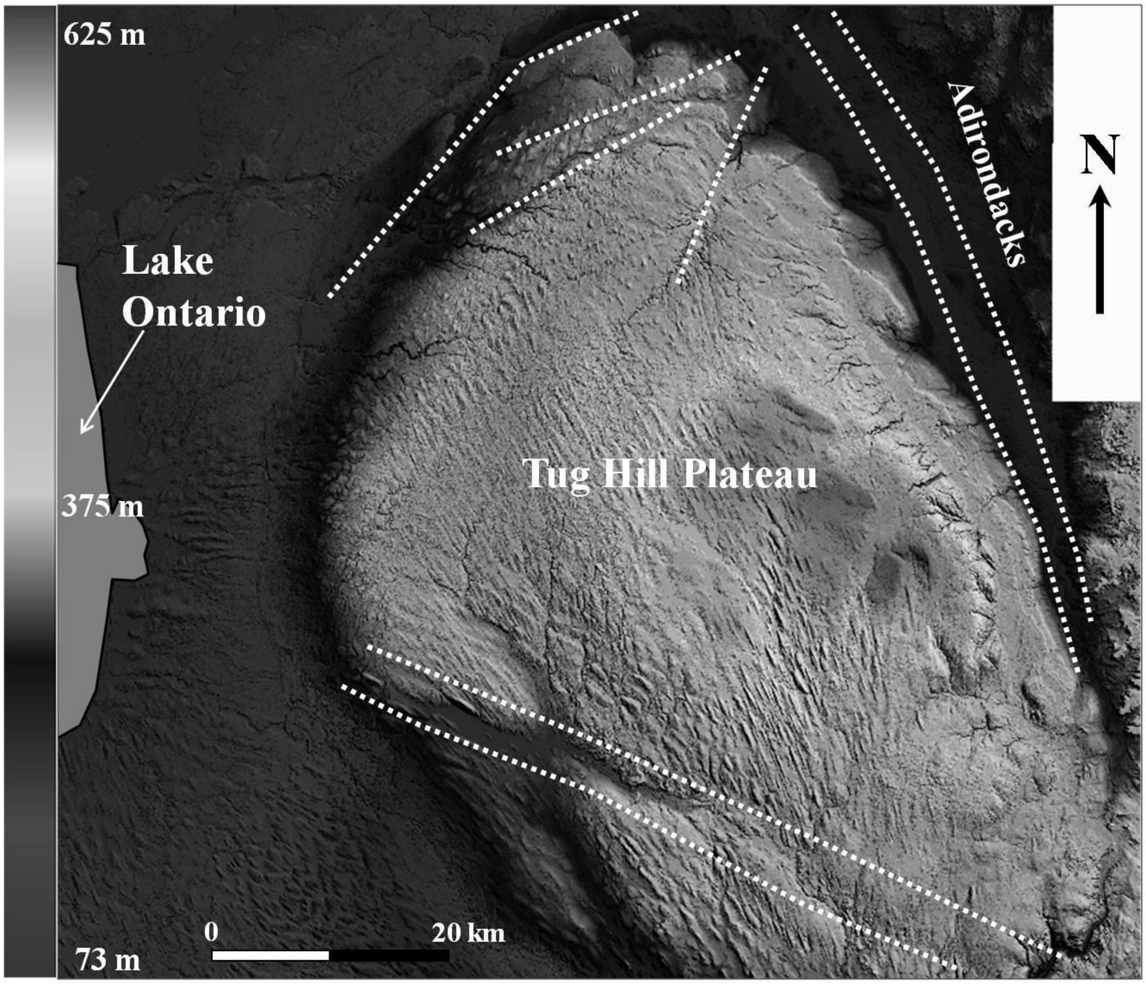

Figure 2. *Digital elevation model for the Tug Hill plateau. The dashed lines are proposed faults of Wallach and Reault (2010).*

TUG HILL PLATEAU GENERAL STRATIGRAPHY

The middle Ordovician strata of the Tug Hill plateau record the transition from marine to terrestrial deposition associated with carbonate sedimentation during the late stage of the Laurentian passive margin (Black River and Trenton groups), and siliciclastic deposition related to the onset of the Taconic orogeny (Lorraine Group) (Patchen, 1966; 1978) (Figure 3). The Tug Hill plateau sequence includes the Black River and Trenton Group carbonates that reside nonconformably on the crystalline basement, and the unconformity is exposed in the Black River valley. The thick limestone sequence that makes up the Trenton Group is directly overlain by the Utica black shale across an abrupt disconformity contact. With a progressive increase in the occurrence of siltstone beds, the Utica Formation transitions upward into the Whetstone Gulf Formation, and in turn the Whetstone Gulf Formation grades upward into the Pulaski Formation with the increase in the number of sandstone beds. Finally, with a marked decrease in the number of shale beds, the Pulaski Formation grades upward into the thin to thick bedded sandstone of the Oswego Formation. Although earlier geologic maps (Isachsen and Fisher, 1970) show a few isolated occurrences of Queenston shale, the cap-rock for the Tug Hill plateau is primarily the Oswego sandstone. The Oswego sandstone is interpreted to represent the transition from marine to fluvial deposition, with source region to the south-southeast based on paleocurrent indicators (Patchen, 1966; 1978).

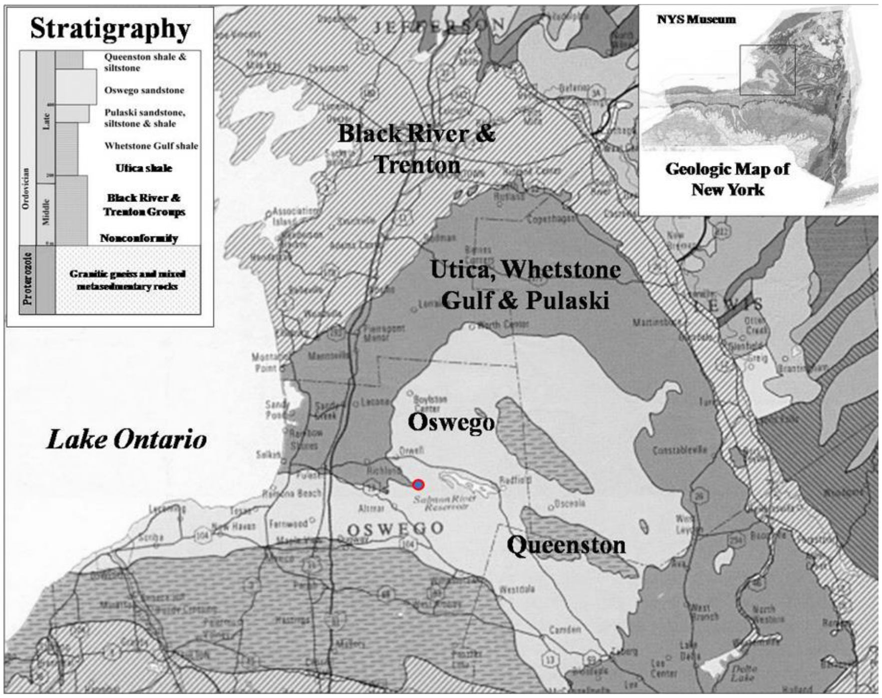

Figure 3. *Geologic map of the Tug Hill plateau (Isachsen and Fisher, 1970). The inset shows the general stratigraphy.*

Figure 4. *Photograph of the strata at the Salomon River falls. The gradational lithologic transition from the Pulaski formation (sandstone and shale) to lowermost Oswego formation (dominantly sandstone) is best exposed at this location. The stratigraphic column to the left represents the occurrences of sandstone (yellow) and non-sandstone (purple) strata (shale and siltstone beds).*

JOINTS, FRACTURE ZONES AND FAULTS

The orientation and density of various joint sets, fracture zones and faults in the strata of the Tug Hill – Eastern Lake Ontario region have been documented (Stilwell et al., 2005; Valentino et al., 2011). This field trip will focus on the variation in fracture systems within the Pulaski-Oswego formation transition that is fully exposed at the Salmon River falls, and this information was first presented in 2011 at NYSGA (Valentino et al., 2011).

Pulaski Shale-Siltstone-Sandstone.

Within the Pulaski Formation there are joints with strike variation on a bed-by-bed scale, where the shale, siltstone, and sandstone beds exhibit different fracture sets and densities (Figure 5A). Overall, there are northeast, east-northeast, southeast and south-southeast striking fractures that appear to be tied to specific lithologies. This is the only formation in the Tug Hill plateau that exhibits both joint sets that appear to be associated with the Appalachian basin (J1 & J2) (Engelder and Gross, 1993; Engelder and Lash, 2008) and with the crystalline basement (Valentino et al., 2016). In addition, the shale units in the Pulaski Formation have extensive tightly spaced fracture cleavage ranging from 2 to 5 fractures per centimeter (Figure 6A).

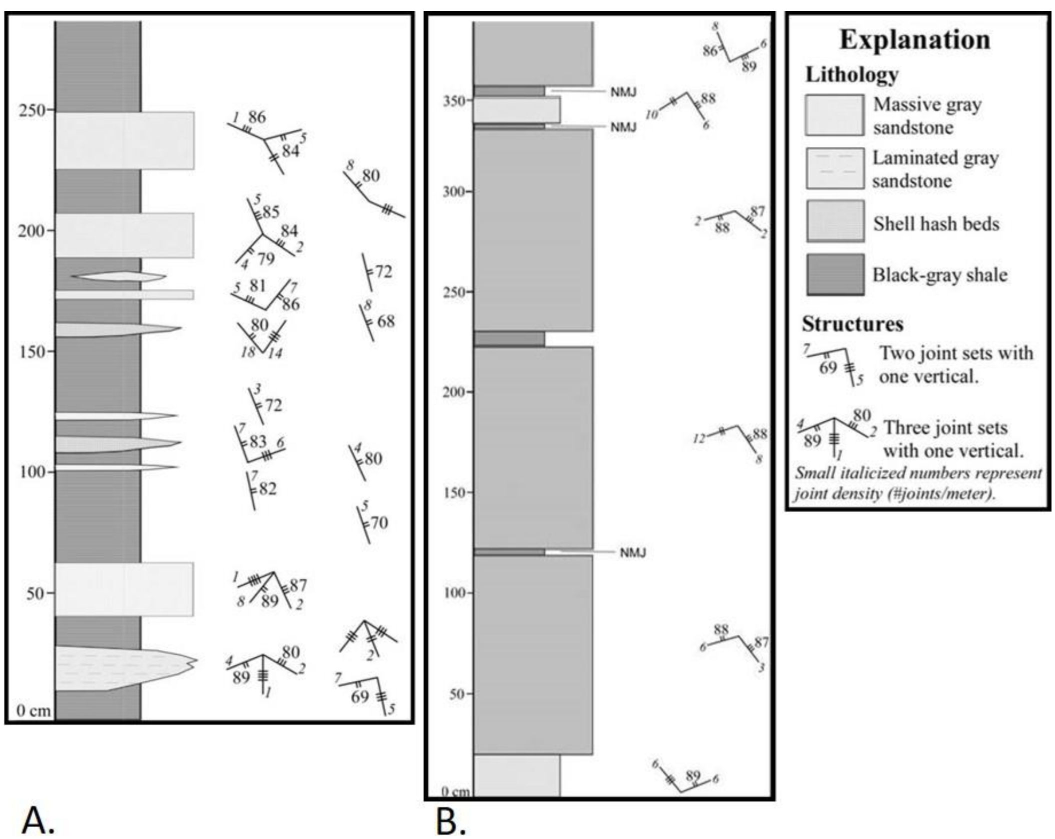

Figure 5. *Representative stratigraphic column for the Pulaski Formation (A) and Oswego Formation (B) at Salmon River Falls, Oswego County.*

Oswego Sandstone

Within the Oswego Formation, there are dominant east-northeast and southeast striking joint sets that persist from the region of Lake Ontario to the fringe of the Tug Hill plateau (Figures 5B). These joint sets are consistent throughout the sandstone and only vary in strike where beds are either very thick (> 40 cm) or very thin (<2 cm). There is evidence for minor left lateral shear (2 – 20 cm) on the east-northeast striking joint set, where the southeast striking joints have been displaced horizontally. Small en-echelon fracture zones with left lateral geometry have a consistent strike with the individual fractures (Stilwell et al., 2005; Valentino et al., 2011). At the Salomon River Falls, the large pavement outcrop reveals meter-scale right stepping fracture array associated with the east-northeast striking set (Figure 6).

REGIONAL FRACTURE SYSTEMS

Joint density (joints/meter) is the inverse of joint spacing. For example, a joint density of 5 joints per meter is equivalent to a rock with average joint spacing of 20 cm. Joint spacing was plotted using box-and-whisker (Mcgill et al., 1978) diagrams (Figure) for comparison within rock formations across the region, and for comparison between the different formations, including the Pulaski and Oswego Formations. Two integrated box-and-whisker diagrams were plotted to portray variation in the strike direction of the joint sets against the variation in the joint spacing. This type of plot allowed for visual and quantitative comparison within and across rock formations. Mean joint spacing of 10 cm is consistent throughout much of the Pulaski Formation,

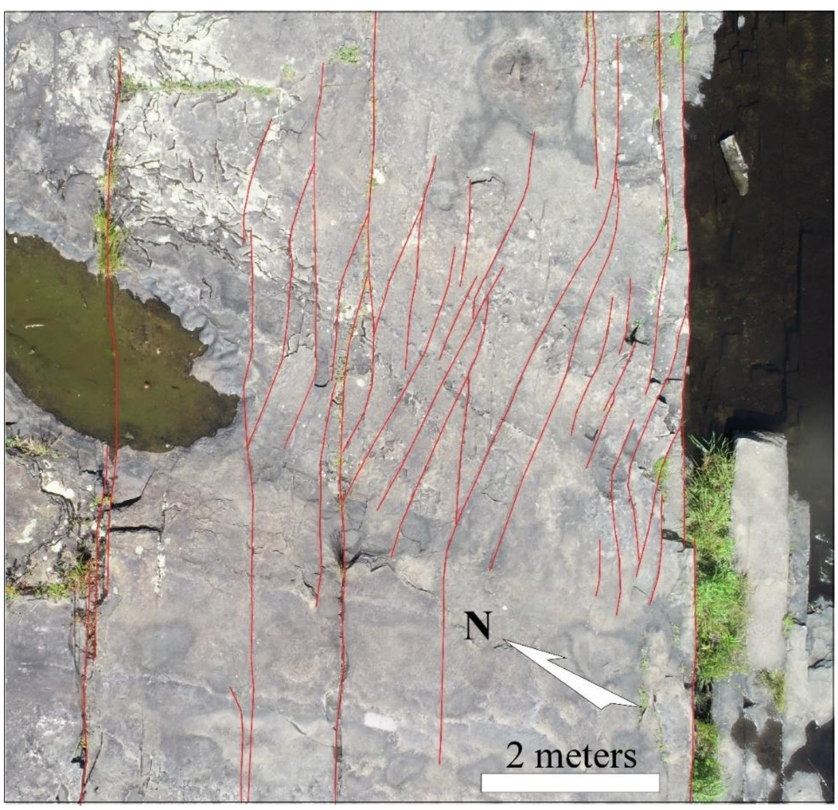

Figure 7 - Right stepping complex fracture array associated with the E-NE striking set. This is the same set of fractures that have sinistral en-echelon zones at the top of the Oswego sandstone.

where the southeast, east-northeast and northeast striking joint sets are dominant. Within the sandstone of the Oswego Formation, there are only southeast and east-northeast striking joint sets with mean spacing of 25 cm (Figure 7A). The Oswego Formation contains by far the most consistent joint-set occurrence in the entire plateau (Valentino et al., 2011), strike and spacing, probably reflecting the thick beds and relatively homogenous lithology of the formation.

From the field observations during this study, it is apparent that joint density and strike are variable at the outcrop scale. However, at the regional scale, it is also apparent that joint variability is related to the proximity to the crystalline basement both laterally and stratigraphically, and related to the variation in rock type with mechanical characteristic control on the fracture and fault formation (Figure 7). Starting at the base of the plateau, the limestone units have the most variability in joint strike with all sets that are present. The joint densities are relatively moderate with ~5 joints/meter in the rocks of the eastern flank of the plateau, and 2-3 joints/meter in the rocks of the western flank. In addition, the limestone units contain the highest number of southeast striking normal faults, and they are roughly parallel with the Black River fault (Figure 2) (Wallach and Rheault, 2010; Wallach, 2002). Within the shale-bearing rock formations, faults are not common, but the fracture density is high in all of the sets that are present, with density values that exceed 20 joints/meter. There are dominant northeast, east-northeast and southeast striking fracture sets with only a minor occurrence of the north-south striking Appalachian basin cross fold joints (Engelder and Lash, 2008). Finally, in the sandstone that caps the plateau, the fracture sets have a generally low density of <1-4 joints/meter (Figure 7A), and only the east-northeast and southeast striking fracture sets are present. The east-northeast striking joint set has evidence for left lateral shear (Stilwell et al., 2005; Valentino et al., 2011), which is evident in the offset of southeast striking fractures and primary en-echelon fracture zones.

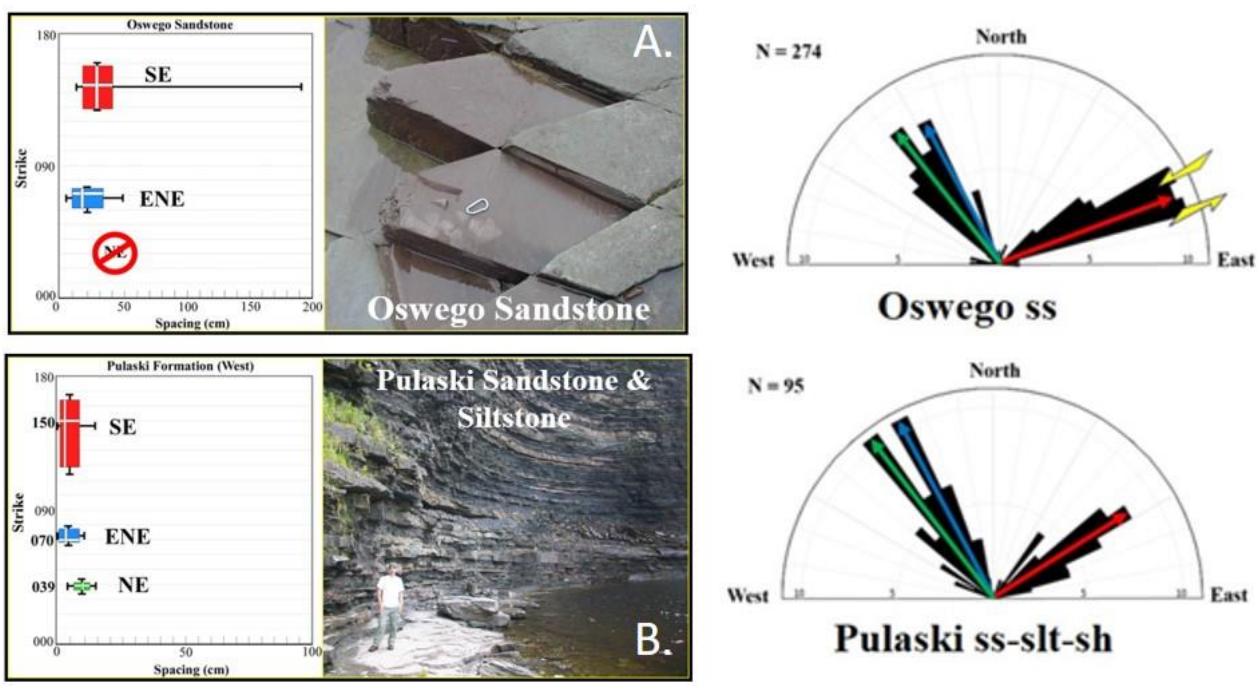

Figure 7. *Box and whisker plots and rose diagrams showing the orientations and spacing of fracture sets in the Oswego (A) and Pulaski (B) Formations at Salmon River falls.*

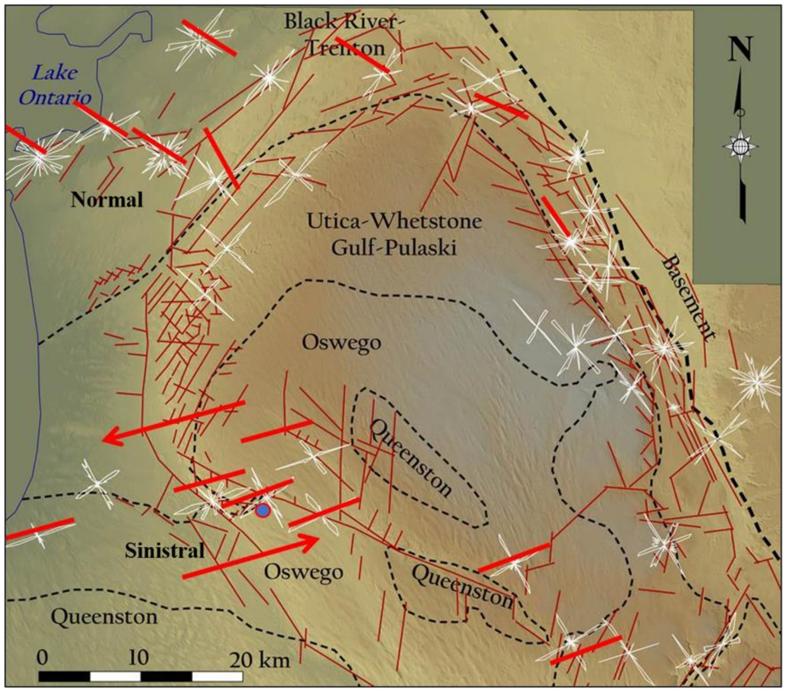

Figure 8. *Digital elevation model for the Tug Hill plateau with geologic formations, joint rose diagrams, and faults. The Black River fault is from Wallach and Rheault (2010). Minor faults are represented by short heavy lines represent observed faults. Those faults exhibit normal displacement in the northwest region and sinistral displacement in the Oswego Formation. The lighter weight red lines are interpreted topographic lineaments that are not related to glacial deposits.*

Four systematic joint sets were observed in the strata of the Tug Hill plateau. The east-northeast striking (J1 of Lash and Engelder, 2009) joints are well developed within all of the Ordovician rock units of the Tug Hill plateau, and there may also be some occurrences in the underlying basement. Note that these fractures are well developed in the underlying Trenton and Black River Group carbonates, suggesting that they are related to basement structure. Considering the persistence of these joints in all of the Ordovician strata, then perhaps they reflect the general east-west subhorizontal compression associated with the Alleghanian orogeny in the northern Appalachians, and the impact of that event on the underlying basement. On the contrary, the north-south striking cross-fold joints (J2 of Lash and Engelder, 2009) are only of minor occurrence, most likely due to the distal location relative to the Alleghanian thrust and fold belt.

The northeast and southeast striking fracture sets that dominate the crystalline basement (Figure 8) are also developed in the overlying Ordovician strata of the plateau, but the occurrence is variable. The northeast striking fractures have direct correlation with a suite of steeply dipping northeast striking normal faults in the southern Adirondacks (Roden-Tice, 2000; Valentino et al., 2011), however, no dip-slip deformation has been documented for the northeast striking joints in the Tug Hill plateau. Barosh (1990, 1992) proposed the northwest trace of oceanic fracture zones into eastern North America, showing several possible extensions into the Adirondack and Tug Hill regions. These proposed fracture zones are the likely cause of the southeast striking basement fracture set, in addition to the southeast striking faults that occur within the Trenton Group. In the statistical analysis of Adirondack faults, Deneshfar and Ben (2002) also concluded that the northwest striking faults to be candidates for seismic activity. Wallach and Reault (2010) demonstrated that southeast striking faults account for the southwestern regional tilt of the Ordovician strata and accommodated the vertical rise of the crystalline basement during formation of the Adirondack dome. The east-northeast striking sinistral faults in the Oswego Formation cross-cut the southeast striking fractures. This relative timing suggests that the sinistral deformation post-dates, or is synchronous with uplift of the Adirondack dome. This is most likely reactivation of the earlier developed J1 Appalachian joints (Lash and Engelder (2009) as the result of differential uplift of the Adirondacks. These faults were only observed in the Oswego Formation, suggesting that the underlying shale units accommodated strain through partial plastic deformation. Note the development of cleavage in the Pulaski Formation, as discussed above.

THE FALLS GEOMORPHOLOGY

Above the falls, the Salmon River flows generally toward the west-northwest (290^o). And continues on this path immediately downstream of the falls, until it enters two abrupt bends forming a "horseshoe shaped" gorge, then flows south-southeast parallel to the upper section (Figure 9). The river then takes another abrupt bend tow the southwest. Superficially, it appears that the river below the Salmon River falls is confined to a gorge that parallels the local fracture systems that occur in the Pulaski Formation. Despite this apparent correlation, the river flow across the Oswego Formation above the falls does not correlate with the fracture sets in the sandstone, but the face of the falls is clearly controlled by the northeast striking fractures (Figure 1). Although normal faults are not common in the Pulaski Formation, there is a minor fault that occurs in the strata on the southeast face of the falls and it traces upward into the Oswego sandstone where it can be observed at the lip of the falls. In the Pulaski Formation, this fault has minor normal offset (<1 m), but exhibits no displacement in the sandstone at the top of the falls. This fault is the location of the current face of the falls, and therefore, most likely contributed to the current orientation.

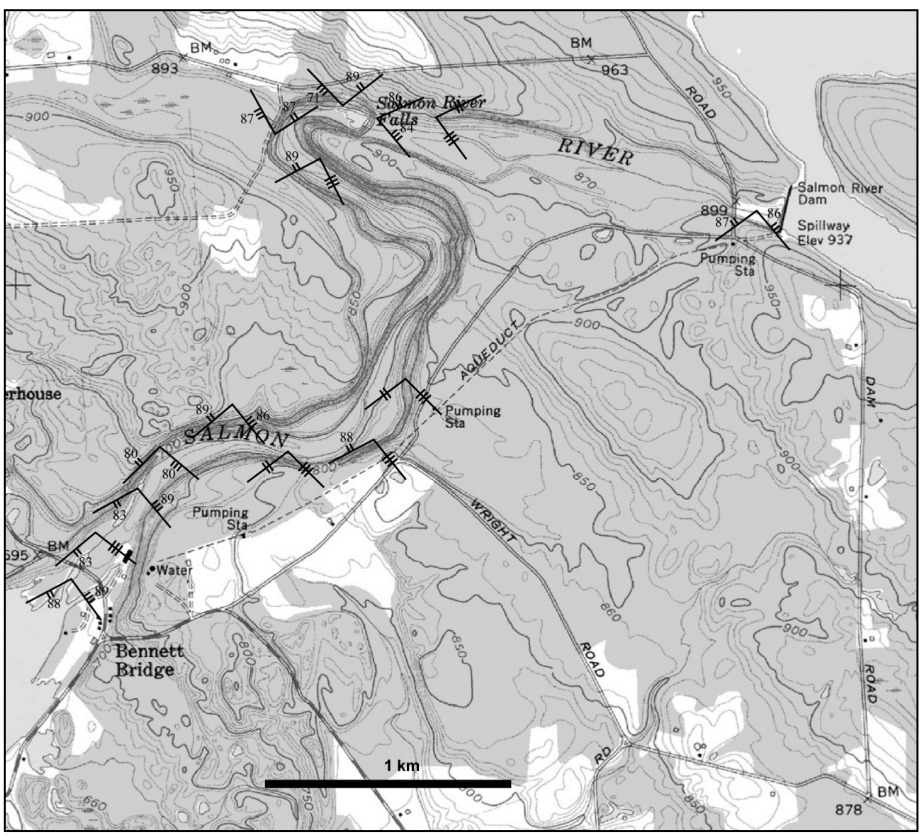

Figure 9. *Joint map for the Salmon River gorge, Oswego County, New York.*

ROAD LOG AND OUTCROP DESCRIPTIONS

Drive to the following coordinates: 43°32'56.47"N; 75°56'37.41"W

Park vehicle in the designated place. Follow the path northeast to the top of the Salmon River Falls.

This stop description was first prepared for the 2005 NYSGA conference and appears in Stilwell et al. (2005). Follow the steep path (with stone stairs) into the Salmon River gorge. The path splits in two directions at the river level. Follow the river bank to the right (southwest) about 200 m to the waterfall. Stop 5 is the series of outcrops that start at the waterfall and end at the sharp bend in the river downstream. At the river level, thick beds of sandstone comprise the lower part of the Pulaski Formation. Long joints (up to 30 m long), strike about 025 and are steeply dipping. These fractures are spaced about 1-2 meters and intersect discontinuous joints that strike northwest (Figure 9). The thick sandstone beds are overlain by interbedded sandstone, siltstone and shale that is progressively more dominated by fine grained sedimentary rocks upward. This sequence of rocks is best viewed in the cliff face that occurs in the sharp bend in the Salmon River. Walk back to the path and climb up the gorge. Follow the path eastward to the top of the Salmon River Falls. At this location, please obey all safety signs and stay in the designated areas. Stop 6 is the pavement outcrop at the top of the falls of Oswego Formation. Here the rocks are green-gray sandstone with abundant ripples and large cross-beds. Joints transect the entire outcrop for many meters and occur in complex zones. The dominant joint set strikes about 060, but is associated with some complex joint patterns. In places the individual joints with an average strike of 035 curve and are continuous with the 055 striking fractures. Additionally, there are en echelon fracture zones with apparent left lateral geometry that merge with the dominant fracture set, and large right-stepping fracture zone (Figure 7)

End of trip.

REFERENCES

Barosh, P.J., 1986. Neotectonic movement, earthquakes and stress state in the eastern United States, Tectonophysics, v. 132, p. 117-152.

Barosh, P. J., 1990. Neotectonic movement and earthquake assessment in the Eastern United States, Reviews in Engineering Geology, v. 8, p. 77-109.

Barosh, P.J., 1992. Northwest-trending basement fracture zones in the Eastern United States and their role in controlling neotectonic movement and earthquakes, Proceedings of the International Conference on Basement Tectonics, v. 7, p. 409.

Daneshfar, B. and Benn, K., 2002. Spatial relationships between natural seismicity and faults, southeastern Ontario and north-central New York state, Tectonophysics, v. 353, p. 31-44

Engelder T. and Gross R., 1993. Curving cross fractures and the lithospheric stress field in eastern North America. Geology, v. 21, p, 817-820.

Engelder, T. and Lash G., 2008. Fracturing within the outer arc of a fore bulge at the onset of the Alleghanian Orogeny. Journal of Structural Geology v. 29 p.774-786.

Isachsen, Y.W. and Fisher, D.W., 1970, Geologic map of New York: Mohawk sheet: New York State Museum, Map and Chart Series 15, scale 1:250000.

Lash G. and Engelder T., 2009. Tracking the burial and tectonic history of Devonian shale of the Appalachian Basin by analysis of fracture intersection style. GSA Bulletin; v. 121, p. 265-277.

Patchen, D. G., 1966. Petrology of the Oswego, Queenston, and Grimsby Formations, Oswego County, New York, M.S. Thesis, State University of New York at Binghamton, 191 p.

Patchen, D. G., 1978. Depositional environment of the Oswego sandstone, Oswego County, New York, in (ed. Merrian, D. F.) New York State Geological Association, 50th Annual Field Trip Guidebook, p. 368-385.

Roden-Tice, M. and Tice, S., 2009. Regional-scale mid-Jurassic to Late CretaceousUnroofing from th Adirondack Mountains through central New England based on apatite fission-track thermochronology, Journal of Geology, v. 113, p. 535-552.

Roden-Tice, M. K., Tice, S. J. and Shofield, I. S., 2000. Evidence for differential unroofing in the Adirondack Mountains, New York State, determined by apatite fissiontrack termochronology, Journal of Geology, v. 8., p. 155-169.

Stilwell, S., Valentino, J., Gawron, J. and Valentino, D., 2005. Late sinistral shear in the Ordovician rocks of Oswego County, New York: a look at faults and related fractures, in (ed. Valentino, D. W.) New York State Geological Association, 77th Annual Field Trip Guidebook, p. 29-44.

Joshua D. Valentino, David W. Valentino, Alex P. O'Hara and Hannah R. Valentino, 2011. Effects of Adirondack basement uplift on joints and fault development in the Appalachian basin margin, New York, New York Geological Association, 83rd Guidebook, p. 1-22.

Valentino, D., Valentino, J., Chairenzelli, J. and Inclima, R., 2016. Faults and fracture systems in t Adirondack Mountains, New York, in (Selleck, B. and Chiarenzelli, J. eds.) The Geology of the Adirondacks, Adirondack Journal of Environmental Science, v. 21, p. 101-117.

Wallach, J., 2002. The presence, characteristics and earthquake implications of the St. Lawrence fault zone within and near Lake Ontario, Tectonophysics, v. 353, p. 45-74.

Wallach, J. and Rheault, M., 2010. Uplift of the Tug Hill Plateau in northern New York State, Canadian Journal of Earth Science, v. 47, p. 1055-1077.