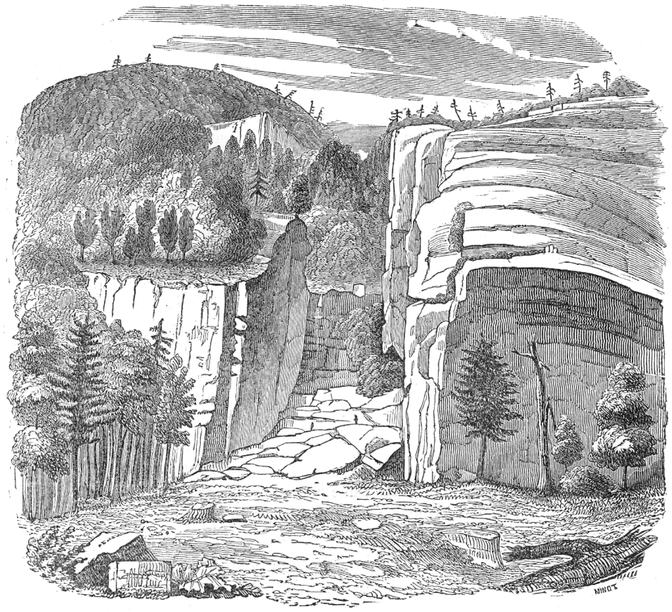
New York State Geological Association

Fieldtrip Guidebook



Great Trap Dyke at Avalanche Lake.

above: Sketch of the great trap dyke at Avalanche Lake. From Emmons, 1838

below: The 'Great Unconformity'. ca. 510 Ma Cambrian Potsdam Sandstone overlying ca. 1100 Ma Proterozoic gneiss, Putnam Center, NY

NYSGA

Colgate UNIVERSITY

80th Annual Meeting

Lake George, New York

September 26-28, 2008

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New York State Energy
Research and Development Authority

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A PRIMER ON NEW YORK'S GAS SHALES

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ABSTRACT

Though New York's first shale well was drilled in 1821, shale has not been a major contributor to natural gas production in the state. Recent price increases and the development of more efficient drilling and completion technology now make these rocks attractive for exploration. The resource in New York is significant: Previous estimates of the state's shale gas resource range from 163 to 313 trillion cubic feet (Tcf) out of just the Devonian section.

The New York State Energy Research and Development Authority (NYSERDA) has been investigating New York’s shale resource for more than two decades. Early work completed in the 1980s targeted the Devonian shales, including the Marcellus. Starting in the mid-1990s, the NYSERDA began to look at the possibility that all the state’s shale formations may offer production potential. Recent NYSERDA sponsored projects are helping to characterize both the Devonian and Ordovician shales. Prospective shales include the Ordovician Utica, the Middle Devonian Hamilton (Marcellus), and suite of Late Devonian shales that are separated by silt/sandy layers.

Experience developing shale gas plays in the past 30 years has demonstrated that every shale play is unique. Each individual play has been defined, tested and expanded based on understanding the geology, resource distribution, natural fracture patterns, and limitations of the reservoir, and each play has required solutions to problems and issues required for commercial production.

ANATOMY OF PRODUCTIVE GAS SHALES

Though shale production dates all the way back to 1821, the conditions for wide-spread field development have only become apparent in the last decade. The key driver is the application of technology and reservoir management practices that increase production levels considerably over those seen in the Eastern Gas Shales program. Tightening North American supply also make “unconventional resources” more attractive by creating a bullish pricing environment. Finally, the Antrim and Barnett shales, which serve as endpoints in the shale spectrum between adsorbed gas production and fracture gas production, prove that multiple exploration and extraction models exist for shale (Drake 2007).

Gas shales are often the origin of hydrocarbon stored in conventional reservoirs. These hydrocarbons have been expelled, migrating upward into a trap of reservoir quality rock below a sealing unit (often shale). In gas shale systems, the shale is all three: hydrocarbons are generated, stored and held in place. The preserved organic matter is "consumed" through biogenic or thermogenic processes to generate smaller chain hydrocarbons (gas or liquid). The remaining carbon that cannot be converted (dead carbon) and clay minerals form a storage mechanism through adsorption, which increases tremendously the potential storage volume. The relationship between temperature, pressure, available volume and the general attractiveness of methane (partial pressure) will define ultimate adsorbed storage capacity. Even after a great amount of generated gas is expelled out of the shale (as source rock), there can remain an enormous quantity as adsorbed gas. Gas will also reside in rock matrix pore space and fractures if there is a "seal" to keep the gas in these spaces (Figure 1).

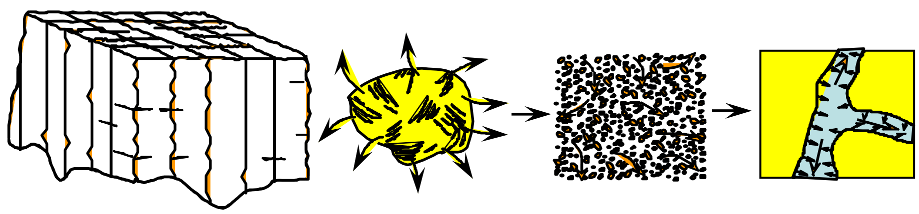


Figure 1. Gas Production Methods (after McDonald, 2002)

Productive oil and gas shale range in age from the Cambrian to the Jurrassic. Age is less relevant than the volume of rock available. Productive shales are usually vertically substantial (30 meters or more) and geographically prevalent (hundreds of square kilometers). Most are marine but there is at least one productive lacustrine (Green River Shale) formation (Chornoboy, 2007). Key to producibility is that the shale must be deposited in an anoxic environment to preserve enough organic material for gas generation.

Hydrocarbon generation in organic black shales take on varying physical and chemical characteristics. One constant is the presence of organic material that provides for a source of liquid and gaseous hydrocarbons and a potential storage site. The amount of gas present in organic rich shales (at a given locality) is dependent on three factors: 1) the amount of organic matter originally deposited with the rock, 2) the relative origins of the different types of organic matter and the original capacity of each for gas generation, and 3) the degree of conversion of the organic matter to hydrocarbon natural gas. The first two factors are largely dependent on conditions present at the site of deposition, and the third is determined by intensity and duration of post-depositional heating, or load metamorphisim due to maximum depth of burial. This also assumes that the natural gas has remained, to some extent, trapped in the source to become a "reservoir."

The amount of organic carbon present in the rock is not only important as a source rock, but it also contributes to the natural gas storage by adsorption and or solution within the reservoir system. In the Appalachian Basin, darker zones within the Devonian Shale (higher organic content) are usually more productive that the organic-poor gray zones (Schmoker 1980).

Knowing the type of kerogen that is present in the rock provides information on hydrocarbon source potential and depositional environment. Kerogen type can also influence the amount of natural gases stored by adsorption as well as diffusion rate. The classification scheme for kerogen evolved initially from the optical maceral analysis of coal. Elemental analysis was later applied to kerogen analysis. The elemental analysis is based on the quantification of the hydrogen/carbon (H/C) and oxygen/carbon (O/C) ratios from Van Krevelen (1961). A plot of the ratios, called the Van Krevelen diagram, was developed to diagrammatically determine kerogen types and thermal maturation. The ratios on the Van Krevelen diagram were replaced with the indices (HI and OI) from Rock-Eval data resulting in a modified Van Krevelen diagram (Espitaliè 1977). This modified diagram is used to determine kerogen types. Figure 2 is a further breakdown and description of the four common types of kerogen.

Kerogen Type	Depositional Environment	Organic Precursors	Hydrogen Product
I	Lacustrine	Algae	Liquids
II	Marine, Reducing Conditions	Marine Algae, Pollen, Spores, Leaf Waxes, Fossil Resins	Liquids
III	Marine, Oxidizing Conditions	Terrestrial-Derived Woody Materials	Gas
IV	Marine, Oxidizing Conditions	Reworked Organic Debris, Highly Oxidized Material	None

Figure 2. Kerogen Types (Waples 1985)

The maturation level of the kerogen is used as a predictor of the hydrocarbon potential of the source rock. It also is used to high-grade areas for fractured gas shale reservoir potential and as an indicator for investigation biogenic gas within a shale reservoir system. Thermal maturation of the kerogen has been found to also influence the amount of natural gas that can be adsorbed onto the organic matter in shale. Thermal maturation can be determined by several techniques, including Rock-Eval, vitrinite reflectance, thermal altercation index and conodont alteration index. Multiple techniques should be employed to help determine thermal maturity of a shale.

Reflectance of coal macerals in reflected light has long been used to evaluate coal ranks. Reflectance measurements have been extended to particles of disseminated organic matter occurring in shales and other rocks (kerogen) and have been the most widely used technique for determining maturity of source rock. Typical analysis normally shows a distribution of reflectance corresponding to the various constituents or macerals of the kerogen. Because humic or vitrinite particles are generally used for reference to the coalification scale, the mean random reflectance of vitrinite (Ro) is preferred to other particles. In some cases, there may be several groups of vitrinite particles with different reflectance present. In these situations, it is recommended that only the group with the lowest reflectance should be used. Other groups with higher reflectance are considered “reworked.” Figure 3 is a breakdown of the different stages of maturation with vitrinite reflectance. Vitrinite reflectance and organic content can be used to develop an adsorption isotherm that displays the ability of the shale to chemically adsorb methane at different pressure and temperature conditions (TerraTek 2004).

Other optical measures of thermal maturity include conodont alteration index and the thermal alteration index (TAI). The TAI uses of progressive changes of color and/or structure of pores, pollen or plant-cuticle fragments is also used as an indicator of thermal maturation of the kerogen. Kerogen coloration is reported on a scale of 1 to 5, and is referred to as Thermal Alteration Index (TAI) (Staplin 1969). The thermal maturity of shales can also be inferred from published conodont alteration indices (CAI), a scale of color alteration in conodonts (a marine fossil) (Epstein 1977). In general, the CAI of a conodont increases with depth and temperature as a result of metamorphisim.

Vitrinite Reflectance	Comments
R_o <0.5 to 0.7%	Diagenesis stage, source rock is immature
0.5 to 0.7% >R_o<1.3 %	Catagenesis stage, main zone of oil generation
R_o >2.0%	Metagenesis stage, methane remains as the only hydrocarbon (dry gas zone)
R_o is the mean reflectance in oil.	

Figure 3. Vitrinite Reflectance Categories for Thermal Maturity

The degree of kerogen conversion to liquid or gas hydrocarbons is measured by a series of indices, many created through RockEval. Rock-Eval can be used to assist in determining the thermal maturation level of kerogen. Peters (1986) defined the thermal parameters in which T_{max} (maximum temperature) can be used to define the dimensions of the oil window (Peters 1986). The top of the oil window is generally assumed to occur between T_{max} values of 435ºC and 445ºC and the bottom of the oil window occurs at 470ºC. Plotting T_{max} and hydrogen index can proxy the thermal maturation and kerogen type of the samples. Estimates of hydrogen index, transformation ratio and production index are also be used to distinguish remaining hydrocarbon potential from generated hydrocarbons. These are particularly useful in situations of high maturity since Tmax becomes unreliable. Productive shales may be in the generating window (like the Antrim and Woodford) but can also be beyond generation (Fayettville and some Barnett fields).

The amount of gas in place in shale is dependent on the presence of organics and clays as well as the ability for methane to adsorb onto the solid lattice internal surface. Organic content and quality give an idea of the storage capacity while rock characteristics give an idea of the ability to deliver the gas from the rock to the borehole. Mineralogy and rock fabric help define the ability of the rock to move gas out of the storage and matrix and into the larger fracture network.

Open, orthogonal or multiple sets of natural fractures increase the productivity of gas shale reservoirs due to the extremely low matrix permeability of shales (Hill 2000). Fractures must be present, whether natural or induced through hydraulic fracturing. Finding these natural fracture systems are critical to commercial production of natural gas and is considered one of the primary exploration strategies. Identification and characterization of natural fractures is typically done either at the surface through outcrop studies or in-situ through the use of geophysical logs or core. In addition, indications of natural fractures are often associated with natural gas shows while drilling a well, especially on air or under-balanced.

SHALE GAS IN NEW YORK

The Appalachian Basin in the northeastern United States is an important hydrocarbon province that has been producing oil and gas since the early 1800’s. More than 40 trillion cubic feet (Tcf) of natural gas and millions of barrels of oil have been produced from reservoir rocks of all ages. Devonian-age shales are a significant resource in the basin. Their coal-like appearance, wide spread distribution, and stratigraphic nearness to the surface led to interest and use as an energy source dating back to the 1700’s. The Devonian Shale of the basin has been estimated to contain up to 900 Tcf of natural gas, and an estimated 120,000 wells have produced roughly 3.0 trillion cubic feet (Tcf) of natural gas in the past 30 years (Milici, 1996). In addition to Devonian Shale, other stratigraphically older and deeper black shales are present in the basin, and the organic-rich Ordovician shales are believed to be a principle source rock for many of the productive reservoirs in the basin. These shales, though not frequently produced, are often noted in driller’s logs to have significant gas shows when drilling through them. As of 2007, exploratory drilling was underway to begin producing the Utica Shale of eastern and central New York State.

Curiosity about the black shales of New York from a geologic perspective and as a fuel source dates back to the late 1700's. The black coal-like appearance and slightly combustible nature of the shales were of interest to the coal industry, and gas seeps in creek beds motivated early explorationists to study the rocks and find use for them. The first know commercial shale gas well was drilled in 1821 in the town of Fredonia, Chatauqua County, New York near a gas seep along Canadaway Creek (de Witt 1997). The well, drilled by William Aaron Hart, was completed as a gas producer in the shallow Dunkirk shale. The well was connected to pipeline and provided natural gas to Fredonia's main street businesses and street lamps in the 1820's. Following Hart's success, the development and use of shale gas proliferated along the south shore of Lake Erie, eventually spreading southward into Pennsylvania, Ohio, Indiana, and Kentucky. By the turn of the century hundreds if not thousands of wells had been drilled along the lake shore and in the basin, and were producing shale gas for domestic and small commercial use. However as exploration advanced, the development of shale gas wells diminished in favor of more productive conventional oil and gas horizons. It was observed early on that shale gas was tight, and while successful wells produced steadily over long periods of time, production volumes were extremely variable and unpredictable, but usually low (<100 mcfd). The mechanisms controlling production from these wells were not understood, and the technology to optimize production was in its infancy.

In the late 1960's, as natural gas reserves in the United States began to diminish, the U.S. Energy Research and Development Administration (ERDA, later the U.S. DOE) initiated a program to evaluate the Nation's gas resource. Recognizing that the Devonian and Mississippian black shales were a major gas resource that required advanced production methods for recovery, the ERDA launched the *Eastern Gas Shales Project* (EGSP) in 1976. The project was a joint research project between the DOE and numerous State, Federal, and private industrial organizations, which were brought together to participate in the research. NYSERDA entered the project in 1979 by initiating a 4 well R&D program. NYSERDA has continued research and testing to help define the gas shale potential in New York State.

GEOLOGY

New York forms the northern edge of the Appalachian Basin that exists from southern Ontario to Tennessee. With few exceptions, the state's bedrock primarily consists of Devonian-age and older formations. The younger rocks lie to the south and all sedimentary formations outcrop to the north, at the edge of the Adirondack uplift. The Ordovician and Cambrian become visible again in the St. Lawrence Lowlands.

The rocks in New York have been impacted by at least one of the three major Paleozoic tectonic events. This has left the subsurface folded, fractured, and compressed. Also, numerous sea level changes created significant unconformities including the Knox Unconformity. Studies indicate that the Devonian age and older rocks underwent deep burial before being uplifted to their current elevation. This tectonic history created the environment for hydrocarbon development and the trapping mechanisms to accumulate economic quantities of oil and natural gas. Lake Ontario and the Adirondack Mountains form the northern boundary, the eastern margin is formed by the Hudson Lowlands and Taconic Mountains, and to the west terminates at the shore of Lake Erie. Paleozoic rocks overlying the Precambrian crystalline basement outcrop along the northern extent of the Allegany Plateau, and dip gently to the southwest. In the southern portion of New York, a series of small-scale folds are present, extending from Chatauqua to Tioga counties. The folds are small anticlines, dipping less than 2^{o}, which are associated with the Appalachian Fold Belt, an arcuate belt of anticlines and synclines that extend southward into West Virginia (Frey 1973).

In the last 1 million years, New York has endured significant continental glaciation, with ice thicknesses approaching one mile. According to Robert Milici, "glacial loading and post-glacial isostatic rebound in the gas-producing regions to the south of the Great Lakes appears to have created the fractured pathways for gas to have migrated from black shale source rocks into intercalated brittle silty and sandy reservoirs, as well as to have fractured and enhanced the storage capacity of these reservoirs (Milici 1996). The ice at its maximum extent is estimated to have been over 1 mile thick, and the shear weight of the ice sheet caused the region to compress and sag (Isachsen 2000). When the ice melted, ocean water temporarily flooded low-lying areas in the Champlain and St. Lawrence valleys that had been depressed forming the Champlain sea. Many marine deposits of this sea are now found at elevations exceeding 300 feet, indicating rebound of the region occurred. In the south where the glacial ice was thinner the rebound was less, however in the north where the ice was thicker, the rebound is over 400 feet. The uneven rebound is seen throughout northern New York. Glacial lake deposits that were once horizontal are now inclined to the north, and in the Lake Ontario region, the whole are has been tilted north to south. Post glacial rebound is now complete in New York, however the near-surface joint system has been enhanced and opened by the release of the glacial weight (Charpentier 1982). The presence of horizontal fractures in the Devonian is mentioned in well records, and has been attributed to glacial unloading (Imbrogno 2003). Gas-charge horizontal bedding plane fractures also existing in the shale sections of the Upper Trenton Formation and may owe their existence to glacial loading and unloading as well (Kerr 2006; Smith 2008).

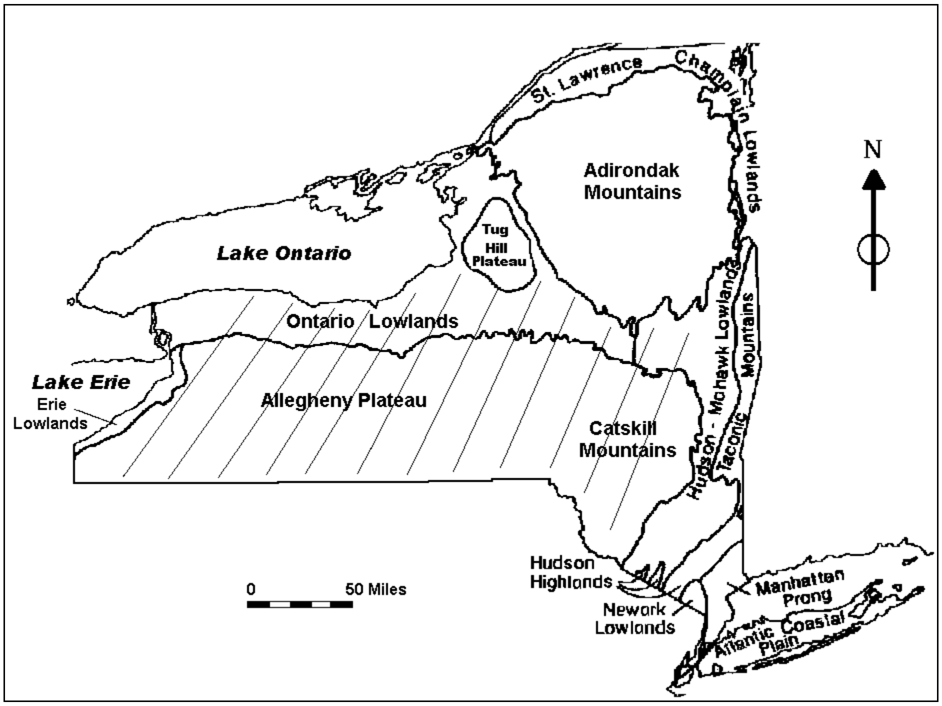


Figure 4. Physiographic Regions of New York

Both gas and oil have been produced from rocks of many ages in New York, and the primary targets for operators in the past have been the gas-bearing sands in the Oriskany, Medina, Queenston, Chemung and Fulmer Valley formations. The organic rich black shales are the principal source rock for much of the oil and gas in the basin (Milici 1992). In addition, gas shows have been noted frequently in drillers logs and petroleum related hydrocarbons have been observed in cuttings from the Ordovician-age Utica Shale (Robinson 1989).

Stratigraphy

Organic-rich black shale beds are found in many different age rock formations in New York. Some are massive and very widespread correlating well to the shales in other regions of the Appalachian Basin, while others are thin and limited in area. The following section provides an overview of the stratigraphy of the primary black shale intervals in the Paleozoic section of New York. A large volume of literature exists that thoroughly discuss the many stratigraphic units and variances in New York.

Ordovician. --- One of the oldest and most widespread black shales is the Ordovician-age Utica Shale. It was deposited very broadly across the Appalachian Basin and into Ontario, and covers thousands of square miles. In New York the Utica is found in outcrop along the west and south-southeast sides of the Adirondack Mountains, and is well exposed in several locals along the northern margin of the Allegany Plateau. It is deeply buried over most of the state of New York, and from outcrop it dips to depths over 9,000 feet in the southern portion of the state (Wallace 1988). Oil and gas shows have been reported in the black shale of the Utica and in its Dolgeville member, including a recent report of 1 MMscf/day (Trevail 2003). In 2007, Two test wells were drilled in Otsego County but results remain inconclusive.

The Utica is a massive, fossiliferous, organic-rich, thermally-mature black to gray-black shale deposited in a subsiding trough that generally trended north-south. Source rock for the organic-rich black shale was supplied from the eroding Taconic highlands to the east. As the deep marine trough was filled in, the deposition of the lower members of the group onlapped westward over the carbonate platform. The westward migration was periodic which is reflected in the presence of multiple facies intervals, which are bounded by unconformities or condensed beds (Lehmann 1995). Each unit represents a pulse of subsidence and subsequent sedimentation in the basin, and all have several similarities. Each interval onlaps argillaceous limestone, condensed interval, and each appears to record a localized deepening event. The overlying and has shifted westward with respect to the underlying unit. The base of each unit is defined as a disconformity and/or stratigraphically black shale unit is thinner than the previous unit (Lehmann 1995).

The Flat Creek Member is the basal unit of the Utica and sits unconformably on units of the Trenton carbonates (Joy 2000). It is a transgressive deep basinal calcareous shale that represents the first mud flux from the erosion of the Taconian Island Arc to the east (current orientation). The Flat Creek is usually dark gray to black, variably calcareous shale with minor thin beds of argillaceous micrite and biomicrite (Lehmann, 1995) as shown in Figure 5). Vertical calcite filled veins cut the Flat Creek Member as shown in Chuctanunda Creek, Florida, NY (Fig. 5d).

Figure 5. Clockwise from left: a-c) Flat Creek Member along Route 5S, Florida, Montgomery County, d) Flat Creek Member along the Chuctanunda Creek, Florida, Montgomery County (Martin, 2008, Nyahay, 2008)

The depocenter for the Flat Creek may very well have been east of the Hudson River, New York State. The unit is extremely thick in the Hudson Valley but thins and subcrops to as one moves to central New York. Nyahay identifies the pinchout as the point where the Steuben Limestone appears and thickens to the west (Nyahay, 2008).

The middle member of the Utica is the Dolgeville, a “pulse” carbonate/shale with turbiditic attributes (Mehrtens 1988). Shale members resemble Flat Creek; limestone members resemble Denley Limestone (Trenton Group). The Dolgeville, interpreted as a slope facies peripheral to the Trenton platform, interfingers with the basal Flat Creek black shale member (Nyahay, 2008). Microfolding created significant fracturing for gas migration from the organic-rich shale interbeds (Figure 6).

Figure 6. Dolgeville Member from left to right: a) microfolds near Paradise Road, Herkimer County, b) top of the Dolgeville along the NYS Thruway near Little Falls (Martin 2008, Nyahay 2008).

The uppermost member of the Utica is the Indian Castle, a transgressive, fissile shale with some calcareous interbeds. The upper units are more monotonous and fissile while the lower units are more blocky with impure limestone beds (Figure 7).

Figure 7. Indian Castle Member from left to right: a) Upper Indian Castle at Little Falls exit of the NYS Thruway, b) Lower Indian Castle at Paradise Road. Herkimer County (Martin 2008, Nyahay 2008).

As a result of basin development, the thickest section of the Utica is found along the Mohawk Valley and was deposited in the subsiding trough where it is well over 2,000 feet thick. It thins to the north and west to less than 100 feet along the Lake Erie shoreline where it becomes somewhat silty. Over much of western New York State, the Utica is less than 300 feet thick (Figure 8) (Zerrahn 1978). The Utica is overlain by coarser clastics of the regressive Lorraine Shale that consists of shale, siltsone and fine-grained sandstones. The Lorraine was deposited as the marine environment prograded westward and deltaic deposits pushed across New York from the east.

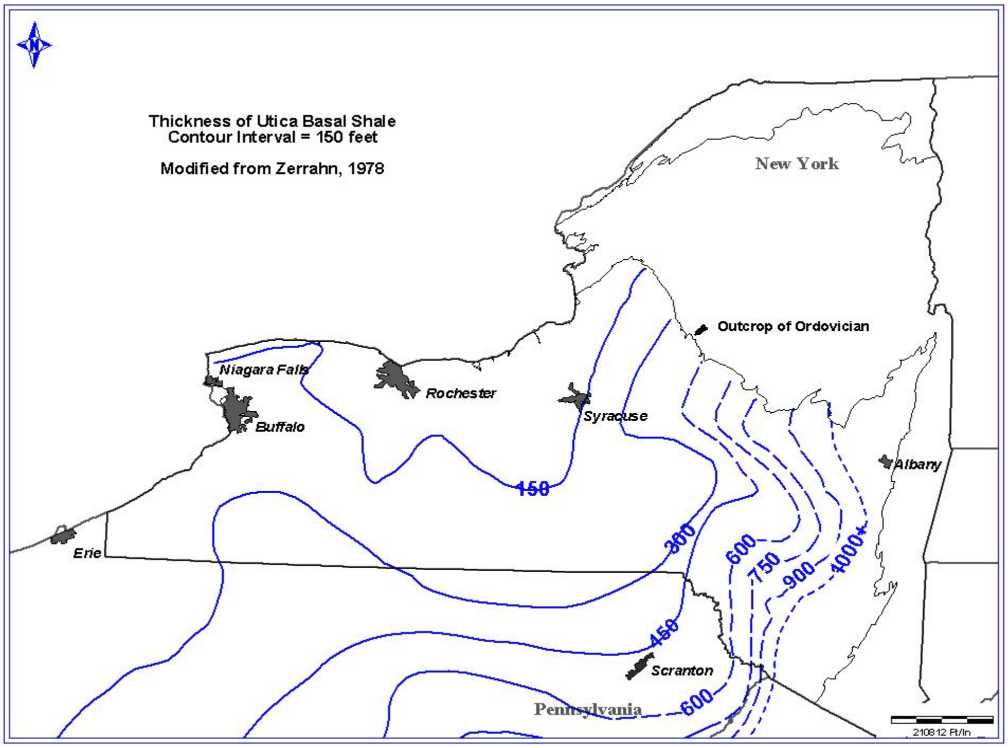


Figure 8. Isopach of the Utica Shale.

Silurian. --- The Silurian rocks of New York were deposited in the northern end of the Appalachian foreland basin during a relatively quiet tectonic time. They represent a short interval of geologic time, roughly 20 million years, however reflect a wide variety of depositional environments. Many of the Silurian rocks are extremely fossiliferous, indicating deposition in relatively shallow warm water. Silurian rocks in New York consist primarily of dolostone, limestone, evaporites, medium-gray and greenish-gray shales, and thin but persistent beds of phosphatic nodules and oolitic or fossil-rich hematite. No information regarding the organic content and thermal maturity of Silurian shale has been found. As they are primarily gray shales (there is one black shale member) they are not organically rich in general; however two shales in the Clinton Group are of interest because of their close proximity to the gas-charged productive horizons, and because two wells are reported to produce natural gas from Clinton Groups shales. However it may be that the gas in the producing rocks migrated there from other source rocks (Martin 2003).

The Sodus Shale was deposited near shore in shallow warm water, and contains a readily identifiable "pearly shell" limestone layer, which formed as a result of a very dense population of small shellfish. The shale is greenish-gray to purplish and was probably deposited in shallow, stagnant, low energy water. One well is reported to produce from the Sodus Shale in Seneca County. Overlying the Sodus is the Williamson Shale, a black shale which was deposited in deep, almost lifeless, anoxic water which was created by the presence of a great deal of iron in the sediments. In a drastic change of environment, the Williamson is overlain by a fossil rich limestone bed and the Rochester Shale. The Rochester Shale is brownish-gray, calcareous, and fossiliferous with interbedded argillaceous limestone layers, and is well exposed in numerous road cuts and creeks (Figure 9). One well is also reported to produce from the Rochester Shale in Seneca County.

Overlying the rocks of the Clinton Group is a continuing sequence of near-shore/marine rocks of the Lockport Group. The alternating layers of sand, shale, limestone are rich in fossils. The overlying Salina Group was deposited near-shore, and contains shales, dolostone, and numerous evaporite beds. The salt beds of the Salina had a great influence on the structural deformation of overlying rocks in the basin. The salt layer divides the rocks of the Allegheny Plateau horizontally, separating the youngest Silurian and Devonian rocks above from lower Paleozoic rocks below (Isachsen 2000). The salts, which are extremely malleable, provided a zone of weakness that allowed the younger rocks above to slide to the northwest during regional compression without significant folding and faulting. The resulting horizontal fault, or *décollement* separates the fixed rocks below form the transported rocks above.

Figure 9. Rochester Shale Outcrop in New York (University of Rochester 2002)

Devonian. --- The Devonian section covers approximately 22,500 square miles in south-central New York (Figure 10), and represents some 50 million years of history. It crops out along the northern and eastern margin of the Allegheny Plateau and is roughly 3,000 feet thick near Lake Erie, where it is composed primarily of rocks with marine origins. To the southeast, it thickens to over 9,000 feet, and is composed primarily of rocks of continental origin (Isachsen 2000). Depth to the base of the group increases from outcrop to over 4,000 feet in southern New York (Figure 11). The black shales in the Devonian section generally are thickest in the western and central portion of the Allegheny Plateau. To the east, they thin and pinch out, grading into coarser gray shales and siltstone. Interbedded are several thin, but widespread limestone units, which serve as marker beds used to differentiate between the numerous formations.

The gas-bearing shale portion of the Devonian in New York occurs in the Middle and Upper Devonian, and extends from the top of the Onondaga Limestone through the Perrysburg Formation (Van Tyne 1978). They are in ascending order: the Hamilton Group, Genesee Formation, Sonyea Group, West Falls Formation, and Canadaway Group. The rocks of the Hamilton Group are the oldest strata of the Devonian gas shale sequence. The group overlies the Onondaga Limestone, and consists of black and dark gray shales in the lower part, and limestone, light gray shale and mudstone in the upper part. The Hamilton Group outcrops along the northern margin of the Allegheny Plateau, and thickens eastward from 250 feet near Lake Erie to over 2,500 feet in Ulster and Green counties. The Hamilton has been subdivided into four units: the Marcellus, Skaneateles, Ludlowville, and Moscow, which are separated by thin limestone beds. The basal unit of the Hamilton is the Marcellus Shale. The Marcellus formation is highly radioactive and regionally extensive, covering most of the Allegheny Plateau and extending southward through the Appalachian Basin. It is a “sooty” black/brown to dark gray fissile shale with interbedded layers of medium-gray shale and limestone nodules or beds of dark gray to black limestone. It ranges from 25 feet to over 100 feet in thickness.

The Stafford Limestone overlies the Marcellus and marks the base of the Skaneateles Formation, which is a dark to medium gray fossiliferous shale and mudrock, containing a thin, black shale, the Levanna Shale. The Skaneateles is more clastic in nature than the Marcellus and contains some sandy layers. It is overlain by the Centerfield Limestone, which marks the base of the Ludlowville Shale. The Ludlowville is a dark gray basal shale, overlain by a lighter shale.

The Geneseo Shale is the basal unit of the Genesee Formation, and is the primary black shale in the formation. It is a fissile, organic-rich shale which when broken emits a distinct petroleum odor (de Witt 1993). The Geneseo attains a maximum thickness of 125 feet in central Steuben County. The Lodi Limestone overlies the Geneseo and consists of large discoidal limestone nodules in a bed of dark-gray fossiliferous siltstone. The overlying Penn Yan and West River shales are dark gray to medium gray organic-rich shale and mudstone, with some beds of black shale that extend into the Renwick. A thin limestone, the Genundewa, is found between the two shales in central New York, but pinches out southward and the shales grade into each other.

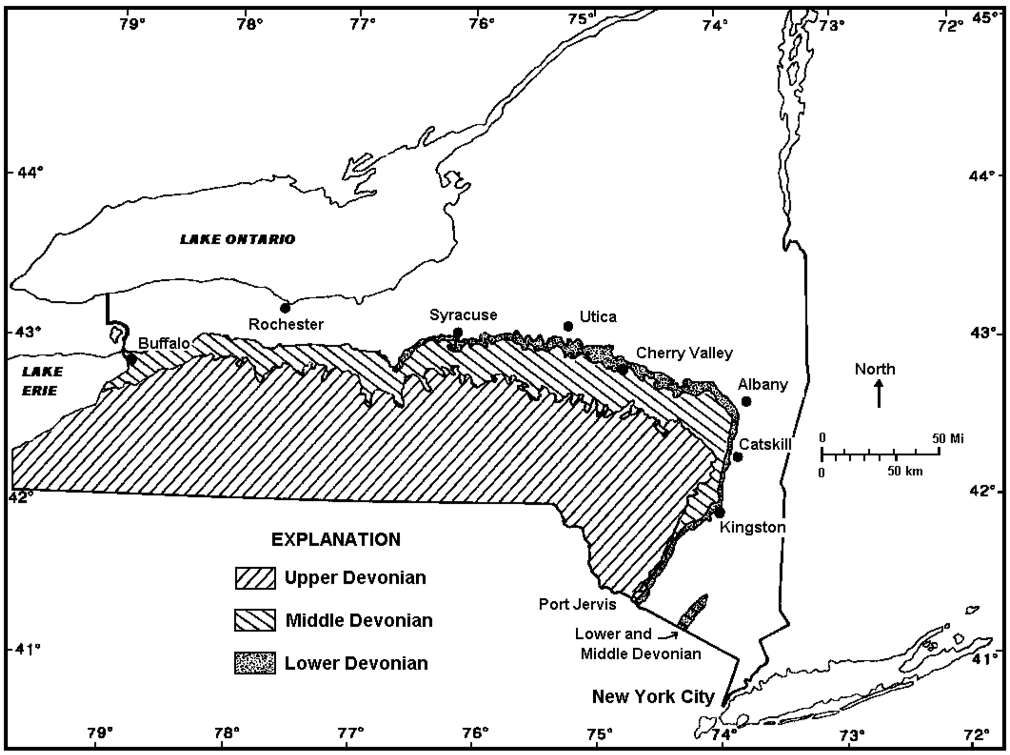


Figure 10. Devonian Outcrop in New York (modified from Isachsen et al 2000)

The Sonyea Group overlies the Genesee Formation and is subdivided into the Middlesex Shale and the Cashaqua Shale. Thickness of the Sonyea increases from approximately 10 feet at Lake Erie to over 800 feet in Tioga County. Like the Geneseo the Middlesex is a black organic-rich shale in western New York. Interbedded are layers of dark gray and brownish-black shales. It covers much of southern New York, and averages 65-75 feet thick in Yates and Steuben counties, and thins to the west to less than 10 feet. The Cashaqua is a gray shale with an abundance of flat ellipsoidal limestone nodules, and a few thin layers of black shale. The two shale members grade eastward into a thickening sequence of siltstone and silty shale, which is part of a common turbidite facies of the Catskill Delta.

The West Falls formation overlies the Sonyea Group, and consists of two shale-bearing formations, the West Falls Formation and the Java Formation. The Rhinestreet is the basal shale unit of the West Falls Formation. It is a thick, fissile, black shale outcropping in Chatauqua County where it is about 140 feet thick. To the east it thickens rapidly as it grades into and interfingers with the overlying gray Angola shale reaching a thickness of over 1,200 feet at the Allegany-Steuben County line, however the black shale component of the Rhinestreet thins eastward to less than 5 feet in Allegany County. The Overlying Java Formation ranges from 100 feet in thickness in western New York to over 600 feet in Steuben County. At the base of the Java is the thin, black Pipe Creek Shale. It is persistent, organic-rich black shale throughout its lateral extent. It is thin, not more than 25 feet at its maximum in south-central Cattaraugus County, and pinching out in northern Steuben County. The Hanover Shale is a gray shale, with some interbedded black shale beds. It thickens to the east grading into silty shale, siltstone, and sandstone.

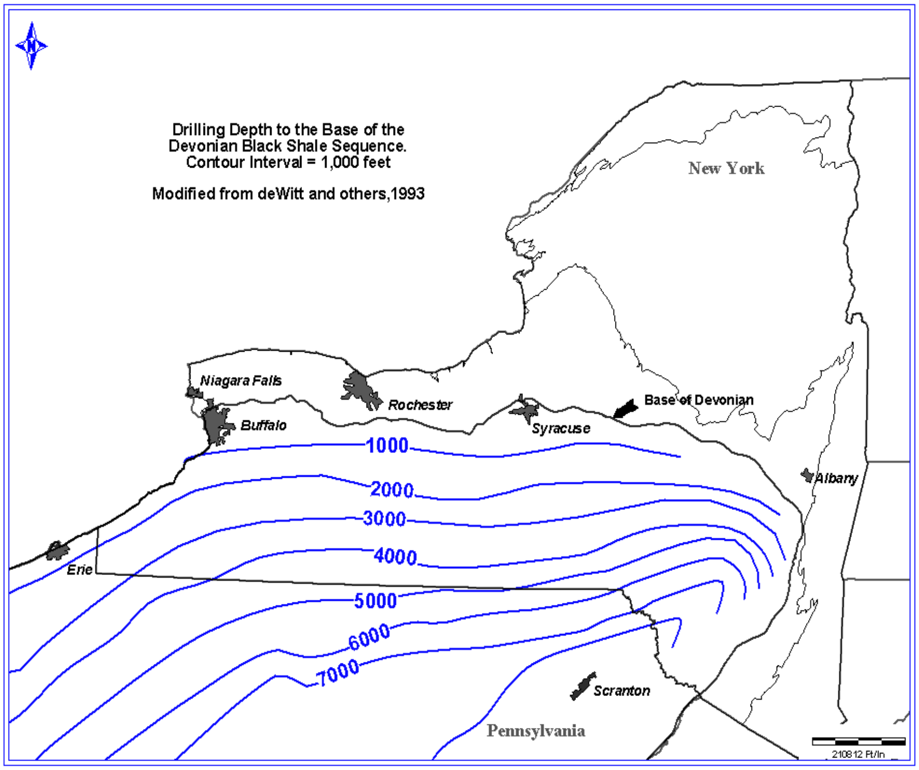


Figure 11. Drilling Depth to Base of the Devonian.

The uppermost unit of the black shale sequence is the Perrysburg Formation of the Canadaway Group. It is approximately 300 feet thick near Lake Erie and increases in thickness to over 700 feet in Cattaraugus County, thinning again toward Steuben County. The Perrysburg consists of a basal black shale, the Dunkirk Shale, overlain by the gray Gowanda Shale Member. The Dunkirk is another extensively deposited, organic-rich, black shale in the basin with equivalent shales (the Huron and Ohio) extending south into Alabama. In New York, the Dunkirk is a grayish-black to black shale containing some medium gray shales and siltstones in the upper part. It crops out and is well exposed in the vicinity of Dunkirk, Chatauqua County, ranging from 50 feet in thickness in the east to 110 feet in central Erie County. The black shale component of the Dunkirk varies from 50 feet in Chautauqua County to less than 25 feet in south Cattaraugus County (Van Tyne 1978). The overlying Gowanda Shale is a gray shale with siltstone and very fine-grained sandstone, and an occasional black shale bed. In central and eastern New York the black shale content diminishes rapidly as the two formations grade into one another. From the Late Devonian into the Early Permian, the basin continued to fill with coarse clastics primarily of continental origin, which were deposited as the delta migrated to the west.

Depositional Environment and Thickness

The environment of deposition for the marine shale sequences in the Appalachian Basin consisted of four broad regions; an alluvial plain, a shelf-delta front, the base of the slope, and the deep basin as depicted in Figure 12 (Kepferle 1993). Deposition of the Paleozoic sediments occurred as mountains, generally located to the east of the basin, were eroded. Sediments were then transported westward via a massive delta complex and deposited on the alluvial plain and into the marine environment.

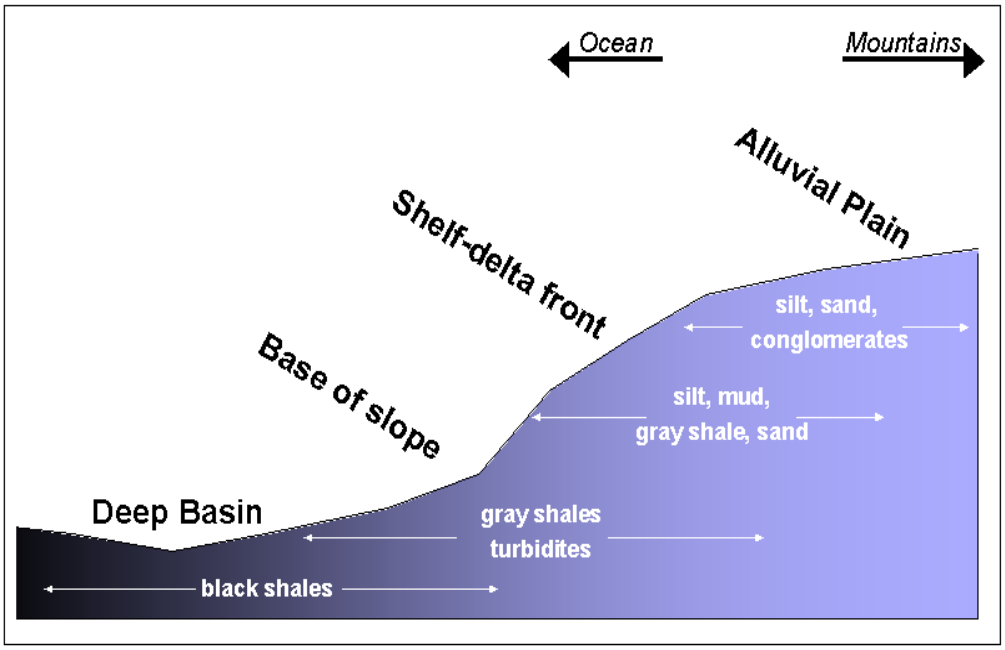


Figure 12. Deposition Model; Organic Rich Black Shales (modified from Kepferle 1993).

The resulting clastic wedge that formed from the eroding highlands consists of very thick coarse conglomerates and sands in the east that grade westward (seaward) into finer-grained beach deposits, and open marine deposits (de Witt 1993B). The black shales coalesce in western New York where the deep marine basin existed quite constantly, and most extend south and west into Pennsylvania, Ohio, and West Virginia to varying extents. Changes in sea level and fluctuations in the rate of sediment supply caused the transgression and regression of the marine environment. During marine transgression, the deep marine basin environment would expand up the slope, onto the shelf and perhaps even across the shore zone, spreading east and south. This is reflected in the deposition of black shales over gray and green shales and sands of the near shore environment. When sediment supply increased, or sea level dropped, the marine environment regressed and the delta complex and associated clastic rocks pushed westward, depositing the interbedded gray shales, siltstones and sandstones. This type of cyclic deposition occurred repeatedly during Devonian time affecting the extent of deposition of each interval.

The thickness of the Devonian black shale has been evaluated by several authors (Van Tyne 1978 and 1997; de Witt et al. 1993; Roen 1984). The thickness of each black shale bed is not depicted in this report but is well depicted by these authors in numerous reports, however thickness of each unit varies somewhat by author depending upon their methodology. Most black shales are easily recognizable on gamma-ray logs by their strong positive deflections (Figure 13). Thickness is determined by picking shale where the gamma-ray log exceeds 20 API units in positive value above the gray-shale base line. However, Van Tyne noted that “in certain cases, much of the black shale present does not exceed the 20 API limit and thus constitutes a thicker section than that measured in this way from the log," which was particularly true in the Genesee and Hamilton groups, where differences in thickness in the sample studies exceeded the log response pick by a factor of 10 or greater (Van Tyne 1978 1997).

de Witt noted in his evaluation that throughout much of the western and central Appalachian Basin the 20 API criteria are applicable in picking black shale thickness; however that it cannot be applied with assurance in the eastern part of the basin where the black shales lose the positive gamma-ray deflection (de Witt et al. 1993).

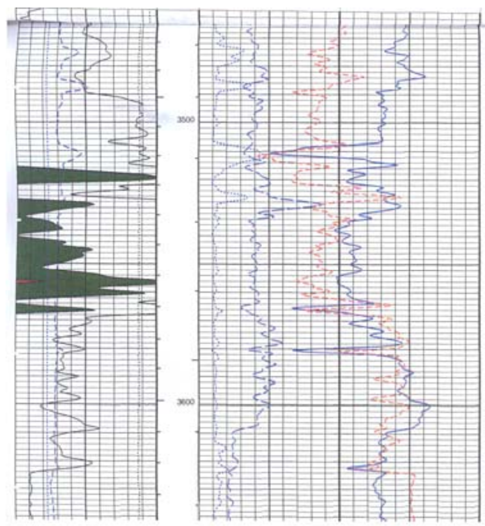


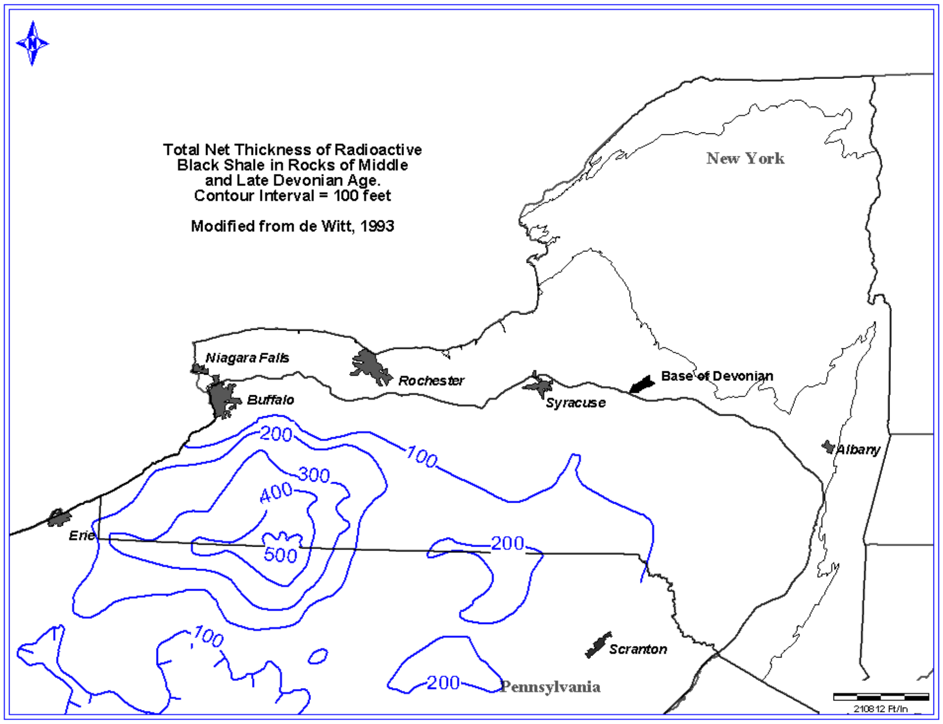
Figure 13. Sample Gamma Ray Signature of Radioactive Black Shale.

As determined by de Witt using the 20 API gamma ray cut off, the net thickness of radioactive Devonian black shale ranges from less than 100 feet to over 500 feet (Figure 14) (de Witt et al. 1993). Generally the individual black shale units thicken from the western-central portion of the Allegheny Plateau toward the south/southeast. They are thickest in south-central New York, and thin to the east, grading into gray shales. The Hamilton Group black shales (primarily the Marcellus) range from less than 50 feet to about 100 feet over much of southern New York, but locally thicken to over 250 feet in northeast Tioga County. The Genesee Formation black shale is present in the western and central portion of the Allegheny Plateau, and ranges from less than 25 feet to over 125 feet in southern Steuben County. The Sonyea Formation (Middlesex black shale) is fairly thin, and not widespread, and ranges from less than 25 feet to just over 75 feet. The West Falls Formation, containing the massive and extensive Rhinestreet Shale, ranges from less than 150 feet to over 300 feet in southwestern New York. The Perrysburg Formation is present in southwestern New York, and thickens from less than 50 feet Chatauqua County to over 100 feet in central Erie County.

Natural Fracturing

The rocks of the New York were strongly affected by the various stress regimes operating during the orogenic events mentioned above, and several episodes of natural fracturing occurred in the region. Natural fracturing is visible at many locations in New York. The regional fracturing patterns in the Devonian rocks of New York have been studied in depth by numerous authors including Parker (1942), Engelder (1980), Evans (1994), Gross (1991), and Loewy (1995).

Geologic studies of natural fracturing in New York indicate that various different vertical joint sets are present within the rocks of Middle and Upper Devonian age. Several different types of fractures (*joints*) are observed in the rocks, and each formed at different times and under varying circumstances (Wallace 1988). *Alleghanian joints*, are planar cracks that formed during the Alleghanian Orogeny in response to the compression exerted upon the rocks. *Release joints* formed during the Mesozoic Era, and resulted from rock expansion as erosion removed many layers of overlying rock. *Unloading joints* formed later as the rock cooled and reflect the present stress field in the region (Wallace 1988). The orientation of joints is related to the trajectory of the stress setting during the time of fracturing, and the orientation of maximum principal stress changed frequently during the regional geologic history. Generally, vertical joints propagate normal to the least principal stress, following the trajectories of the stress field at the time of propagation, thus the joint systems in the basin have varying orientations.



Figur14. Radioactive Black Shale Thickness in New York.

Organic content appears to have been a significant factor in joint development, and not all joint sets are present in all rocks. Black shales in particular have higher joint densities than adjacent gray shales, and joints often terminate at lithologic boundaries. Some beds contain several different joint sets, while adjacent beds may contain only one set, thus the total joint density in a rock layer depends on the organic content and propagation mechanisms of the jointing episode. Five joints sets have been categorized by Loewy (1995), and were determined based on their "clustering of orientation and similarity of morphology (Figure 15)." Timing of each jointing episode was determined by comparing abutting relationships between the different joints. Where a joint terminates against another joint, the terminating joint (abutting) is younger than the abutted joint.

Little literature exists that discusses the natural fracturing in the Silurian and Ordovician. Natural fractures are present however, as seen in road cuts through the Utica Shale (Figure 16). Earth Satellite Corporation performed a remote sensing and lineament analysis of the Appalachian Basin in New York (Earth Satellite Corporation 1997). The basis for their assessment was geologic interpretation of ten digitally-enhanced, LANDSAT Thematic Mapper ™ images within the outcrop of the Utica Shale in New York.

Order	Type	Orientation	Rock Type	Timing / Stress
1	Cross-fold (CF)* ***Set I****	N-NW (320 ^{o} to 010^{o})	Gray shale, siltstones, and Black Shale	Natural hydraulic fracturing during N-NNW compression during the Alleghanian Orogeny, formed at depth.
2	070^{o} ***Set III****	070^{o}	Black/Gray Shale ***below*** Rhinestreet fm.	Release joints, propagating during basin uplift and regional rifting, formed at depth.
3	E-W	085^{o}	Rhinestreet & Ithaca Formation Black shales only	Relaxing of formation tension resulting in release joints, formed at depth.
4	Fold Parallel (FP) ***Set II****	E-NE to E-W 070^{o} east to -045^{o} west	Gray/Black shale Most dense in and above the West Falls Group	Unloading, Release joints propagated during uplift.
5	E-NE	E-NE	Gray shale, siltstones, and Black Shale	Unloading-type, Parallel present stress field (Neotectonic), shallow depth.
* Consists of several joint sets that are not differentiated ** Regional set nomenclature by Parker (1942)				

Figure 15. Primary Joint Sets in New York

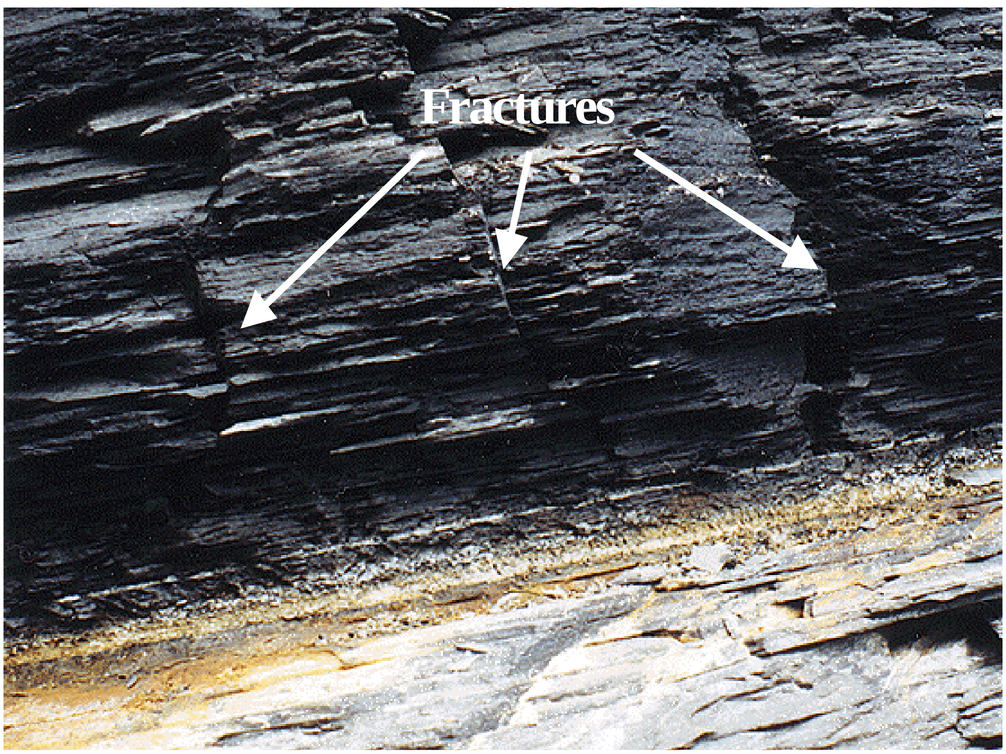


Figure 16. Natural Fracture in the Utica Black Shale (University of Rochester 2002)

The orientation of natural fractures was evaluated for four map sheets within the entire study area as depicted in Figure 17, and the rose diagrams generated by Earth Satellite Corporation (1997) show that the dominant fracture orientation varies from west to east. On sheets 1 and 2, where the upper Devonian sequence is very near the surface, northeast to eastern fractures dominate. On sheet 3, the wide variation in fracture orientation observed is “indicative of the fact that this sheet contains both the northeastern-most portion (mostly

Ordovician section of the Appalachian Basin) and the St. Lawrence Lowlands portion of the study." Sheet 4 shows that "the effect of basement reactivation of trends associated with the Taconic orogeny (N5^{o}-20^{o}E and N5^{o}-20^{o}W trends) and the trends associated with the eastern-most portion of the Appalachian Basin (N45^{o}-55^{o}W, N25^{o}-40^{o}E and N45^{o}-60^{o}E) are evident."

Joints in Ordovician-age rocks are more profoundly influenced by preexisting basement structure, the Taconic Orogeny and related structural grain, and by the Salina Salts, which would have transferred "almost all the stress from subsequent tectonism to the overlying sequences and resulted in a muted effect in the underlying Ordovician Rocks" (Earth Satellite Corporation 1997).

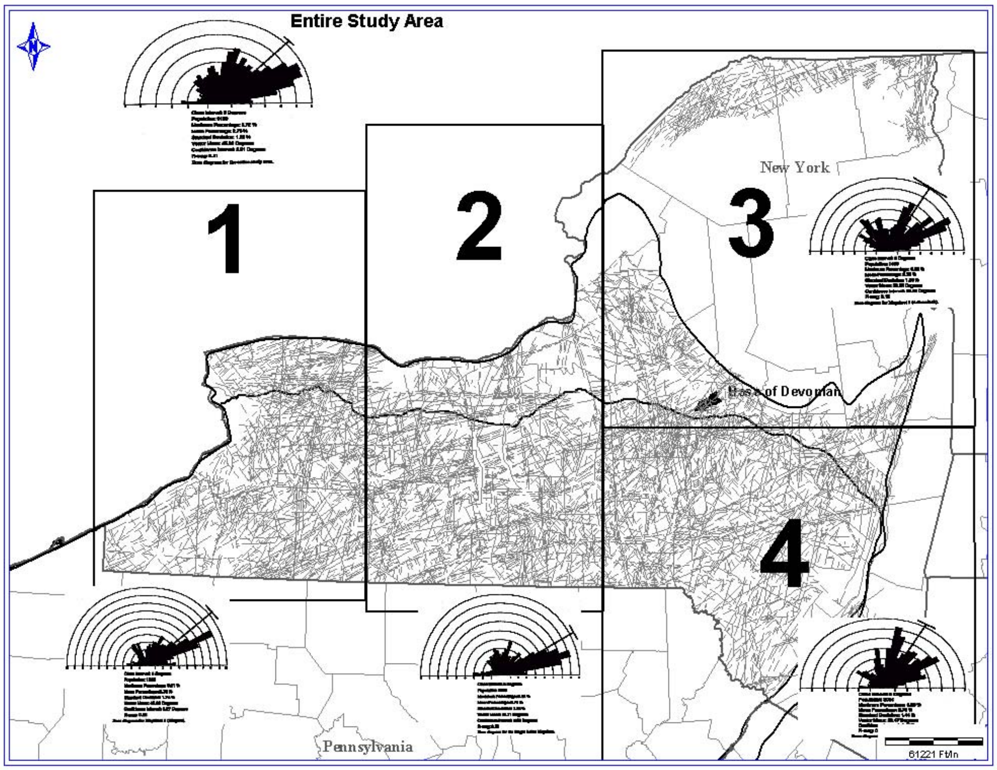


Figure 17. Rose Diagrams of Four Study Areas

GEOCHEMISTRY

Total Organic Carbon

Total organic carbon (TOC) measurements have been made on both core and drill cuttings in the Devonian Shale in New York. Figure 18 summarizes the measurements from core samples by formation (Streib 1981; Zielinski 1980) which are averages of multiple data points from individual wells and from multiple wells by Devonian Shale member. TOC values range from low values less than 0.5% in the upper Devonian shales to over 6% in the middle Devonian shales. The data also show a general trend of increasing TOC going from central New York to western New York as well as a general trend of increasing TOC with depth or age of Devonian Shale (Figure 19). However, Weary, et. al. describe a Middle Devonian Marcellus cuttings sample taken from Livingston County with a measured TOC of 11.05% (Weary 2001). Finally, recent measurements taken from samples collected by the New York State Museum show total organic content in the Marcellus Shale to range from less than 1% to over 9% TOC (Martin 2006).

Group	Member	Van Tyne 9 Well Data Set Average All Wells / All Depths Total Organic Carbon (%)	USGS 20 Well Data Set Average All Wells / All Depths Total Organic Carbon (%)
	Dunkirk Shale		1.14
Canadaway			
	Hanover Shale		0.80
Java			
	Pipe Creek		
	Angola Shale		0.89
West Falls			
	Rhinestreet Shale	1.95	1.47
Sonyea	Cashaqua Shale		0.65
	Middlesex	2.83	1.50
	West River		1.34
Genesee	Pen Yan Shale	2.40	1.58
	Geneseo Shale	4.00	0.92
	Hamilton Shale		0.80
Hamilton			
	Marcellus Shale	6.05	3.87

Figure 18. TOC from New York Devonian Shale Drill Cuttings

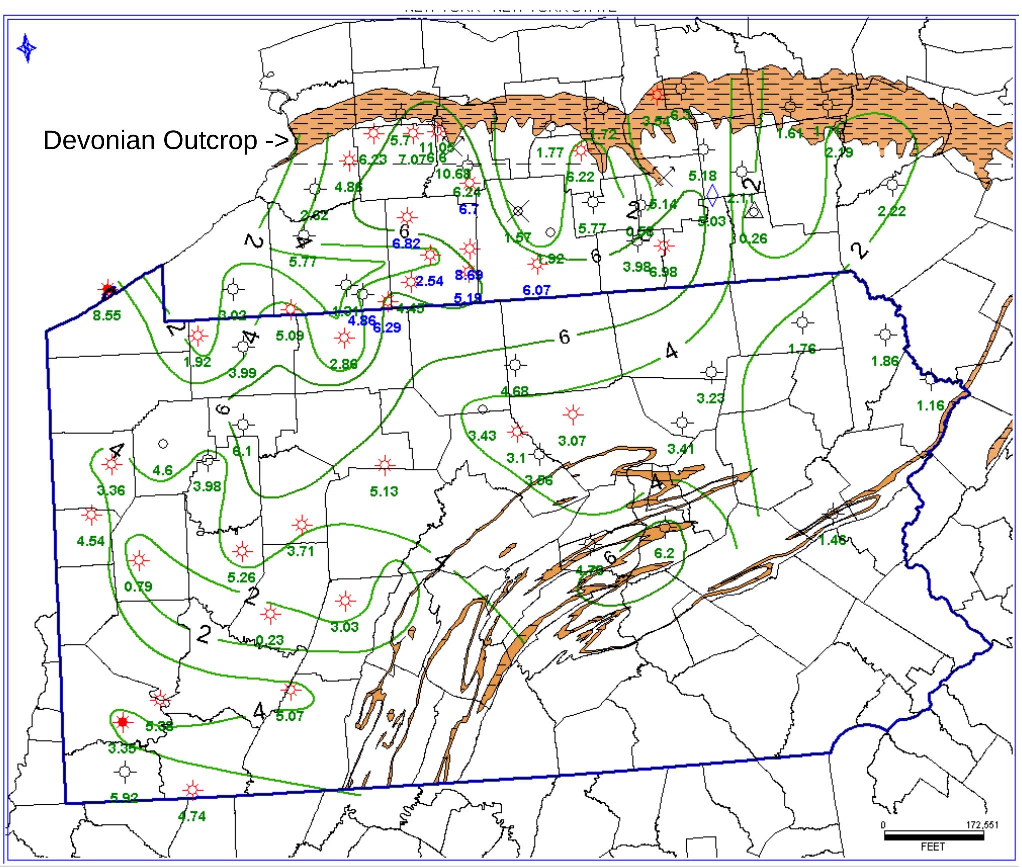


Figure 19. Average Organic Content (%) of the Marcellus Shale in New York and Pennsylvania (SOURCE Weary 2001; Repetski 2002, 2005, graphic by Frank Maio)

Measurements of total organic carbon in the Utica Shale have been reported in literature (Hay 1989; Hannigan 1994; Ryder 1998; Wallace 1988). The range is from approximately 0.16% to 4.0%. Recent measurements from samples collected by the New York State Museum range from less than 1% in western New York State to 5% in the east (Nyahay 2007).

Kerogen Type

Published Rock-Eval data for the Marcellus shale and the Utica Shale in New York State was plotted on a modified Van Krevelen diagram (Figure 20) (Weary 2000). The data show that the Marcellus is primarily Type II kerogen with a mixture of Type III and the Utica is primarily Type III kerogen with a mixture of Type II. Both shales with these kerogen assemblages are capable of generating liquids and gases. No Rock-Eval data is available for the Silurian shales in New York.

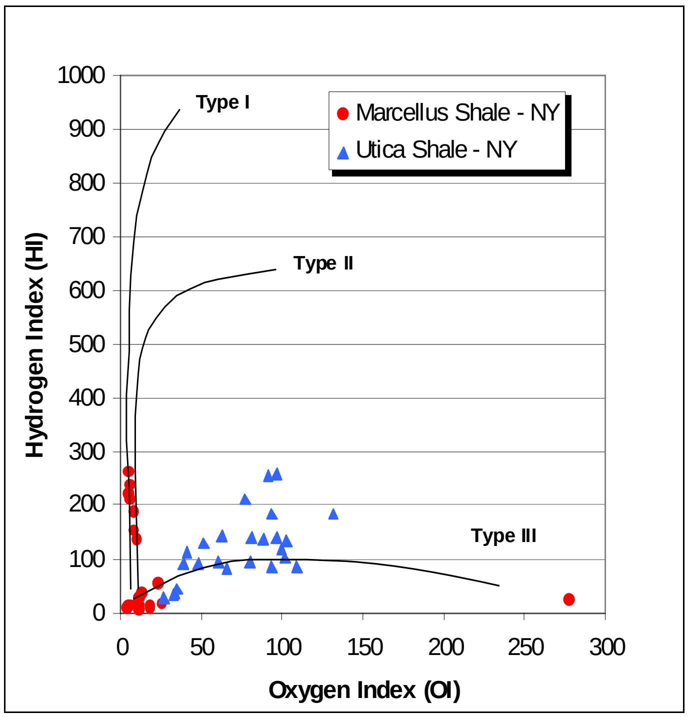


Figure 20. Published Rock-Eval Data for Marcellus and Utica Shales in New York Plotted on a Modified Van Krevelen Diagram.

Thermal Maturity

Rock-Eval. --- Published Rock-Eval data for the Marcellus Shale and the Utica Shale in New York State was plotted using the technique after Peters (Figure 21) (Weary 2000). This figure shows the spread of maturity of the samples measured. The samples were from different depths and ranged from central New York to western New York. The bimodal disctribution of Marcellus samples most like reflects the variability of shale maturity as shown by vitrinite reflectance and conodont alteration estimates (see below). The "scattershot" nature of the Utica data reflects the inability to pick a clear Tmax with extremely mature shales and may not be significant.

Vitrinite Reflectance. --- Figure 22 summarizes vitrinite reflectance data from nine wells in the Marcellus Shale (Van Tyne 1993). There is a general trend of increasing thermal maturity going from western New York toward central New York. This general trend in the Marcellus Shale is further supported by the vitrinite reflectance data reported from drill cuttings in the USGS report by Weary (2000) (Figure 23).

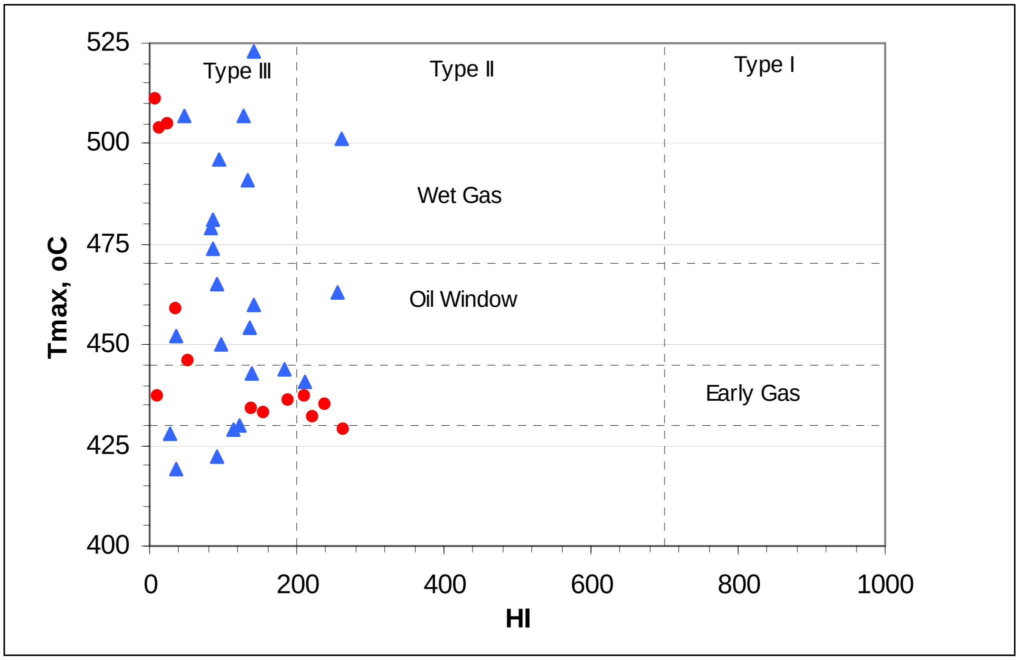


Figure 21. Published Rock-Eval Data for Marcellus and Utica Shale. Plotted After Peters (1986)

Well / County	**Depth (ft)**	**Ro (%)**
St. Bonaventure , Cattaraugus County	3,600-3,640	1.23
Portville Central School, Cattaraugus County	4,140-4,180	1.2
Houghton College #1 Allegany County	2,270-2,290	na
Houghton College #2, Allegany County	2,380-2,410	1.18
BOCES Fee #1, Allegany County	3,240-3,290	1.27
Meter #1, Livingston County	1,570-1,600	1.31
Alfred University #1, Allegany County	3,950-3,960	1.65
Hammel #1, Allegany County	4,662-4,690	1.65
Valley Vista View #1, Steuben County	3,882-3,895	1.65
Average All Wells / All Depths		**1.39**

Figure 22. Summary of Thermal Maturity Data; Marcellus, New York

For the Ordovician rocks, Utica Shale vitrinite reflectance calculations were calculated by TerraTek, Inc. from samples collected by the New York State Museum from graptolites identified in core taken in the Mohawk Valley. Graptolytes were used because no vitrinite was present at the time the Utica was deposited in the Ordovician (TerraTek 2004). Average vitrinite reflectance values ranged from 2.28 to 4.32. The higher value, which reflects a super-mature sample, was the farthest east sample taken in Saratoga County. This reflects the increasing maturity in the Utica as one moves closer to the Taconic front.

Thermal Alteration Index. --- The uses of progressive changes of color and/or structure of pores, pollen or plant-cuticle fragments is also used as an indicator of thermal maturation of the kerogen. Kerogen coloration is reported on a scale of 1 to 5, and is referred to as Thermal Alteration Index (TAI) (Staplin 1969). Different types of spore or pollen grains can show different sorption values at low levels of maturation. TAI averaged 3.20 for the Rhinestreet Shale interval from NY#3 well in Steuben County New York that was cored from 1,203 to 1,263 feet (Streib 1981). Similar TAI values were measured from the NY#4 well in Steuben County, 3.2 for the Geneseo (2,970 – 3,080 feet) and 3.4 for the Marcellus (3,780 – 3,842 feet) (Streib 1981). All samples indication maturation levels above 150^{o}C. No TAI values were available for the Silurian or Ordovician shales.

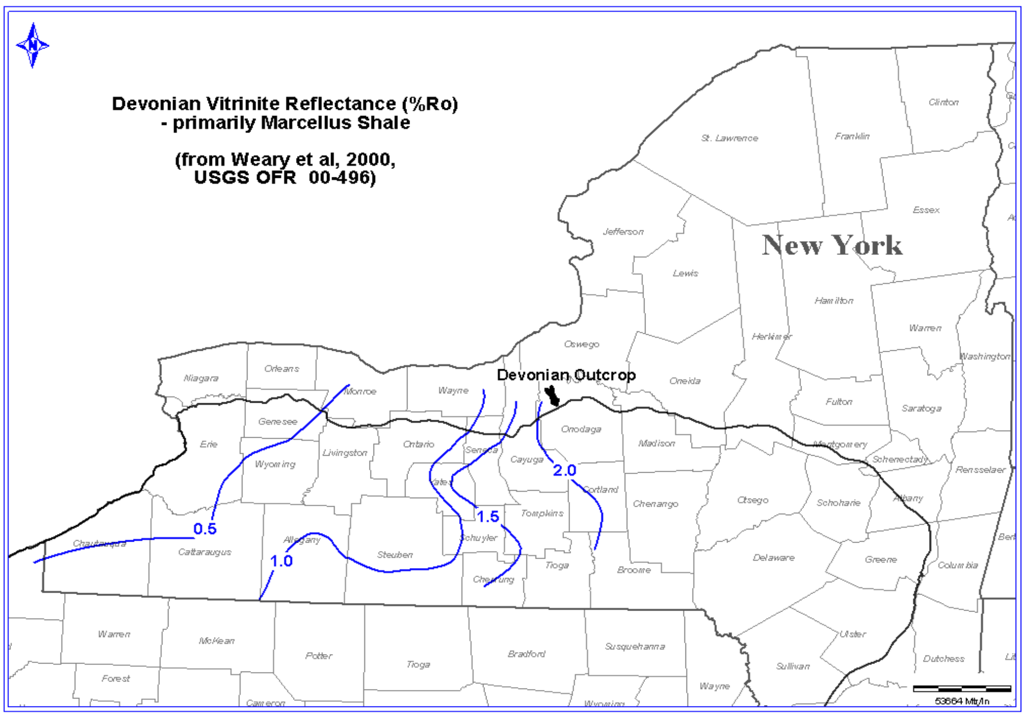


Figure 23: Devonian Vitrinite Reflectance (%Ro)

Conodont Alteration Index. --- The thermal maturity of shales can also be inferred from published conodont alteration indices (CAI), a scale of color alteration in conodonts (a marine fossil) (Epstein 1977). In general, the CAI of a conodont increases with depth and temperature as a result of metamorphisim. A recent study of thermal maturity in Ordovician and Devonian rocks has been completed by the USGS and New York Geological Survey (Weary 2000). In the Upper Devonian shales, CAI values range from less than 1.5 to 2.5 west to east. In Middle Devonian shales, CAI increases from about 1.5 in western New York to 2.5-3 in the central area (Figure 24) (Tetra Tech 1980). Silurian CAI values are similar to the Middle Devonian. Upper Ordovician rocks in western New York have CAI values of 2-3, which put them within the more advanced stage of wet gas generation. In southern New York, where CAI values are 3-5, the Ordovician rocks are prospective for dry gas (Figure 25) (Weary 2000).

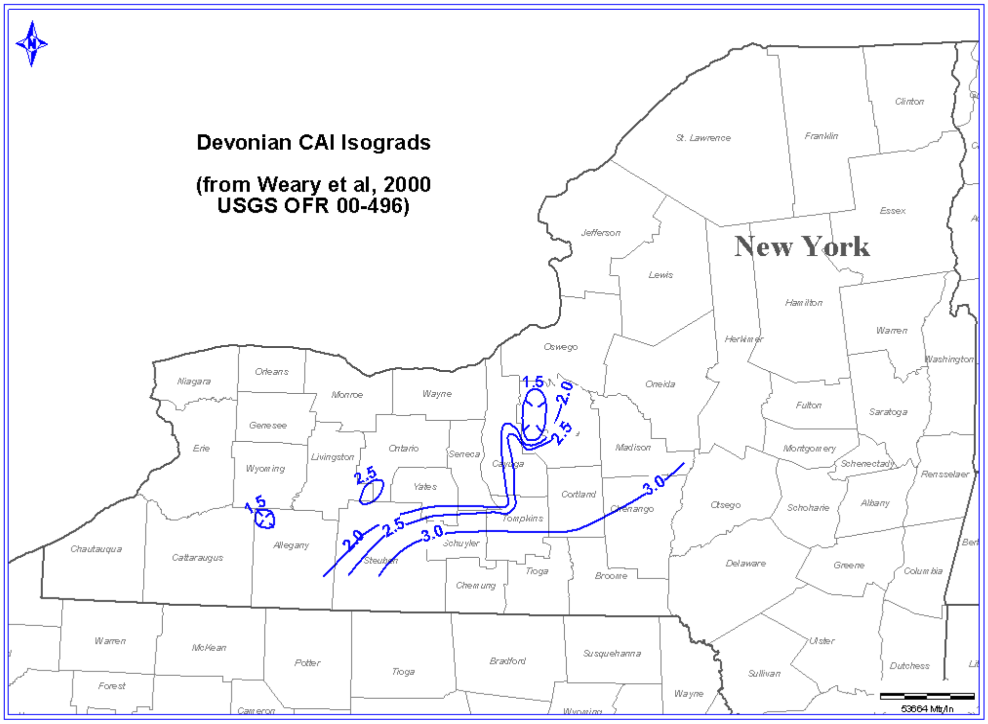


Figure 24. Devonian Conodont Alteration Index (CA) Isograds.

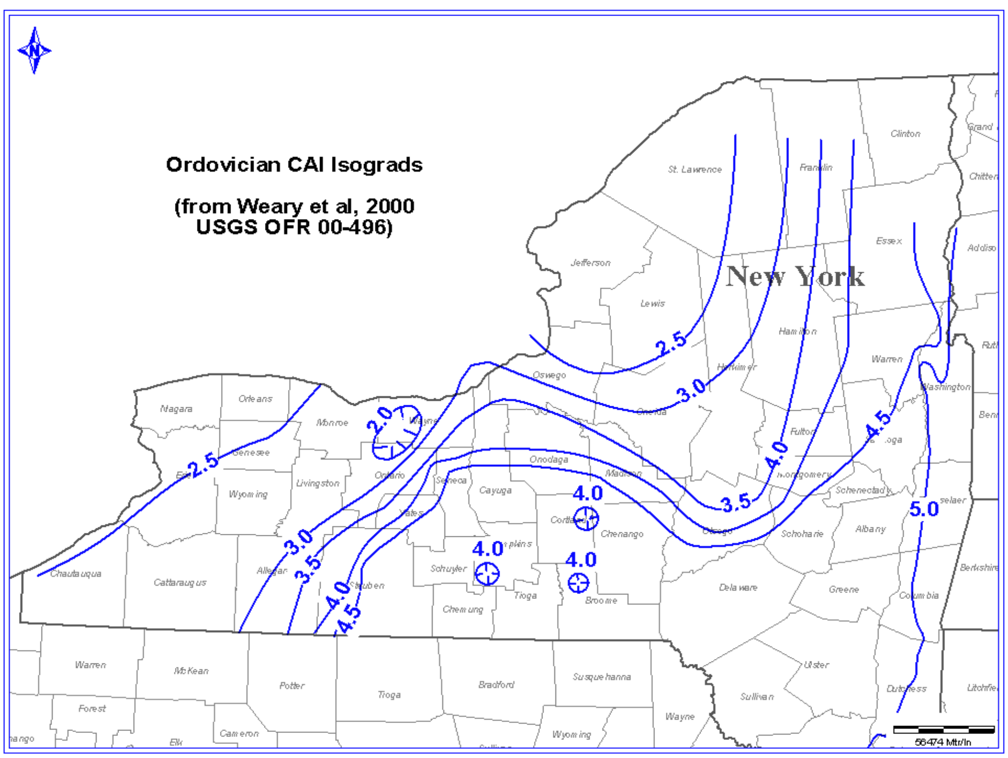


Figure 25. Middle and Upper Ordovician Conodont Alteration Index (CAI) Isograds

Adsorption

Adsorption isotherms have been created for both the Devonian and Ordovician shales. According to Terratek, “the testing consisted of exposing, at constant temperature, the E.Q. moisture prepared coal sample to methane gas at a series of pressures, calculated to yield the desired equilibrium adsorption pressures (TerraTek, 2004; TerraTEk 200_).” As expected, the Marcellus sample, with TOC of 8.27%, can yield much higher adsorption figures compared to the Utica given the higher organic content, with TOC of 2.167%. (figures 26 and 27). Other factors such as clay content also affect adsorption.

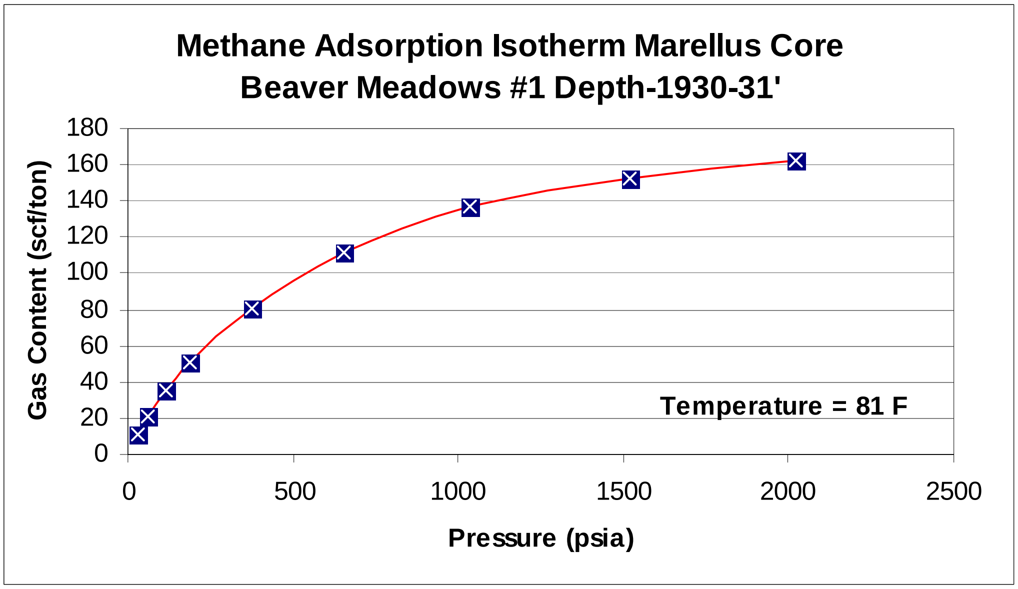


Figure 26. Methane Adsorption Isotherm for a Marcellus Shale Sample, New York

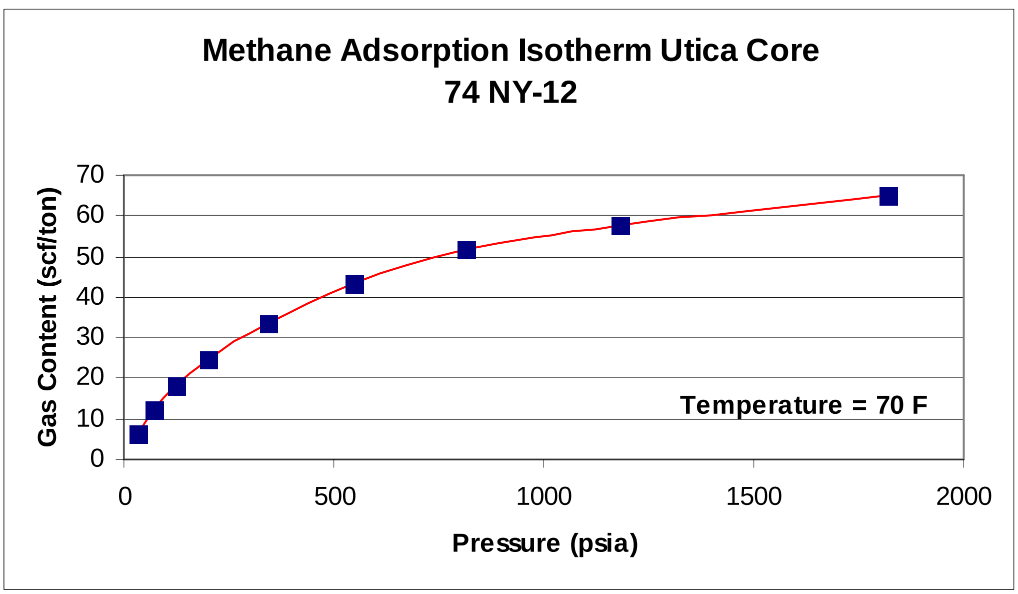


Figure 27. Methane Adsorption Isotherm for a Utica Shale Sample, New York (Terratek 2004)

RESERVOIR CHARACTERIZATION

Reservoir characterization in gas shale reservoir systems focuses primarily on natural fractures because most known productive gas shale reservoirs are gas saturated with extremely low permeability and required multiple sets of open natural fractures for commercial production of natural gas. There are other properties that are also important in characterizing the reservoir potential of shale. These properties are covered below and information is provided for the shales in New York where available. Unfortunately, there is very little published data on the reservoir properties of the shales in New York. The majority of the data comes from the three wells cored and studied as part of the US DOE Eastern Gas Shale Project. Additional data has been published on drill cuttings.

Mineralogy

Both the Ordovician Utica and Devonian Marcellus shales are calcareous shales. Xray diffraction was completed on a number of outcrop samples. Figures 28 and 29 show ternary diagrams of the relationship between three major constituent categories as derived from XRD data (Nyahay 2008B). The relationship shown here is quite similar to the productive units of the Barnett shale (Nyahay 2008B). To full assess the mineralogical composition, far more data is needed.

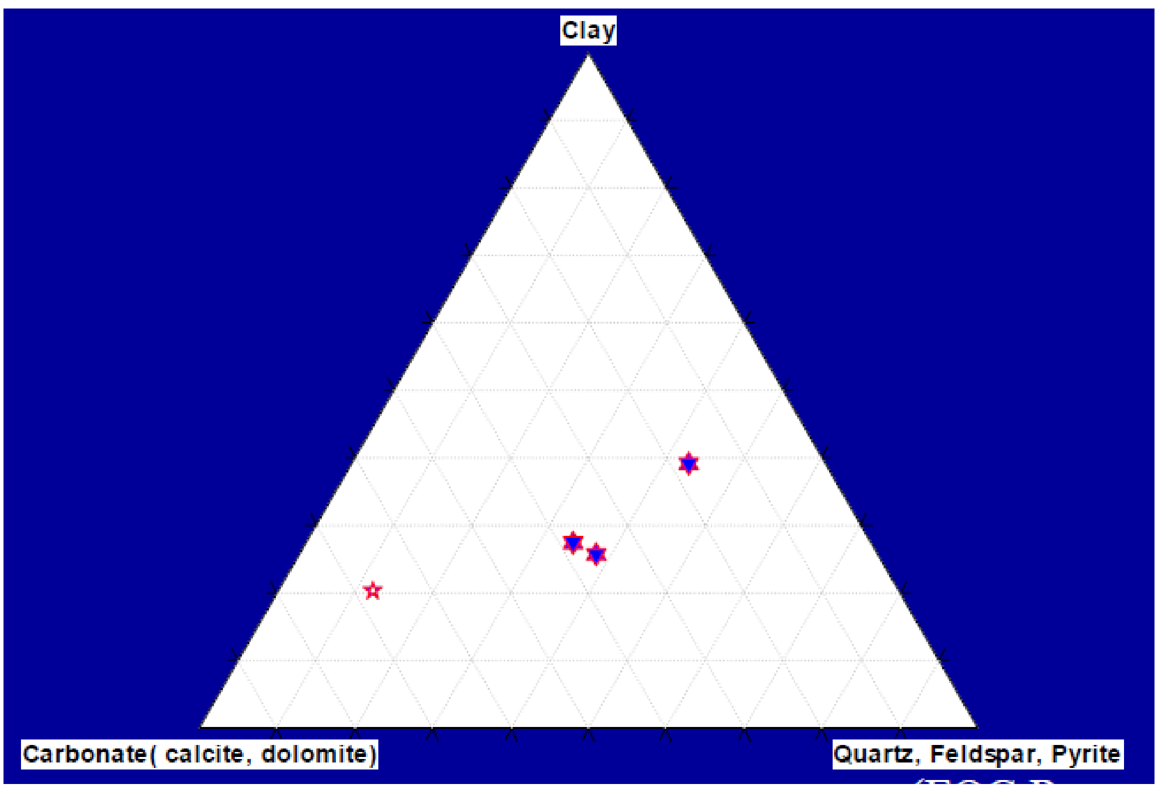


Figure 28. Utica Shale Ternary Diagram from Outcrop Samples (Nyahay 2008B)

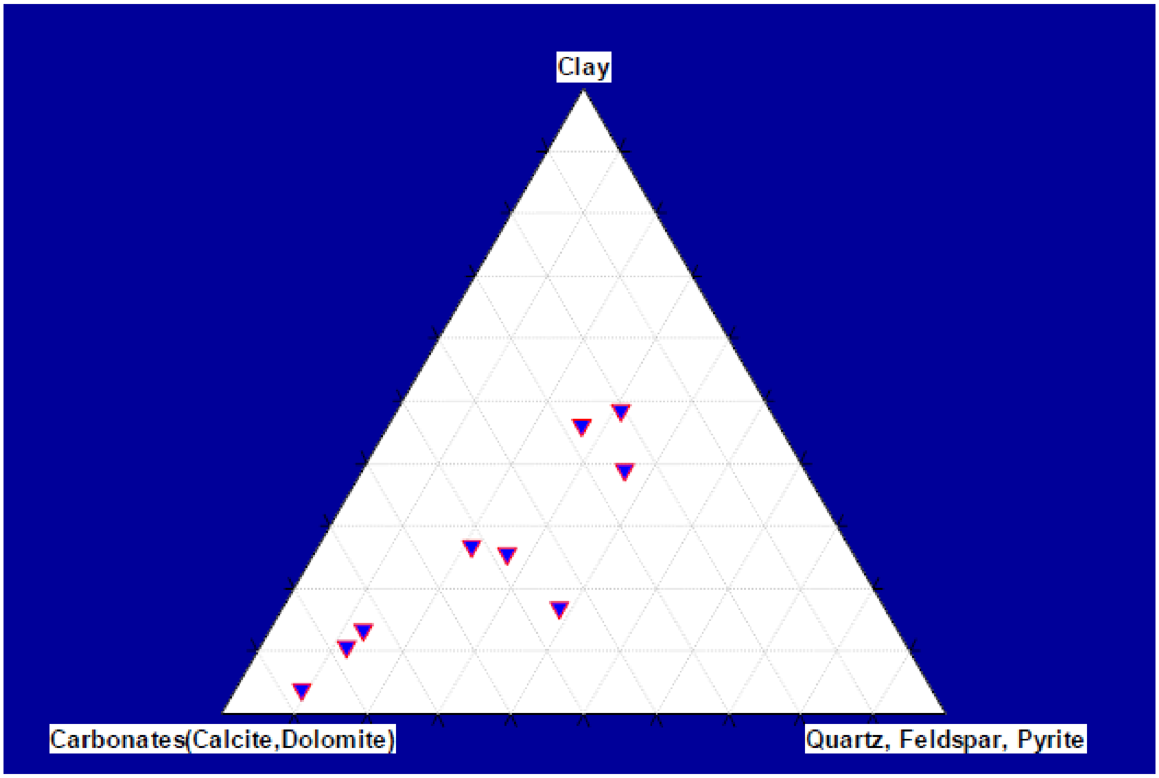


Figure 29. Marcellus Shale Ternary Diagram from Outcrop Samples (Nyahay 2008B)

Natural Gas Composition

The composition of produced natural gas can have an impact on the overall economics of a gas shale play as well as provide information related to its source. In several fractured shale gas plays, the composition of the produced natural gas impacts economics and provides evidence of microbiologic and thermogenic processes (Walter 1997 2001). Unfortunately, very little gas chemistry and gas and water geochemistry is available from shales in New York. Thus it is difficult to attempt to draw comparisons to other gas shale plays, such as the shallow biogenic Antrim shale play in northern Michigan Basin.

The best source of natural gas composition from gas shales in New York was from the USBM project that looked at produced natural gas composition across the United States (Moore 1987). In this report, six wells with natural gas production from Devonian Shale were analyzed along with one water well and one natural seep. The data shows methane concentrations of 80-95% and concentrations of ethane and propane from 3% to 15%. The heating value of the gas measured (BTU) ranges from 901 to nearly 1300 BTU's. The majority of the data points were sampled in 1979. No detailed geochemistry is available to investigate the biogenic or thermogenic processes. No gas composition data is available for either the Silurian or Ordovician shales of New York. However, analysis of the Utica Shale in Quebec indicate methane concentrations ranging from 88% to 96% (Beiers 1976).

Natural Fractures

Natural fracture formation was addressed previously. In New York, only a minimum amount of oriented core has been taken in the shale reservoir systems for natural fracture characterization and no formation imaging logs or down-hole cameras results have been published for natural fracture characterization. Three wells were cored in New York during the DOE Eastern Gas Shale Program and natural fracture characterization was published for them (Cliff Minerals, Inc. 1980B 1980 1981).

Figure 30 summarizes the natural fractures identified in the oriented core from the Devonian Shale for the three research wells. No significant shows were associated with the cored intervals. The NY #1 (NYSERDA #3-6213) well was completed in the Marcellus Shale (which was not cored) and was producing intermittently at the end of 2001.

Group	Unit	NY #1 Depth (ft)	Fracture Orientation	NY #3 Depth (ft)	Fracture Orientation	NY #4 Depth (ft)	Fracture Orientation[1]
Canadaway	Dunkirk Shale	370-515	N85°W (1) N85°E (2)				
Java	Hanover Shale	515-546					
		963-984					
	Pipe Creek	984-1018	N85°E(2)				
West Falls	Angola Shale	1018-1021					
		1328-1355					
	Rhinestreet Shale	1335-2345	N35-45°W (4) N70-90°W (7) N70-90°E (14)	1203-1263	N2°E (2) N48°E (1)		
Sonyea	Cashaqua Shale	2345-2359	0				
		2486-2495					
	Middlesex	2495-2629	N45-65°E (4) N25°W (1)				
Genessee	West River	2629-2664	N40-70°E (1) (4)				
		2723-2730					
	Genundewa	2730-2737	N20°W (1)				
	Pen Yan Shale	2737-2866	N70-80°W (3) N35°E (1)				
	Lodi Limestone	2866-2876	0				
	Geneseo Shale	2876-2924	N35°E (1) N80°W (1)			3010-3080	N50°W (1)
Hamilton	Tully Limestone	2924-2929	0			3080-3084	N50°E - N60°E (major)
	Marcellus Shale					3790-3842	N50°W - N60°W (minor)

[1] Six feet of the Onondaga formation was cored and included 2 joints, 3 microcracks and 10 faults - major trend is N20°W-N30°W. (Cliff Minerals, Inc. 1980B 1980C, and 1981)

Figure 30. Devonian Shale Natural Fracture Orientation from Oriented Core.

Based on available information, cumulative production from the Devonian Shales is 15.89 Mmcf. The greatest number of fractures from the core analysis was in the Rhinestreet shale, which was not completed in this well. The NY #3 (Scudder #1) well was not completed and was plugged and abandoned. The NY #4 (Valley Vista View #1) well tested the Rhinestreet which proved to be poor and was eventually completed in the Marcellus Shale and produced for a short period of time before it was plugged and abandoned. Unfortunately, due to completion circumstances and poor well performance, no observations can be made for improved well performance related to the presence of orthogonal natural fractures. No core or subsurface natural fracture descriptions are available for the Silurian or Ordovician shales.

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A QUICK LOOK AT NEW YORK'S UNCONVENTIONAL GAS SHALES: SUMMARY AND ROAD LOG

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The purpose of this trip is to take a look at the organic rich shales that are driving unconventional shale exploration in New York. These shales are the Devonian Marcellus and the Ordovician Utica. Over the last fifteen years Ver Straeten has worked on the Marcellus Formation and Brett, Baird, Mitchell, Jacobi, and Joy have worked on the Utica Group to determine new and improved correlations. These shales have joined the list of names such as the Barnett, Antrim, and Fayetteville to rejuvenate interest in organic rich shales for natural gas exploration

Technology has been able to make these highly impermeable shales that have long been considered nuisance gas in the past highly profitable. This then begs the question why are these shales highly productive? In order to answer this shales are being studied intensely across the nation. Of all the shales being studied the Barnett shale is the one everyone looks to use as the model. The Barnett shale exploration was inspired by the Eastern Gas Shale Project in the Appalachian Basin that took place during the late seventies and early nineteen eighties (Steward 2007).

The Utica Group is been divided into three formations, the Upper and Lower Indian Castle, the Dolgeville and the Flat Creek. The Dolgeville is coeval with the Rust and Steuben of the lower Trenton Group. The Flat Creek is coeval with the Denley of the Lower Trenton Group.

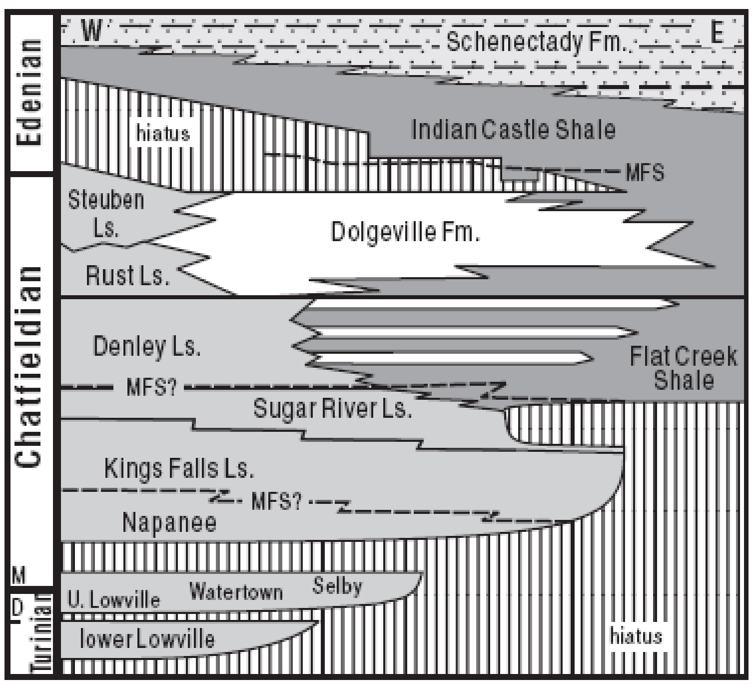

Figure 1. Utica Chronostratigraphy (Joy, et al. 2000)

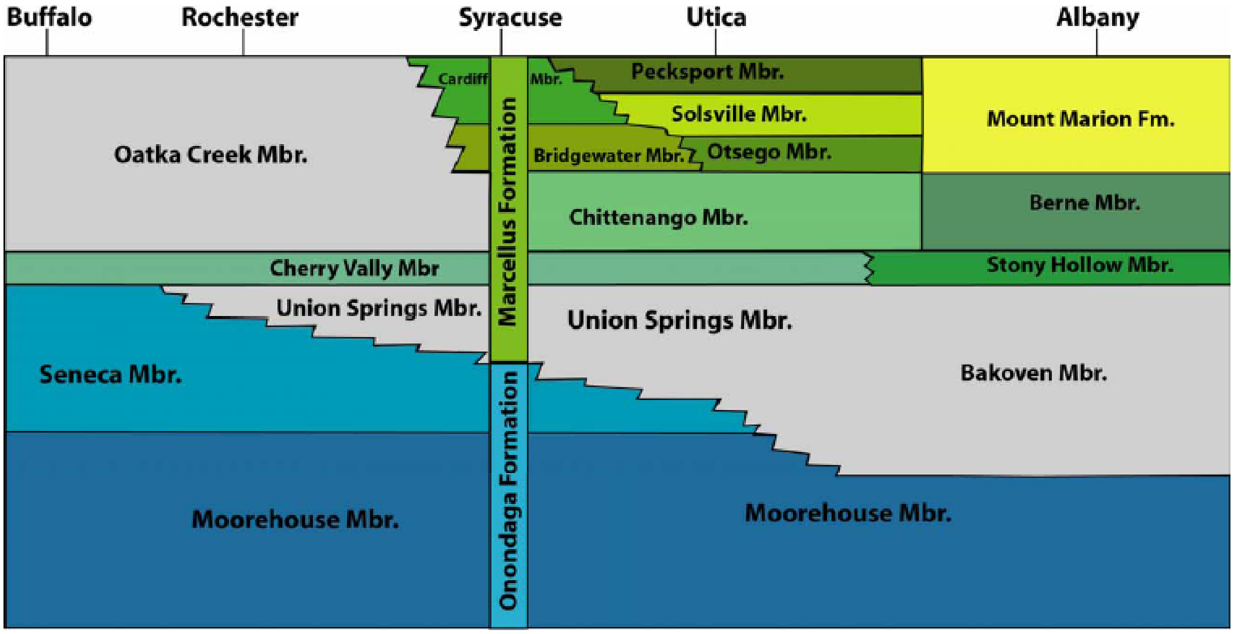

Figure 2. Marcellus Stratigraphy (Rickard 1975)

The Marcellus members we are interested in are the Chittenango, the Cherry Valley, and the Union Springs. Interestingly enough, the patterns of the Barnett, Lower Marcellus and Utica are similar. In the Barnett, the upper and lower shale units are separated by the Forestburgh Limestone. In the Lower Marcellus, the Chittenango, and the Union Springs are separated by the Cherry Valley Limestone. In the Utica Group, the Indian Castle and Flat Creek are separated by the ribbon limestones of the Dolgeville.

The deepening upward sequence that is evident within the Trenton Group and the overlying Utica Group was interpreted as a product of lithospheric flexure caused by the loading of the Laureantian margin during the Taconic Orogeny (Jacobi and Mitchell 2002). Numerous normal faults occur in the region and many underwent significant displacement during the deposition of the Trenton and Utica Groups as a consequence of the attempted subduction of the Laurentian margin (Jacobi and Mitchell 2002). Differential rates of subsidence across the region produced accommodation rates that increased by an order of magnitude from west to east. High rates of siliciclastic sediment supply and even higher subsidence rates resulted in thick deeper water succession to the eastern part of the study region (Joy, et al. 2000). The coeval Trenton Group succession in the west formed under conditions of moderate sediment supply and lower subsidence rates resulting in net shallowing in the upper succession (Joy, et al. 2000).

Baird and Brett (2002) noticed that both the Ordovician Indian Castle and Devonian Marcellus developed as onlap disconformities in a similar western cratonward side of a synorogenic foreland basin. These units record the transgressive drowning of a carbonate shelf with the ensuing establishment of a stratified dysoxic to anoxic bottom conditions.

The Dolgeville Formation formed as a short duration pulse of calciturbidites (Merthens 1988). The Dolgeville formation records a turbidite deposition on a subtidal slope well below fair weather base (Baird and Brett, 2002). The presumed source of the turbiditic carbonate was from the carbonate shelf to the north and west, storm energy as well as seismic activity disturbed sediments in the in the upper slope triggering gradient current and density flows(Baird and Brett 2002).

The Flat Creek Formation was formed as the basin deepened to the east, suggesting that differential tectonic subsidence was the dominant mechanism creating accommodation space (Joy, et al. 2000).

The Indian Castle Formation is divided into two units, the lower Indian Castle and the Upper Indian Castle separated by the Honey Hill disconformity (Baird & Brett 2002). It is overlain by the Schenectady and Frankfort formations and underlain by the carbonates of the Trenton to the west and the Dolgeville Formation to the east. The lower Indian Castle corresponds to the Climacograptus (Diplacanthograptus) spiniferous assemblage zone of graptolites and possibly to the lowermost Geniculograptus Pygmaeus Zone. The lower Indian Castle is hard blocky to sheety black shale with bundles of tabular impure limestone (Baird & Brett 2002). The upper Indian Castle corresponds to the Geniculograptus Pygmaeus Zone and is composed of fissile to platy shale that lack limestone beds (Baird and Brett 2002).

The Dolgeville Formation is distinctive unit that is composed of closely spaced rhythmic dark gray shale and calcisiltic ribbon limestones. The Dolgeville is overlain by the Thruway unconformity that bevel the soft sediment slumps and fault scars of the Dolgeville and is overlain further to the west by the onlapping upper Indian Castile and is underlain by the Flat Creek to the east of the Little Falls fault and lower Trenton Group carbonates west of the fault. The Dolgeville corresponds to the time range of Orthograptus ruedemmani assemblage zone of graptolites with only a very narrow segment of the basal Climacograptus (Diplacanthograptus) spiniferous assemblage near its top (Goldman, et al. 1994). Baird and Brett have traced the Dolgeville as far east as the Hoffman's Fault. Goldman, et al. 1994 describe the Dolgeville as a thin to medium bedded tabular silty black micrite and interbedded black shales. In subsurface cores at the NYSGS core repository the Dolgeville will show a white weathering color that distinguishes them from the Flat Creek below and Indian Castle above.

The Flat Creek Formation rest disconformably over the Sugar River of the Trenton Group (Joy, et al. 2000) and is gradationally overlain by the Dolgeville Formation (Goldman et al. 1994). The Flat Creek corresponds to the the Corynoides Americanus graptolite Zone assemblage to 8 meters below the base of the Dolegville where is corresponds to the Orthograptus ruedemmani assemblage zone (Goldman et al. 1994). The Flat Creek is a calcareous, laminated black shale with interbedded calcilutites.

The Chittenango is a jet black, very fissile shale, with no limestone beds as in the Union Springs. It is noncalcareous until the basal section near the top of the Cherry Valley.

The Cherry Valley is a black to brown argillaceous limestone that gives a petroleum odor when broken. It is divisible into two massive parts separated by a shaley limestone or shale containing limestone nodules, and pyrite.

The Union Springs is the basal Marcellus which is a black, pyritiferous, fissile, calcareous shale with concretions, and thin limestone layers.

ROAD LOG

Cummulative mi.	Miles from last Point.	
0.0	0.0	Start at Exit 27 of NYS Thruway
0.1	0.1	Take right and proceed to NYS RT 5S.
1.0	0.9	At the light take a left onto RT 5S and proceed to Stop1

Stop 1. South Chuctanunda Creek and Rt. 5S Roadcut

This outcrop features the Flat Creek Formation. This shale can be further subdivided into a black fissile shale overlying a dark gray calcareous mudstone with black shale interbeds. A north trending vein displays a nearly vertical fault that might show possible lateral slip, with horizontal slickensides next to a clastic dike (Neptunian?) that strikes north. Could this be evidence of seismic activity that triggered the turbiditic flows of the Dolgeville? Vertical calcite veins cut the roadcut and strike nearly E-W. The roadcut as seen from the east shows a synclinal structure. Bentonite beds are can be easily picked out by their rusty colored weathering. The creek shows these E-W vertical calcite veins that show extensional horsetail features, breccias and possible small pull aparts. In the black fissile shale there is evidence of thrust faulting. Before Baird and Brett (2002), this outcrop could have been interpreted as Utica Shale undivided, but with the Dolgeville extended as far east as Hoffman's fault, this make this shale a good candidate to be the Flat Creek shale. Collections of graptolites would confirm this more correctly.

Cummulative mi.	Miles from last Point.	
2.0	1.0	Turn right on to Rt. 5s going east to Rt. 30. Bear right on to Rt. 30 south then quickly turn left on to the New York State Thruway (90) entrance ramp. Proceed west on Thruway to Exit 29A Little Falls
35.3	33.3	Dolgeville outcrop on both sides of the NYS Thruway (90)
39.3	4.0	Exit 29A, proceed to toll both. Black Canyon outcrop of Indian Castile to the south on ramps going east.
39.8	0.5	Turn left on road leading to Rt. 5S
40.1	0.3	Turn right on Rt. 5S and go west to Paradise Road.
42.8	2.7	Turn left on to Paradise Road and proceed south.
43.3	0.5	Turn left and park by gated fence.

Stop 2. Paradise Road NYS Thruway Road Cut

This roadcut displays Lower Indian Castile Formation overlying the Dolgeville Formation separated by the Thruway unconformity. To the eastern end we see a normal fault, a splay of the Little Falls fault. Weathered bentonite beds outline the dark gray Lower Indian Castile showing the blocky limestone beds and indurated shales overlying the alternating tabular beds of dark gray to black limestone and shale with fold trains that have a west vergence. Following the weathered bentonite beds small sags can be seen in the Lower Indian Castile. A fault at the eastern end displays the Upper Indian Castile down dropped to the east against the Lower Indian Castile and Dolgeville formations. Notice the lack of weathered bentonite beds in the Upper Indian Castile. The Upper Indian Castile is a dark gray fissile shale. These laminations can be seen from the perch of the farm field we are standing on.

Cummulative mi.	Miles from last Point.	
		Turn around and proceed north to Rt 5
43.8	0.5	Turn right on to Rt. 5S and go east to Rt. 80 in Fort Plain
58.6	14.8	At the light turn right onto Rt. 80 going south.
59.6	1.0	Turn right into the parking area for the Town of Minden court and road maintenance.

Stop 3. Otsquago Creek Outcrop

This creek cut shows the contact between the Dolgeville tabular alternating beds of limestone and shale overlying the Flat Creek black fissile shale . Calcite filled veins cut both the Flat Creek and Dolgeville Formations. The veins trend N 110° - 130° E and N 20 E forming a brick work pattern. The density of veins decreases away either side suggesting some type of fault activity. Beds dipping toward each other suggest a synclinal feature just around the bend of the creek.

Cummulative mi.	Miles from last Point.	
59.6	0.0	Leave the Town of Minden Court an d Town garage and turn right on to Rt. 80 going south
62.6	3.0	Stop at first outcrop just below the crest of the hill

Stop 4. Hallsville Road Outcrop

At this out crop we will examine the well laminated fissile gray to dark gray Upper Indian Castile. This lamination might be considered papery. This outcrop shows one tabular limestone bed that has a rusty weathering color outlining it. Walking further up the section we see just Upper Indian Castile with no tabular limestone beds.

Cummulative mi.	Miles from last Point.	
62.6	0.0	Continue south on Rt. 80 to the junction of Rt. 20
64.6	2.0	More Upper Indian Castile, fissile gray shale
75.1	10.5	Turn left onto Rt. 20 going east
81.2	6.1	Cherry Valley Limestone on the north side of Rt. 20
84.7	3.5	Turn right on to Otsego County Rt. 54 and park next to outcrop.

Stop 5. Otsego County Rt. 54 or old Chestnut street outcrop

At this classic outcrop we will see the lower part of the Chittenango now renamed to the Oakta creek (VerStraten, et al. 1994). Underlying the Oakta Creek is the Cherry Valley Limestone, and the uppermost section of Union Springs. Talus covers intensely deformed sections with the Union springs that exhibit a small thrust fault that ramps up through the Cherry Valley (Bosworth, 1984). The Union Springs member is the richest organic shale in New York, TOC range can be as high as sixteen percent.

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GEOLOGIC SETTING AND CHARACTERISTICS OF ADIRONDACK ANORTHOSITE AND RELATED MANGERITE-CHARNOCKITE-GRANITE (AMCG SUITE)

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The Adirondack Mountains represent a southwestern extension of the Grenville Province via the Thousand Islands – Frontenac Arch region of the St. Lawrence River (Fig. 1). The region is topographically divided into the Adirondack Highlands and Lowlands separated by the Carthage-Colton Mylonite Zone (CCMZ, Fig. 2). The former is underlain largely by orthogneisses metamorphosed to granulite facies and the latter by upper-amphibolite grade metasediments, notably marbles. Both sectors have experienced multiple deformations resulting in refolded major isoclines. Due to the complexity of the terrain, it was difficult to establish a reliable geologic history prior to the advent of U-Pb zircon geochronology. Silver (1969) pioneered a landmark study based on multigrain thermal ionization mass spectrometry (TIMS). His investigations indicated high-grade metamorphism at ca 1050 Ma and widespread intrusion of granitoids at ca 1120 Ma. These early analyses required very large zircon fractions (hundreds of mg.) and did not have the advantages of hand picking or air abrasion; hence, they represent mixed ages and serve only as rough guidelines. In the 1980's McLelland and Chiarenzelli initiated an investigation in which 30 representative samples were dated by multigrain, air-abraded TIMS methods (McLelland et al 1982, McLelland and Chiarenzelli, 1990). Subsequently, these results have been extended by single grain TIMS and SHRIMP techniques and most of these are summarized in McLelland et al. (1996), McLelland et al. (2001), Wasteneys et al (1999), and McLelland (2004) from which the first portion of this article has been derived. The picture that has emerged is one of a tripartite division of Adirondack geologic history, as summarized below.

The earliest recognized rocks in the Adirondacks consist of calcalkaline tonalites and granodiorites dated at ca 1350 – 1250 Ma and exposed in the southern and eastern Adirondack Highlands (McLelland and Chiarenzelli, 1990). Rocks of similar composition and age have been recognized in the Green Mountains of Vermont (Ratcliffe et al, 1996) and the Proterozoic of western Connecticut (Walsh and Aleinikoff, 2002). In addition, calcalkaline plutonic rocks ranging in age from 1270-1220 Ma are common within the western Central Metasedimentaery Belt (CMB) of the Canadian Grenville Province (Fig. 1). All of these calcalkaline orthogneisses have been interpreted as arc-related additions of juvenile crust to this sector of the Province during the interval ca 1400-1220 Ma (Fig. 3). Moore and Thompson (1980) proposed that this interval was characterized by repeated arc accretion to the southeast margin of Laurentia and referred to this protracted sequence of events as the Elzevirian Orogeny. Because Elzevirian plutons are overprinted by ca 1100-1000 Ma high-grade metamorphism, Moore and Thompson (1980) introduced the term Grenville Orogenic Cycle consisting of an early Elzevirian Orogeny and a younger (ca 1100-1000 Ma) Ottawan Orogeny that, together, replaced the old Grenville Orogeny and emphasized the existence of two major pulses of orogeny separated by an extensional interval of sedimentation represented by the Flinton Group in the CMB. In the Adirondacks, Elzevirian accretion appears to have extended to ca 1200-1160 Ma when the Adirondack Highland-Green Mountain Terrane collided with leading edge of the Frontenac Terrane of the CMB, which was at that time represented by the Adirondack Lowlands (Wasteneys et al 1999). This culminating accretionary event resulted in widespread magmatic and tectonothermal events that are referred to in adjacent Canada as the Shawinigan Orogeny (Corrigan and van Breemen, 1997, Rivers, 1987), and this terminology is adopted here.

Subsequent to the Shawinigan Orogeny, there does not appear to have been any significant regional deformation in the Adirondacks until ca 1090 Ma when, following a brief interval (ca 1103-1090 Ma) of granitic magmatism (Hawkeye suite, McLelland et al, 1996), the entire Grenville Province began to feel the effects of the Ottawan Orogeny that is thought to have been the result of collision with a continental-scale craton (Amazonia?) with the suture zone somewhere to the southeast of the present day Appalachians (Fig. 3). Geothermometry, geobarometry, and seismic investigations document that during the Ottawan Orogeny the Adirondack Highlands region obtained double crustal thickness and underwent granulite facies metamorphism and nappe emplacement (Bohlen et al, 1985, Valley et al, 1990, McLelland et al., 1996). The major contractional phase of the Ottawan Orogeny appears to have been over by ca 1050-1040 Ma, since minimally deformed subunits of the Lyon Mountain Granite were intruded at this time, especially the undeformed fayalite granites at Wanakena and Ausable Forks (McLelland et al, 2001). Locally, younger events of lesser magnitude continued to affect the region until ca 990 Ma and are recorded by metamorphic zircon and monazite.

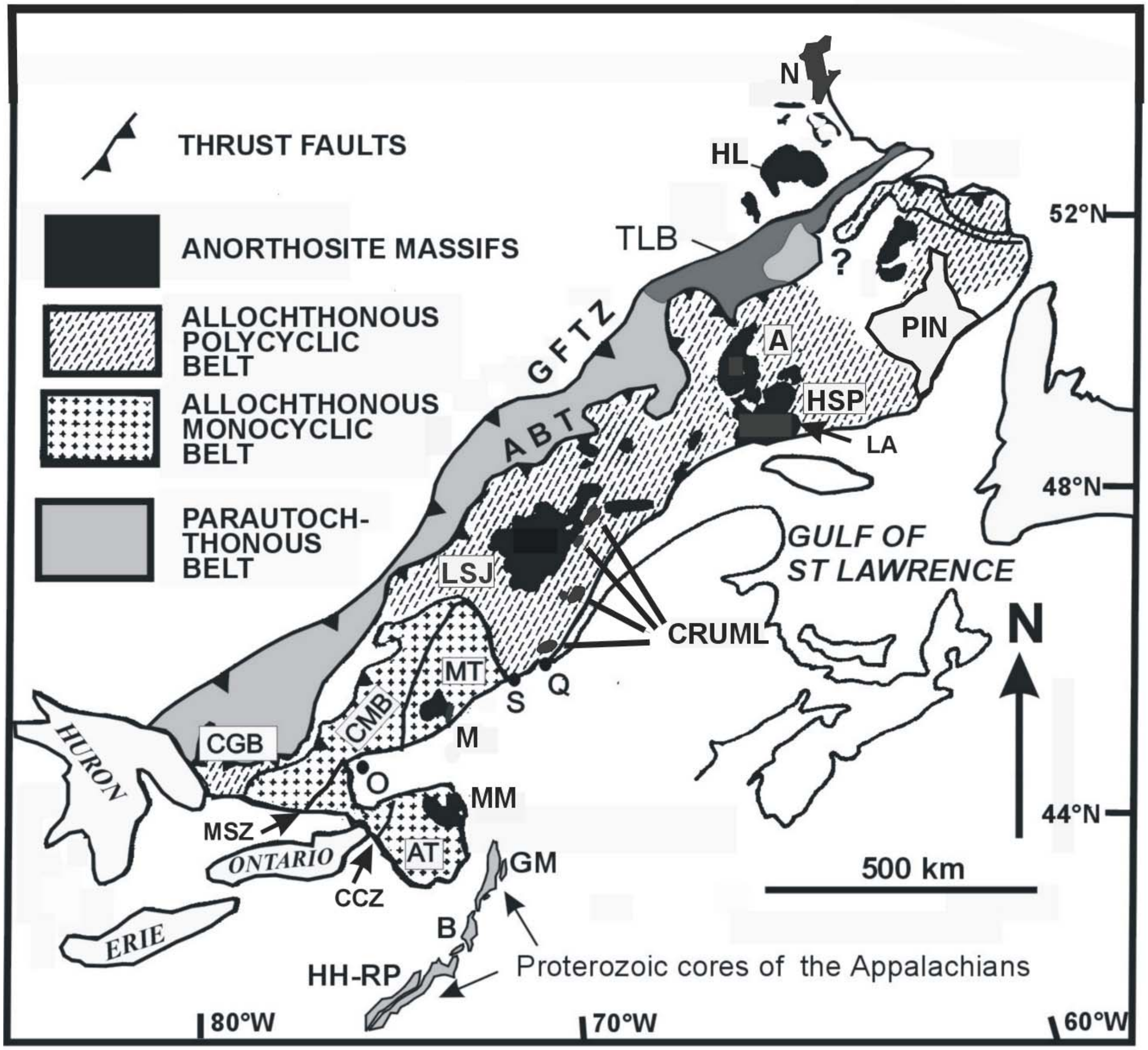


Fig. 1. Generalized location map of the Adirondacks as a southwestern extension of the Canadian Grenville Province whose three major tectonic divisions (Rivers, 1997) are shown. ABT-Allochthonous Boundary Thrust; GFTZ- Grenville Front Tectonic Zone; TLB (dark gray) - Trans-Labrador Batholith, GM – Green Mountains, H – Housatonic Mountains, HH-RP – Hudson Highlands and Reading Prong. The major anorthosite massifs (with ages) of the region are: MM- Marcy (ca 1150 Ma), M) Morin (ca 1160 Ma), LSJ) Lac St. Jean (ca 1150 Ma), HSP) Havre St-Pierre, A) Atikonak (ca 1130 Ma), ME), HL) Harp Lake (ca 1450 Ma), N) Nain-Kiglapait (ca 1300 Ma),; P) Pentecote (ca 1350 Ma). Age references: 1- 2, Hamilton et al. (2002), 3-7, Emslie and Hunt (1989), 4) Higgins and van Breemen (1996), 5) van Breemen and Higgins, (1993) Hamilton et al (1998), 9) Machado and Martignole (1988).

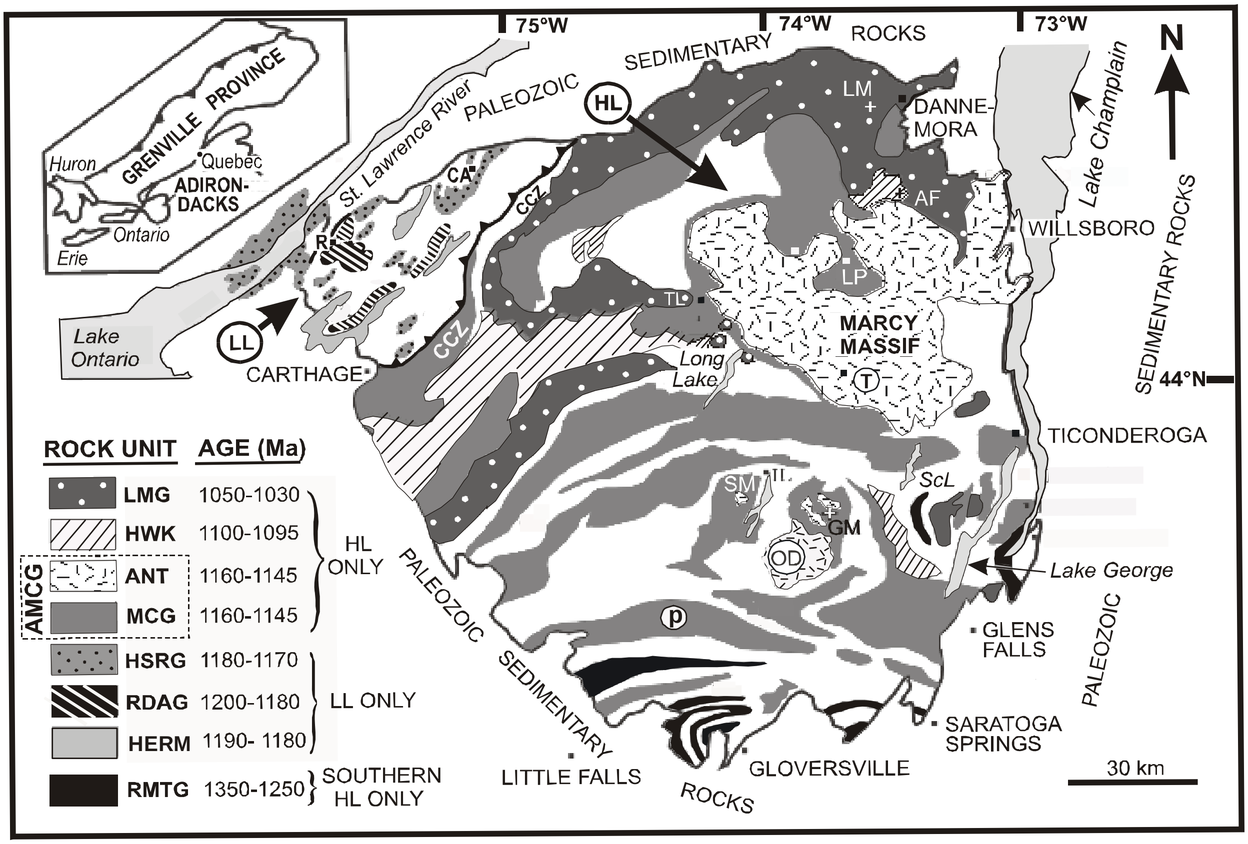


Fig. 2. Generalized geological/geochronological map of the Adirondacks. Units designated by patterns and initials consist of igneous rocks dated by U-Pb zircon geochronology with ages indicated. Units present only in the Highlands (HL) are: RMTG – Royal Mountain tonalite and granodiorite (southern HL only), HWK - Hawkeye granite, LMG – Lyon Mountain Granite and ANT - anorthosite. Units present in the Lowlands (LL) only are: HSRG – Hyde School and Rockport granites (Hyde School also contains tonalite), RANT – Rossie diorite and Antwerp granodiorite. Granitoid members of the AMCG suite (MCG) are present in both the Highlands and Lowlands. Unpatterned areas consist of metasediments, glacial cover, or undivided units. CCZ- Carthage-Colton Mylonite Zone, MM- Marcy anorthosite massif, OD Oregon Dome, SM – Snowy Mountain, LM – Lyon Mountain, IL – Indian Lake, SL – Saranac Lake, GO – Gouvernour, R – Rossie, CA – Canton. Locations for samples discussed in text: a)-Rooster Hill megacrystic charnockite, b)-Piseco leucogranitic ribbon gneiss, c)- Oregon Dome ferrodiorite, d)- Gore Mt. mangerite, e)-Snowy Mt mangerite, f) Schroon Lake granitic gneiss, g) -Minerva mangerite, h) North Hudson metagabbro, i) Woolen Mill gabbro and anorthosite, j)- anorthositic pegmatite and clinopyroxene-plagioclase dike in Ausable River at Jay, L) Yard Hill jotunite, m)- Bloomingdale mangerite, n) mangeritic dike cross-cutting anorthosite northeast of Tupper Lake Village, o)– mangerite southeast of Tupper Lake Village, p)- Rapakivi granite in Stark anticline, q) - Oswegatchie leucogranite r-Diana pyroxene syenite, s-Croghan granitic gneiss, t) Carthage anorthosite.

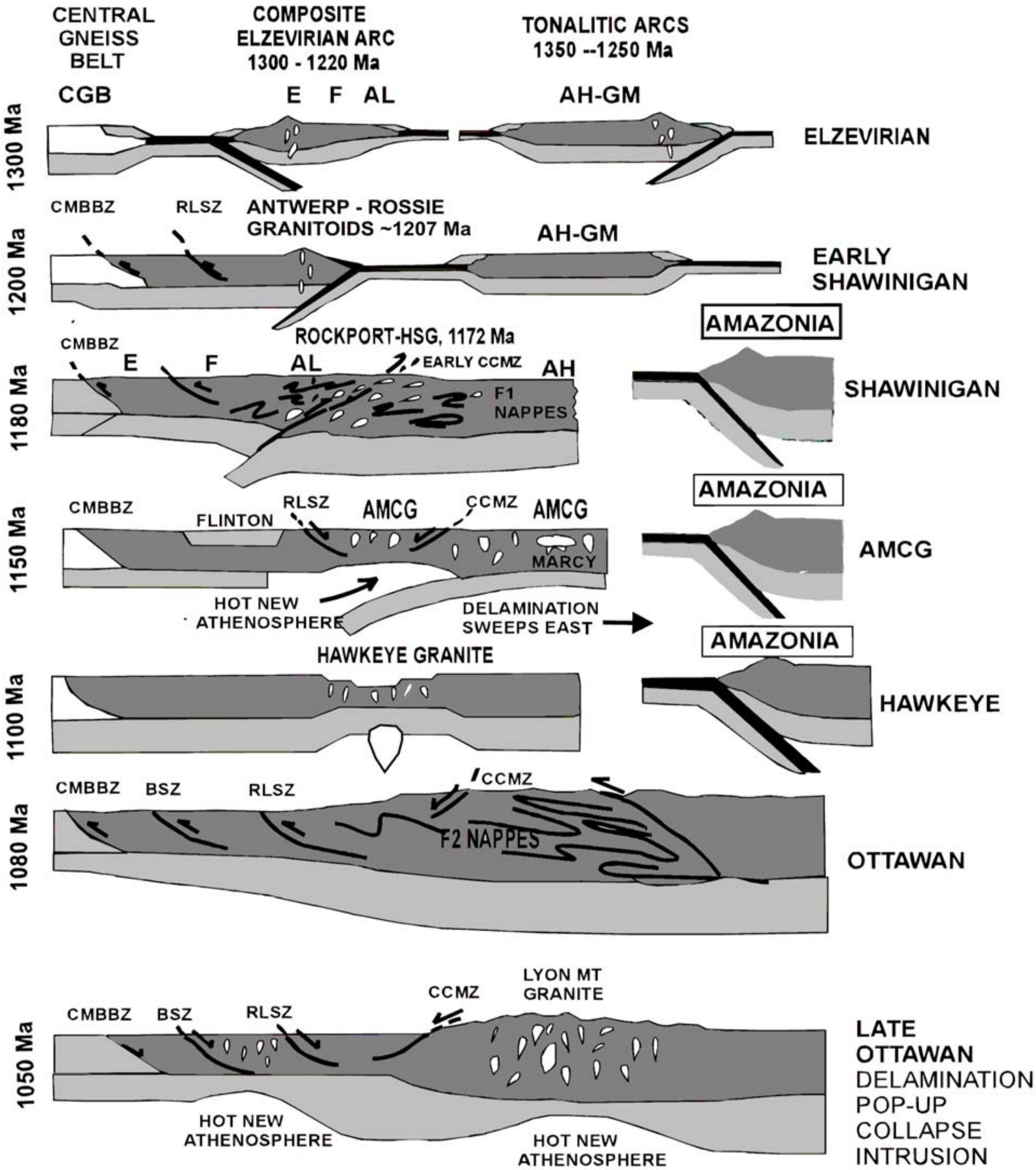


Fig. 3 Plate tectonic synthesis of the Adirondack region

For decades controversy has existed over how and when the Adirondack anorthosite massifs and associated granitoids are related to one another. Buddington (1939, 1969), Davis (1969), and Postel (1952) regarded the anorthosites as early, pre-orogenic intrusions that were slightly older than the AMCG granitoids that crosscut them. In contrast, deWaard and Romey (1969) and Lettney (1969) adopted the model of Martignole and Schrijver (1968) for the Morin anorthosite and proposed that Adirondack anorthosites were synorogenic intrusions and comagmatic with the AMCG granitoids. Silver (1969) emphasized that ca 1050 Ma metamorphic zircons in the
Marcy massif indicated that the anorthosite was older than this event and also predated the ca 1120 Ma AMCG granitoids that crosscut it. The U-Pb zircon results of McLelland and Chiarenzelli (1990a,b) confirmed Silver's interpretation by establishing the ca 1090-1050 Ma Ottawan Orogeny in the Adirondacks (See also Mezger et al, 1991) and documenting that the AMCG granitoids were emplaced at ca 1145 $\pm$ 15 Ma thus fixing a minimum age for the anorthosites that they crosscut. This placed AMCG magmatism subsequent to the Shawinigan Orogeny and ~50 Ma prior to the Ottawan Orogeny and was consistent with the ages of the large anorthosite massifs in the Canadian Grenville (Emslie and Hunt, 1992).

The eastern Grenville Province contains a number of mid-Proterozoic AMCG complexes cored by large anorthosite massifs (Fig. 1) accounting for ~20% of the region's area (McLelland, 1989). Within the Adirondacks, the Marcy anorthosite massif (Figs. 1,2) has long been regarded as typical of mid-Proterozoic massifs worldwide (Ashwal 1993). The smaller Oregon Dome and Snowy Mt bodies (Fig. 2) are compositionally and mineralogically similar to the Marcy massif. Several other small occurrences exist on a scale too small to be shown on figure 2. Adirondack anorthosites are of the andesine-labradorite (An_{45}-An_{55}) variety and contain magnetite and ilmenite reflecting a relatively low fO_2 (Anderson 1987) and resembling other large massifs in the Grenville Province. In addition to these low fO_2 massifs, several smaller and distinctive alkalic and hemoilmenite-bearing andesine ($An_{30}Or_{11}$-$An_{40}Or_7$) anorthosites occur in Quebec (Owens and Dymek, 1999) and are also present in the Proterozoic core of the southern Appalachians (Owens and Samson, 2001). These late- to post-tectonic bodies are clearly more oxidized and younger (ca 1050-1010 Ma) than the larger, more reduced massifs, such as the Marcy, that form the subject of this paper.

The Marcy massif is roughly elliptical (Fig. 2) with its long dimension (NW-SE) extending ~100 km and its short (NE-SW) dimension measuring ~60 km. Gravity models suggest a slab-like shape some 3-5 km thick with two funnel-shaped feeder pipes extending to ~10 km. Like other igneous bodies of this size, the Marcy massif is a composite intrusion consisting of a variety of anorthositic and leucogabbroic members that were repeatedly emplaced in an unspecified number of pulses. Following Buddington (1939, 1969), the principle rock types comprising the massif are anorthosite (>90% plagioclase), mafic anorthosite (90- 75% plagioclase), leucogabbro (75 – 65% plagioclase), and gabbro (<65% plagioclase). Together, anorthosite and mafic anorthosite constitute ~90% of the massif. Leucogabbro is concentrated along the borders of the massif where it occurs together with mafic anorthosite to form a deformed, finer-grained, and more mafic border facies surrounding a largely anorthositic core of coarse, little-deformed blue-gray plagioclase. The leucogabbro is commonly referred to as the Whiteface facies (Kemp, 1898), whereas the anorthositic varieties are known as the Marcy facies (Miller, 1919). Modern workers continue to use these designations but with the recognition that two facies are insufficient to describe the complicated nature of the massif and its multiple intrusive pulses. In addition, the margins of the massif commonly contain a complex border facies consisting of rocks that represent commingling of anorthositic and granitoid magmas. Miller (1919) named these hybrid rocks Keene Gneiss for their exposures near the village of Keene. The mixtures can become very complex, because a variety of AMCG magmas, including ferrodiorites, are involved. Within Keene Gneiss, plagioclase xenocrysts incorporated into granitoids or alkali feldspar incorporated into anorthosite, exhibit reaction rims of perthite or plagioclase, respectively (McLelland et al 2002). In addition to the major massif rock types described above, there exists a widespread, small volume (~5%) member of the suite that is best categorized as ferrodiorite or ferrogabbro (McLelland et al. 1994). These very mafic rocks are concentrated near the massif margins where they occur as sheets and dikes crosscutting the anorthosite. They are also present in small volume throughout the massif as thin, wispy and discontinuous veinlets within the anorthosite. McLelland et al (1994) interpreted these titanium-, iron-rich and silica-poor rocks as late differentiates of the anorthosite that were injected by filter pressing. They also proposed that further fractionation of the mafic magma would lead to formation of the ilmenite-magnetite deposits associated with the Marcy massif.

A striking feature of the Marcy facies is the presence within it of rafts of exceptionally coarse anorthosite containing plagioclase crystals with long dimensions of 15-30 cm and up to 45 cm in length. These are found in subophitic intergrowths with both ortho- and clinopyroxene, some of which qualify as "giant pyroxenes" (Fig. X). Emslie (1975) analyzed some of these and found high concentrations of Al_2O_3 (10-15%) indicative of high-pressure origins. In almost all occurrences these ultra-coarse assemblages are in the form of rafts surrounded by finer grained varieties of leucogabbro or mafic anorthosite   In addition to the exceptionally coarse rafts of anorthosite, there exist large tracts of coarse (4-10 cm) anorthosite consisting of blue gray plagioclase and, 10% mafics. Much of this material exhibits flow alignment of plagioclase laths and must represent mobilized crystal-rich mush. The alignment is interpreted as magmatic rather than tectonic, because neither the plagioclases nor interstitial subophitic pyroxene are deformed. Judging by the high (>90%) concentration of plagioclase laths in these rocks it is probable that they represent cumulates that formed within the post-emplacement massif. The parental magma of these cumulates is thought to be similar in composition to the mafic anorthosite and leucogabbro that enclose the coarse grained rafts and also constitute the Whiteface facies (Buddington, 1939, 1969, McLelland et al, 1994).

DISCUSSION

The results of SHRIMP geochronology of the Adirondack AMCG suite (Fig. 8), together with those of Hamilton et al (1994), are shown graphically in Fig. 7 a, b. These two plots emphasize the narrow range of ages for both the anorthositic rocks and their associated granitoids. This close correspondence is further documented by weighted average age plots for the two groups, i.e., anorthosites, 1155 $\pm$ 6 Ma and granitoids, 1158 $\pm$ 5 Ma (Fig X2, A, B). Additionally, further refinement of the granitoid data reveals that the four granitic members (Fig YY A) yield a weighted average age of 1148 $\pm$ 6 Ma (MSWD = .4, Probability = .93), whereas 7 samples of mangerite and charnockite yield a weighted average age of 1163 $\pm$ 7 Ma (MSWD = .2, Probability = .93). The latter age may be especially relevant to models of AMCG genesis, because the less evolved charnockites and mangerites probably represent the earliest members of the group. All of these results are discussed below.

Direct dating of the Adirondack AMCG suite demonstrates that all members are coeval within the interval 1155 $\pm$ 10 Ma. This now-established result serves as a powerful constraint on genetic models, both within the Adirondacks and elsewhere in the world. Documentation of the contemparaneity of suite members permits the age of undated anorthosites to be inferred from the ages of associated AMCG granitoids. Thus, for example, the ages of AMCG granitoids dated by Emslie and Hunt (1990) in the Canadian Grenville can now be assigned with confidence to their associated anorthosite massifs, as originally suggested by these authors. Such documentation of absolute age relationships greatly facilitates the crucial goal of constraining and establishing the origin, evolution, and tectonic setting of AMCG complexes. In the following, we utilize the coeval nature of the suite to help construct a self-consistent tectonic-petrologic model of AMCG genesis. Many of the specifics are not new, but their synthesis hinges upon the coeval nature of the suite.

The tectonic setting of AMCG suites has been debated for decades. Emslie (1978), Anderson (1983), and McLelland et al (1996), among others, proposed an “anorogenic” environment for these rocks. It was argued that anorthosites represent derivatives from gabbroic magmas, and the best way to produce these derivatives is to pond gabbroic magma under the density lid at the crust-mantle interface, where they can evolve under quiescent conditions and accumulate plagioclase-rich crystal mushes, fractionate olivine and pyroxene, and decrease their density while they also partially melt the lower crust to produce granitoid magmas (Emslie, 1975, 1978; Ashwal, 1993). In contrast, in an environment of rapid rifting the ponded gabbros would quickly ascend and form basalt flows, new oceanic crust, etc. If, instead, the environment were one of contraction, magmas from diverse sources would become hybridized and/or injected upward prior to fractionation. Accordingly, active tectonic regimes were thought to preclude quiescent environments necessary for fractionation and the accumulation of a plagioclase-rich crystal mush. In contrast, so-called “anorogenic” environments were perceived as capable of providing conditions for quiescent fractionation and genesis of anorthositic magmas and AMCG suites. In the “anorogenic” model, the gabbro’s most likely source would be a mantle plume or hotspot. The “anorogenic” model was also consistent with the observation that, while many AMCG suites occur within tectonic belts, they do not appear to be syntectonic, e.g., Nain Anorthosite Complex.

Although a number of investigators expressed skepticism about “anorogenic” environments (cf., Ashwal, 1993), the most demonstrable flaw in the concept arose from the expansion of U-Pb zircon geochronology in the Grenville Province, including the Adirondacks. By the early 1990’s it became evident that emplacement ages of AMCG granitoids overlapped in time with contractional events taking place within the same general region (Corrigan and Hanmer, 1997, McLelland et al., 1996). Accordingly, the concept of “anorogenic” emplacement was no longer tenable in the region. At the same time, the role of delamination in overthickened, contractional orogens was becoming better understood, especially as it applies to late and post-orogenic magmatism (Sandiford and Turner, 1991). Ultimately, both Hanmer and Corrigan (1997) and McLelland et al (1996) applied the delamination mechanism to the genesis of AMCG suites, especially within the late stages of contractional orogenesis. They argued that once a dense lithospheric keel is delaminated from a thickened orogen, hot replacement athenosphere rises to the crust-mantle interface and produces gabbroic magma. Simultaneously, the orogenic crust, deprived of its high-density root, experiences strong buoyancy forces that result in uplift and the neutralization of horizontal contractional forces that may continue to exist. Orogen collapse along low angle normal faults may take place, but the buoyant, tectonically “neutral”, orogen can stay relatively “quiescent” for a substantial period of time. *This set of circumstances results in an orogenic environment that provides all of the stability inherent in supposed “anorogenic” settings but is far superior to the latter, because the concept of a delaminated, buoyant orogen is more consistent with both theory and observables.* An important corollary of the delamination-based model is that it places the genesis of AMCG suites within time intervals that coincide with late- to post-tectonic magmatism, and this is consistent with observations in the Adirondacks and adjacent parts of the Canadian Grenville. Specifically, the voluminous AMCG magmatism dated at ca 1150 Ma (eg, Marcy, Morin, Lac St-Jean, Atikonak massifs, fig 1) immediately follows the Shawinigan Orogeny that affected this portion of the Grenville Province during the interval ca. 1200-1150 Ma (McLelland et al, 1996, Corrigan and Hanmer, 1997, Wasteneys et al, 1999). Similar associations may hold for the ca 1300-1450 Ma Harp Lake, Mistastin, Michikamau and Nain AMCG suites where there now exists compelling evidence that these complexes were associated with funneled, flat-slab subduction in the interval 1460-1230 Ma (Gower and Krogh, 2002). The underiding slab may have involved a spreading center that provided plumbing for ascending mantle melts. Alternatively, instabilities and rollback in the slab may have led to episodes of delamination allowing hot, new athenosphere access to the base of the crust so as to produce AMCG magmatism. On a much smaller scale, the Labrieville, St. Urbain, Roseland, and Montpelier anorthosites fall into the range ca 1050-1000 Ma coeval with late- to post-tectonic events associated with the Ottawan Orogeny (Owens and Dymek,) and its proposed delamination (McLelland et al, 2004).

There currently exist two major petrogenetic models put forward to account for AMCG suite genesis. The first is based upon bimodal magmatism and considers the suite to be coeval but not comagmatic. It has been described in detail by Emslie (1985), Emslie et al, (1994), Ashwal (1993), Weibe (1994), and Fram and Longhi (1992) and represents the most commonly held view of AMCG magmatism. Its principal characteristics are that: a) anorthosites are derived from mantle melts and b) their associated granitoids are the result of melting of the lower crust by heat derived from elevated isotherms and the mantle melts cited in a). The second model utilizes experimental and isotopic evidence to argue for a comagmatic origin for most AMCG suites, and is characterized by a mafic (two pyroxenes and plagioclase) lower crustal (P~ 10-13 kb) source region that melts to produce jotunitic compositions similar to high-aluminum gabbro but containing higher concentrations of Ti, P, and K (Vander Auwera et al, 1998; Longhi et al, 1999). Fractionation of the jotunitic magmas is thought to produce the entire range of AMCG compositions. Both models a) and b) are consistent with a strictly coeval origin for these rocks; however, only the first predicts that granitoids should predate anorthositic members by a narrow time interval. This difference in timing was first noted by Emslie et al (1994) who pointed out that heating of the lower crust by ponded gabbro would lead to early production of granitoid melts leaving behind hot, residual plagioclase-pyroxene restites that could be assimilated by the already fractionating gabbroic magmas. Assimilation-fractional crystallization (AFC) processes would lead to anorthositic magmas characterized by geochemical signatures consistent with lower crustal trace element patterns.

Our strong preference for a coeval but not comagmatic origin for the Adirondack AMCG suite is based in part on observables such as the zones of Keene Gneiss that separate AMCG anorthosite and granitoids and represent disequilibria dominated, physical mixtures between the two magmas. Within these zones, disequilibria is manifested by the reaction rims of perthite around anorthosite andesine and andesine-oligoclase around granitoid perthite (Hamilton et al, 2004). Davis (1968) and Buddington (1968) both drew attention to the absence of correlation between pyroxene (mg#) and plagioclase compositions (An%) within Keene Gneiss, whereas they consistently correlate within the anorthosites and granitoids themselves. These same authors also noted that screens of country rock commonly separate anorthosite from granitoids and imply intrusive relationships. Consistent with this is the observation that ferrodiorites (jotunites) are common in anorthositic rocks but are rare, and of different composition, in associated granitoids. Moreover, the thin veinlets of ferrodioritic material that occur throughout the anorthosites are clearly of local origin and can be seen to occupy small fractures and shear zones just as would be expected from filter pressing. In short, Adirondack ferrodiorite (jotunite) is a late derivative from anorthositic magmas and, contrary to Vander Auwera et al (1998) and Longhi et al, 1999) did not serve as a parental magma for either the anorthosite or the AMCG granitoids. Indeed, the enormous volume of AMCG granitoids in the Adirondacks inveighs against a derivation by fractionation but is wholly consistent with their production as deep crustal melts resulting from mantle-derived heat transported by mantle-derived gabbros whose fractionation at the crust-mantle interface produced ample heat of crystallization resulting in a bimodal suite of granitoid crustal melts and plagioclase crystal mush. Vander Auwera et al (1998) cite de Waard and Romey (1968) in order to claim that “parts” of the Adirondack AMCG suite are examples of fractionation that produces all members of the AMCG as consanguineous derivatives of a single parental magma. The seminal article (de Waard, 1970) setting forth this proposition is based on exposures representing the entire AMCG suite exposed in the bed of Roaring Brook on Giant Mountain in the eastern Marcy massif. McLelland (1992) reinvestigated this locality in great detail and has shown that the relationships present are best accounted for as the result of magma mixing, commingling, and hybridization. Indeed, enclaves originally misinterpreted as xenoliths of Grenville metasediments (Jaffe et al, 1983) in the granitoids can be shown to be the result of commingling between gabbroic and syenitic magmas. At Snowy Mt. (Fig. 2), where deWaard and Romey (1968) collected data that led to their comagmatic proposal, dating reported by Hamilton et al (2004) shows that charnockitic rocks dated at 1174 ± 25 Ma surround a ca 1155 Ma anorthositic core and comagmatism between the two is unlikely. Within the Adirondacks, a supportable case for a comagmatic AMCG suite has yet to be made.

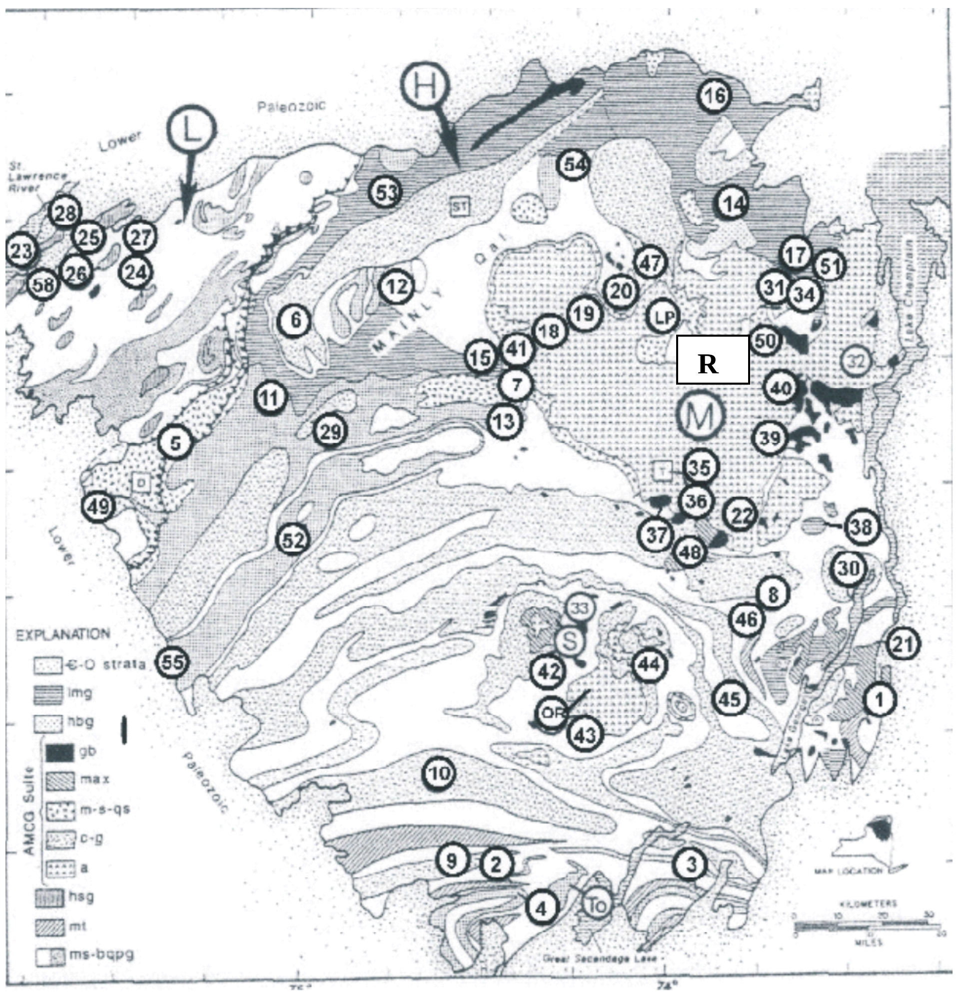


Fig. 4. Locations of samples dated by zircon and listed in Table 1

Table 1

Table 1. Summary of U-Pb Zircon Geochronology For Adirondack Metaigneous Rocks

Map Number	Sample Number	Location	Multigrain TIMS AGE (Ma)	ERR	Singlegrain TIMS AGE (Ma)	ERR	SHRIMP II Analysis AGE (Ma)	ERR	TDM
HIGHLANDS									
Tonalite and Granodiorite									
1	AM87-12	South Bay	1329	37					1403
2	AM86-12	Canada Lake	1302	6					1366
3	LDT	Lake Desolation	>1336						1380
4	AM87-13	Canada Lake	1253	41					
Mangerite and Charnockite									
5	AM86-2	Diana Complex	1155	4			1154	17	1430
6	AM86-15	Stark Complex	1147	10					1495
7	AM85-6	Tupper Lake	1134	4			1169	11	1345
8	9-23-85-7	Schroon Lake	1125	10			ca 1155		
9	AM86-17	Rooster Hill	1156	8					1436
10	AM86-9	Piseco Dome	1150	5					1346
11	AC85-2	Oswegatchie	1146	5					
30	Silver,'69	Ticonderoga	1113	16			ca 1155		
42	AM86-8	Snowy Mt	>1095				1177	22	
44	AM87-3	Gore Mt	>1088				1155	6	
47	AC85-10	Bloomingdale	1133	51			1160	14	
48	AM87-10	Minerva	>1082				1159	12	
49	AM86-1	Croghan	1155	13					
41	Granitedike	Wabeek Quarry					ca 1155		
50	AC85-11	Yard Hill	1143	33					
Anorthosite and Olivine Gabbro									
18	AC85-8	Rt 3, Saranac Lk, ANT	>1113, 1054	22			1149	35	
19	AC857	Rt 3, Saranac Lk, ANT	>1087, 1052	20			1161	12	
20	AC85-9	Forest Home Rd, ILM	996	6					
21	AM87-11	Dresden Station Gab.	1147	7					1331
22	CGAB	North Hudson Gabbro	>1109, 1057	conc			1150	14	
31	BMH01-4	Jay, ANT Pegmatite					1160	15	
34	BMH-01-3	Jay, Cpx-Pgf Dike					1140	18	
35	BMH01-1	Tahawus ANT					ca 1155		
36	BMH01-2	Blue Ridge ANT					1153	11	
37	BMH01-1	Blue Ridge Gabbro					ca 1155		
39	BMH01-19	Exit 29 NWY, ANT					ca 1155		
40a		Woolen Mill Gabbro					1154	9	
40b		Woolen Mill ANT2					1151	6	
43	AM87-8	Oregon Dome Fer'd'r't					1155	6	
Hawkeye Granite Suite									
12	AM86-3	Carry Falls	1100	12					
13	AM86-6	Tupper Lake	1098	4					1314
14	AM86-13	Hawkeye	1093	11					
45	Moon Mt.	Moon Mt	1103	15					
52	NOFO-1	Stillwater Reservoir	1095	5					
53	AM87-6	St. Law/Fran. Co. Line	1090	6					
54	AM87-7	Santa Clara	1080	4					
Lyon Mt Granite									
15	AM86-4	Piercefield	1075	17			1058	18	1576
16	AM86-10	Dannemora	1073	6			1052	11	
17	AM86-14	Ausable Forks, Qt-Ab	1057	10			1041	16	1350
29	CLFG	Wanakena	1113	10	1069	10	1047	10	
38	9-23-85-6	Grasshopper Hill	>1065		1049	3			
51	AM86-11	Ausable Fks., Fay GRT	1089	26	1047	2			
55	PL-3	Port Leyden			1035	4			
LOWLANDS	Selleck	CCMZ			1047	10			
Hyde School Gneiss									
23	AM86-16	Wellesley Island	1416		1172	5			1440
24	AC85-4	Gouverneur	1284						1525
25	AC87-4	Fish Creek	1236		1172	5			1210
26	AC85-5	Hyde School	1230		1172	5			1360
27	AC85-1	Reservoir Hill	ca 1172						
Antwerp Granitoid									
58	ANTG	Antwerp-Rossie	1183	7			1207	20	

ROAD LOG

0 Gathering Point: along west-bound shoulder Rt. 73 at intersection with I-87 (Exit 30).

2.3 Proceed NW along Rt. 73 to junction with Rt. 9N. Bear left on Rt. 9N to Lake Placid

5.6 Turn sharply right into the parking lot for Roaring Brook trail

Stop 1 *Roaring Brook on Giant Mt.* (located near R, Fig. 4)

The valley of Roaring Brook on the southwestern slopes of Giant Mt provides some of the finest exposures of AMCG rocks in the Adirondacks. Over a distance of ~ 1.5 km and a width of 50-100 m there occur an uncommon array of dikes, sheets and plutons of several generations and compositions that have intruded approximately parallel to the stream valley. A plausible interpretation of the association is that it represents a zone of weakness that repeatedly served as a magma conduit during the emplacement of the ca 1155 Ma AMCG suite.

From 1400' to 2000' the brook is underlain by a variety of anorthositic and gabbroic rocks. Near the lip of the high bridal veils falls the valley is defined by erosion of a diabase dike in the anorthosite, Slightly farther upstream z broad (~3m) irregular pyroxene-monzonite dike outcrops on the northwest bank, and clearly truncates magmatic flow foliation in the anorthosite. Coarse comb texture is visible along the dike's margins.

Proceed upstream past several plunge pools to a steep 3-10 m cliff face. A dark, meter-wide eroded and irregular dike of orthopyroxenite is exposed at the base of the waterfall. Erosion of the dike has resulted in the stream channel and can be followed 30-40 m upstream from the cliff edge. Orthopyroxene greatly dominates the mode (90%), but some clinopyroxene, magnetite-ilmenite, and plagioclase are also present. The orthopyroxene appear to be of orthocumulate origin with narrow adcumulate overgrowths. We speculate that the origin of the dike is the result of drainage of orthopyroxene cumulates into a fracture that opened at the base of magma chamber. Several xenoliths of anorthosite are present in the dike, and locally it exhibits soft margins with the country rock anorthosite. The Mg-numbers of the pyroxenes are high (Cpx-75, Opx-65) and suggest that the dike formed early in the fractionation history of the magma.

The smooth outcrop surfaces surrounding the orthopyroxenite dike are dominated by gabbroic anorthosite transitional to gabbro. Several stages in the magmatic history of the anorthosite are recorded by crosscutting relationships. The oldest anorthosite facies are coarse-grained rafts of blue-gray andesine-labradorite anorthosite corresponding to the Marcy facies. They occur as xenoliths within as xenoliths within a medium grained, subophitic, two-pyroxene gabbro that locally grades into norite. This, in turn, is crosscut by a fine-grained gabbroic anorthosite similar to the Whiteface facies. Elsewhere in the outcrop the sequence may be reversed and attests to multiple pulses of magma; however, the coarse grained rafts are always the oldest.

Returning to the trail and proceeding uphill, we cross Roaring Brook and ascend the summit trail to the 2260' (689m) level. Here we leave the trail and scramble downhill to water-smoothed pavement outcrops in the brook valley. The outcrops expose a spectacular intrusion breccia consisting of rounded and angular blocks (10-30 cm on average) that include coarse, white anorthosite and glassy, pink garnetiferous anorthosite, but consist mostly of gray to black, medium to fine-grained, granular pyroxene-plagioclase assemblages. These are set in a medium grained groundmass ranging from gabbroic anorthosite to garnetiferous mangerite and ferrogabbro. These matrix rocks are highly mingled and difficult to separate without the aid of staining.

A number of dark inclusions exhibit narrow (~0.5 cm), light colored layers. Traditionally these were interpreted as primary in origin and associated with a sedimentary or volcanoclastic origin (Kemp, 1921, deWaard, 1970, Mclelland et al, 1982). However, examination of stained slabs and thin sections reveals that the light layers are of syenitic composition whereas the dark layers consist of gabbro. It is common for the light, syenitic layers to be continuous with host rocks of the same composition. Moreover, the gabbroic layers contain acicular rods of orthopyroxene that, together with plagioclase, define a "fence-post" pattern (Fig. 5). The pyroxene-plagioclase layers are essentially gabbroic to dioritic in composition with subequal amounts of magnesian pyroxene and intermediate plagioclase. On the basis of chemistry and textures, we interpret the layered inclusions as igneous enclaves intruded by parallel veins of syenite and mangerite and incorporated into the mixtures of country rock magmas now constituting the breccia groundmass. The “fence-post” textures are identical to comb-textured pyroxenes in orbicular granites (McKinney, 1990) and provide compelling evidence for their igneous origin (Fig. 5).

The outcrops at Roaring Brook also contain numerous non-layered inclusions, and most of these are rich in magnesian clinopyroxene. They are interpreted as disrupted cumulates caught up by ascending AMCG magmas. None of these enclaves contain acicular pyroxene and exhibit coarse, interlocking grains. They are commonly associated with dark gray, relatively fine grained material of approximately dioritic composition and are thought to be cumulates from these magmas.

A number of the inclusions at Roaring Brook exhibit soft, lobate boundaries typical of magma commingling. Some of the anorthositic inclusions have especially well-developed borders of this sort. These features are attributed to the interaction and incomplete mixing of magmas ascending in the Roaring Brook conduit.

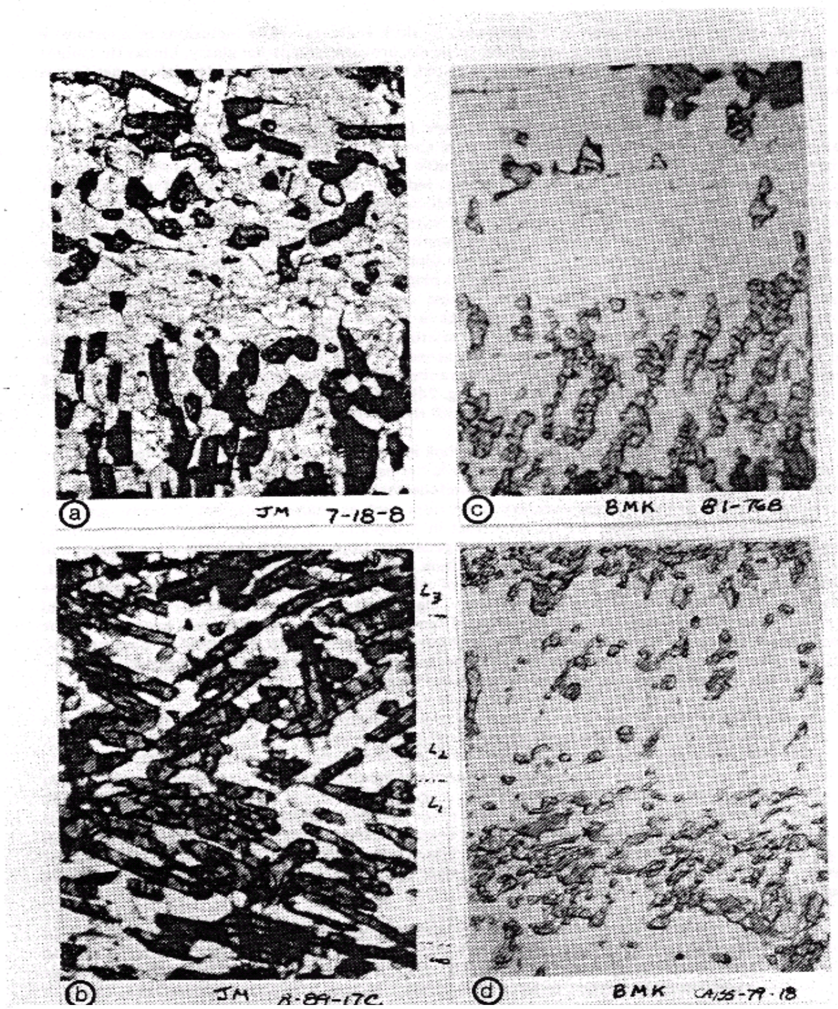


Fig. 5. Comb textures formed by acicular orthopyroxene in enclaves at Roaring Brook (a & b) compared with known com textures with acicular opx (Sierra Nevada plutons) studied by Brooks McKinney.

Table 2

TABLE 6
Major Element Analyses of Roaring Brook Samples

Sample	1 Marcy-type Anorthosite[a]	2 Whitface Anortosite[a]	3 Anorthosite 7-19-89-27	4 Leuconorite 7-19-89-20	5 Fine Grained Gabbroic Anorthosite 7-19-89-21	6 7-19-89-22	7 Gabbro 7-19-89-23	8 Small Mafic Inclusions 7-19-89-25	9 Orthopyroxenite Dike	10 Jotunite
SiO_2	54.54	53.54	54.97	52.55	53.75	53.70	53.19	53.29	50.82	47.16
TiO_2	0.67	0.72	.30	.49	.23	.21	.43	.30	1.27	2.20
Al_2O_3	25.61	22.50	25.09	21.48	25.23	26.54	21.66	23.24	4.7	17.23
Fe_2O_3	1.00	1.26	1.05	1.13	2.18	1.05	4.07	3.31	2.3	2.75
FeO	1.26	4.14	1.19	3.01	---	---	---	---	14.2	9.24
MnO	0.02	0.07	.02	.06	.02	.01	.06	.04	0.29	0.15
MgO	1.03	2.21	.68	3.96	1.84	.71	3.9	2.76	16.52	2.71
CaO	9.92	10.12	9.95	11.79	10.93	10.40	11.43	11.82	8.44	9.04
Na_2O	4.53	3.70	4.7	3.84	4.26	4.63	3.78	3.80	0.71	6.61
K_2O	1.01	1.19	1.46	.66	.75	.80	.65	.71	0.20	2.27
P_2O_5	0.09	0.13	.07	.09	.06	.06	.09	.04	0.14	0.59
H_2O	0.55	0.12	.57	.48	.67	.49	.38	.53	0	0
Total	100.17	100.00	99.86	99.95	99.95	98.57	99.63	99.82	99.59	99.70

Sample	11 Mafic Mangerite	12 Mangerite	13 Charnockite	14 Garnet-Plagioclase Xenolith 7-18-89-3	15 7-18-89-2	16 Cumulates 7-18-89-7	17 7-18-89-21A	18 Intermediates 7-18-89-21B	19 Enclaves 7-18-89-4	20 7-18-89-6
SiO_2	50.05	56.45	62.70	49.23	55.74	52.90	52.75	53.83	57.62	55.28
TiO_2	1.47	1.50	0.44	2.63	.17	.70	.65	.84	.73	.68
Al_2O_3	16.08	15.88	18.41	16.76	2.87	5.50	10.17	12.96	13.54	12.54
Fe_2O_3	2.53	2.60	0.63	2.73	5.22	13.37	8.78	11.40	6.76	7.73
FeO	9.06	8.13	2.37	12.68	---	---	---	---	---	---
MnO	0.20	0.17	0.05	.43	.83	.38	.24	.25	.11	.13
MgO	3.21	1.06	0.40	2.73	14.39	8.18	7.61	4.83	5.10	5.56
CaO	8.20	4.39	2.49	7.01	18.51	15.21	14.38	10.66	8.80	10.06
Na_2O	4.3	3.56	3.85	2.54	.62	1.05	3.00	3.37	2.87	3.03
K_2O	4.33	5.50	7.59	2.51	.9	1.2	.88	1.34	4.16	3.01
P_2O_5	0.57	0.38	0.15	.77	.05	.17	.15	.23	.16	.17
H_2O	0	Nd	Nd	.01	.10	.10	.38	.30		.70
Total	100.09	99.62	00.08	100.02	100.21	98.82	98.99	100.01	99.85	98.89

[a]Buddington (1939)

Table 3

	Split Rock Falls Gab. Anorthosite Roadcut	Rt. 9 Elizabethtown Anorthositic Gabbro Gabbroic Layers	Felsic Veining Block Structure Woolen Mill	Woolen Mill Gabbro In River
SiO2	51.20	41.36	60.36	45.60
TiO2	2.61	4.21	1.07	3.49
Al2O3	17.67	15.96	18.26	14.23
Fe2O3	10.43	16.43	6.62	18.42
MnO	.15	.30	.08	.27
MgO	2.53	4.91	1.06	3.07
CaO	9.37	12.83	6.45	9.25
Na2O	3.10	1.90	2.67	2.63
K2O	1.72	.16	2.71	.79
P2O5	1.02	1.67	.16	1.26
H2O	.27	0	.55	.70
Total	100.06	99.74	100.00	100.23

Turn left (southeast) out of parking lot onto Rt. 73

5.6 10.3 Intersection of Rt 73 & Rt. 9. Turn left (north on Rt. 9 towards Elizabethtown.

2.4 15.9 Turn right into parking area at Split Falls.

Stop 2 *Split Rock Falls* (located near 40, Fig. 5)

The roadcut across from the parking area provides good evidence for multiple intrusions of anorthositic and gabbroic rock. The dominant rock type is gabbroic anorthosite that encloses numerous xenoliths of the anorthositic suite. Subophitic textures are preserved in some of the more gabbroic xenoliths. Garnetiferous gabbro truncates foliation in some xenoliths and has, itself, a different foliations. It is thought that many of the foliations are magmatic in origin. A late mafic facies similar to that at Woolen Mill disrupts all earlier rock types. Late mafic dikes (Phanerozoic?) with well-developed slickensides cut all other lithologies.

From parking area turn right (north) on Rt. 9 to Elizabethtown.
7.7 23.6 Long outcrop on west side of Rt. 9

Stop 3 *Elizabethtown anorthositic gabbro*

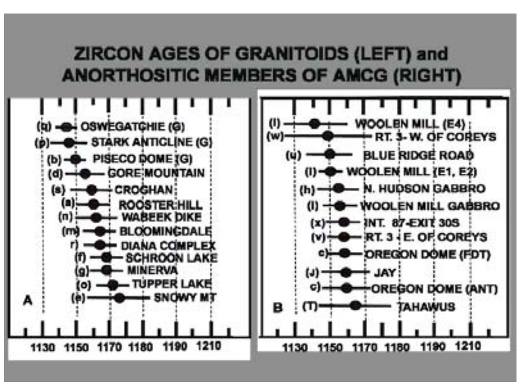
Although most of the outcrop is highly altered, it affords the opportunity to see the effects of Ottawan deformation on AMCG rocks that are situated near the outer margins of the Marcy Massif. Most of the outcrop consists of somewhat gneissic anorthositic gabbro. Some small calcsilicate inclusions consist of grossularite, diopside, and wollastonite. Large cm-scale garnets appear to truncate foliation and may have grown under static conditions. Large black clots of hornblende contain remnant pyroxene cores and may represent deformed, metamorphosed giant pyroxenes.

1.3 24.9 Continue north on Rt. 9 until junction with Rt. 9N. Turn left (west) on Rt. 9N
1.1 26 Pull off into dirt parking area on right and across from large, dark roadcut.
Stop 4 *Woolen Mills Anorthosite and Ferrogabbro* (located near 40, Fig.5)

The large roadcut shows anorthosite intruded by a dark, reddish-brown, fine-grained rock known as the Woolen Mill Gabbro (1154 ± 9 Ma) for the mill that once existed in the stream across the road. It is a clinopyroxene-garnet-oligoclase granulite with considerable opaque oxides and apatite and minor K-feldspar and quartz. Prior to metamorphism, the granulite was a late ferrodiorite representing residual mafic liquids filter-pressed from the anorthosite. The outcrop was used by deWaard (1965) to establish his clinopyroxene-garnet subfacies of the granulite facies that marks the transition to higher pressures, garnet granulites, and ultimately the eclogite facies.

Across the road from the granulite, a variety of anorthositic rocks (1155 ± 8 Ma) are exposed in the streambed. At the west end of the water-polished outcrops, Woolen Mill gabbro clearly intrudes the anorthosite. However, note an inclusion of gabbro in the anorthosite indicative of coeval relationships. Note the “block structure” where blocks of earlier anorthosite have been broken up and intruded by later magmatic pulses.

Fig. 6



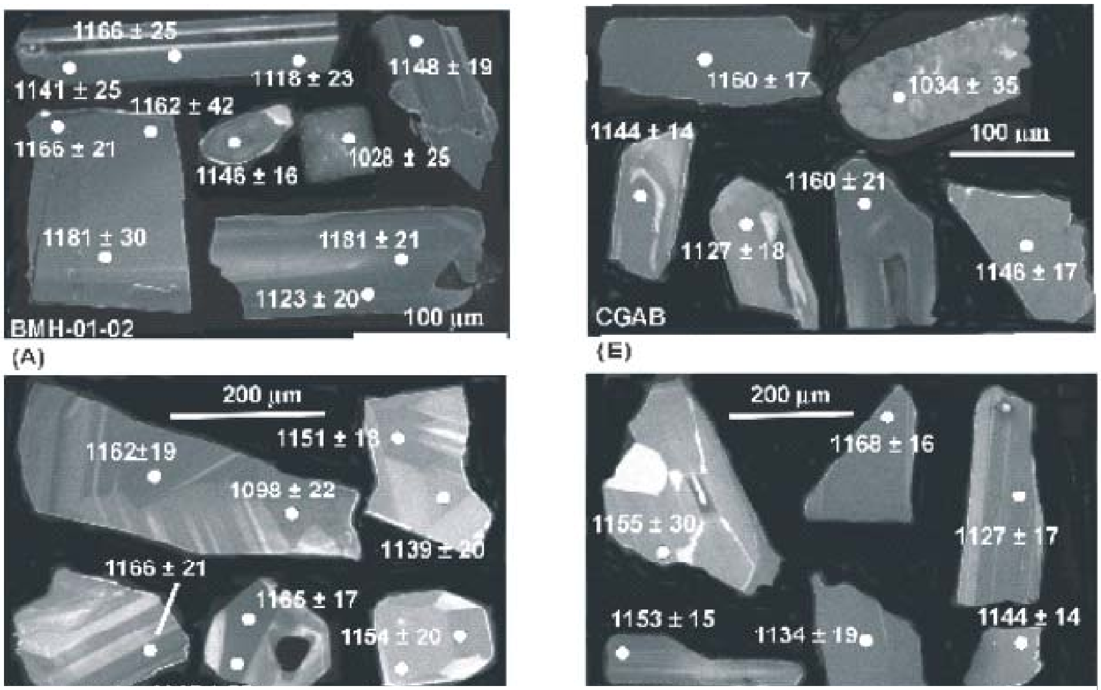


Fig. 7a. CL images of SHRIMPed zircons from Adirondack anorthosite showing ages of spots

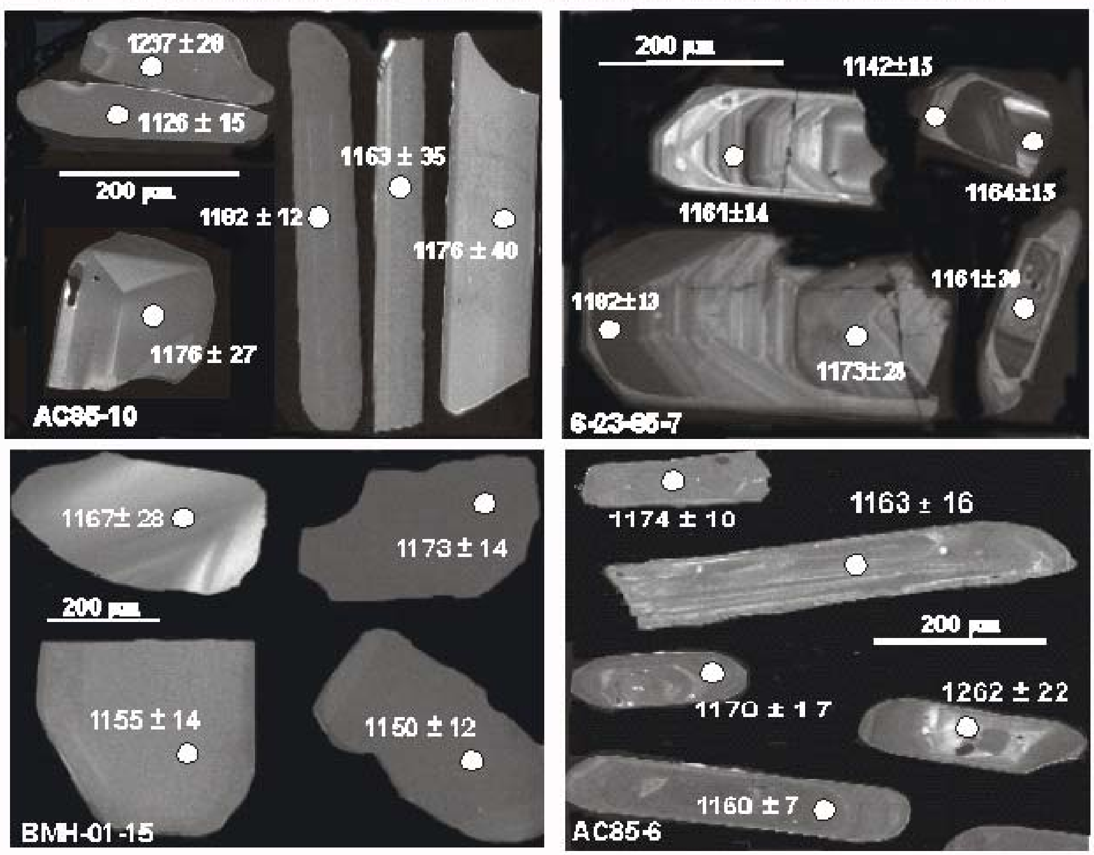


Fig. 7b. CL images of SHRIMPed zircons from Adirondack granitoids showing ages of spots.

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CHEMICAL WEATHERING AND CALCIUM DEPLETION IN ADIRONDACK SOILS

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INTRODUCTION

Acid Rain and the Adirondack Mountains

The term 'acid rain' was first coined in 1872 in *Air and Rain: The Beginnings of Chemical Climatology*, a book published by Angus Smith, an English chemist, who was the first to systematically analyze the chemistry of precipitation in industrialized Britain. The effects of acidic deposition on aquatic and terrestrial ecosystems have been studied intensively since the late 1960's and early 1970's, first in Scandinavia, then in Europe, and eventually in the U.S when acid rain emerged as an important ecological issue (Oden, 1968; Likens et al., 1972).

The Adirondack Mountains receive elevated inputs of sulfur and nitrogen in the form of acid deposition. The acid emissions that cause acid deposition began in the late 19th century and increased well into the 20th century, peaking in the 1970's. Annual emissions of sulfur dioxide in the U. S., for example, peaked in 1973 at 28.8 million metric tons. By 2003, following implementation of the Clean Air Act Amendments (CAAA) of 1970 and passage of Title IV of the Acid Deposition Control Program by Congress in 1990, sulfur dioxide emissions had decreased 50%, to 14.3 million tons.

Prevailing winds from the west carry pollutants emitted in the Midwest, mainly from coal-burning electric utilities, over the northeastern United States and Canada. Precipitation in the Adirondacks averages 100 – 150 cm/yr, with about 30% falling as snow (Johannes et al. 1985). This large quantity of precipitation results in high acid loadings. H^+ deposition averaged about 500 eq/ha/yr over the Adirondacks in the early 1990s, but decreased to about 300 eq/ha/yr by 2002. Average deposition of H^+ and SO_4^{-2} in the Adirondacks has been declining over the past few decades due to significant decreases in SO_2 emissions. Average annual pH values of precipitation have risen over this period from about 4.2 to 4.5. Because precipitation amounts decrease from the western to eastern Adirondacks due to orographic effects, H^+ deposition decreases by about 10% across the mountain range, a distance of about 150 km (Charles 1991, Driscoll et al. 2003). There has been no significant change in nitrogen deposition during this same period because CAAA legislation did not specify limits for NO_x emissions (Driscoll et al. 2001).

Cumulative and lingering effects of over a century of acid rain has led to renewed interest in determining whether terrestrial and aquatic ecosystems in the Adirondacks are showing signs of recovery (i.e., to more natural states, although "natural" is oftentimes difficult to determine because of two centuries of anthropogenic perturbations and lack of historical data). In these studies detailed knowledge of the geology, hydrogeology and geochemistry of the Adirondacks plays a critical role. For example, recent studies of the terrestrial system have shown that base cation depletion in soils, particularly low levels of available calcium in forested watersheds, negatively impacts forest health and the ability of surface waters to recover from acid inputs (c.f. DeHayes et al. 1999, Lawrence et al. 1995, 1999). Because Spodosols (a predominant soil type in the Adirondacks) naturally have low base saturation, leaching of base cations by acidic inputs exacerbates an already fragile balance between nutrient supply for vegetation and organisms that live on or within the forest floor and replenishment of these nutrient cations by litter decay and mineral weathering. Adirondack soils consist mainly of the primary minerals quartz, K- and Na-rich feldspar, and muscovite (April and Newton 1983, April et al. 1986), all of which are highly resistant to chemical weathering. Minerals such as Ca-rich plagioclase, biotite, hornblende, diopside, garnet and calcite are much less abundant in these soils, but they are more susceptible to chemical weathering and therefore can provide base cations to exchange sites at rates comparable with depletion rates accelerated by acidic deposition (April and Newton 1992).

The effects of acid deposition on aquatic systems in the Adirondack Mountains have been studied intensively since the 1970's. Findings show that the acidification status of streams and lakes is strongly influenced by the geology and hydrology of watersheds. Deposition input quantity and quality, the mineralogy and depth of surficial materials, the hydrological properties of soils, groundwater flow paths, wetland processes, snowmelt, etc., all contribute to the final chemical composition of surface waters.

Geology of the Adirondacks

The Adirondack Park comprises about 2.4 million ha (~6 million acres) of predominantly forested land dotted with more than 3,000 lakes and ponds. Water drawn from five major drainage basins flows along 2,400 km of rivers fed by an estimated 48,000 km of brooks and streams. Geologically, the Adirondack region forms the southwestern extension of the Grenville Province of the Canadian Shield with rocks ranging in age from 1.3 to 1.0 billion years old (McLelland and Chiarenzelli 1990, McLelland 2001). The major rock units underlying the area can be broadly divided into three types: granitic gneisses, anorthosites, and metasediments. Marble and other calcite-bearing bedrock occur in a few scattered localities within the metasedimentary units. The mountains and uplands are mantled by till, with deposits thickest in the valleys and thinner at higher elevations. Glacial meltwater deposits composed primarily of stratified sand and gravel fill lowland areas, primarily in the northwestern part of the region. Soils in the region are generally acidic Spodosols, which have developed on till or outwash since deglaciation 12,000 to 14,000 years ago. Soil profiles are typically less than a meter deep.

Primer on Acid Deposition

Deposition of sulfur (S) and nitrogen (N) in the form of acid deposition has impacted soils, surface waters, and biotic components of terrestrial and aquatic ecosystems worldwide. In the northeastern United States (US) the effects of acid deposition on ecosystems has been the focus of numerous field and laboratory investigations over the past quarter century. In acid-sensitive regions, such as the Adirondack region of New York (Adirondacks), acid deposition has reduced fish populations and biological diversity in lakes and streams (Schofield 1976, Baker and Schofield 1982, Gallagher and Baker 1990), reduced growth rates of plants (Worrall 1994, DeHayes et al. 1999), lowered pH and decreased the acid-neutralizing capacity (ANC) of surface waters (Galloway et al. 1983, Driscoll and Newton 1985, Driscoll et al. 1989), caused changes in nutrient cycling and uptake in forests (Cronan 1994, Currie et al 1999, DeHayes et al. 1999), and mobilized metals, such as aluminum, that can be toxic to plants and animals (Tyler 1978, Cronan and Schofield 1990, Cronan and Grigal 1995, Baker et al. 1996, Van Sickle et al. 1996). Acid deposition causes significant changes in soil chemistry, including the depletion of base cations, the release of monomeric aluminum, and the accumulation of S and N compounds, which result in the alteration of soil biotic communities (Wolters and Schaefer 1994).

Quantities of sulfur and nitrogen compounds released by power plants and industry in the United States have markedly decreased pursuant to the 1970 Clean Air Act Amendments (CAAA) and Title IV of the Acid Deposition Control Program of the CAAA of 1990. Amounts of sulfur dioxide (SO_2) in emissions and sulfate (SO_4^{2-}) concentrations in precipitation have shown declines over the past three decades, and sulfate concentrations in surface waters in the northeastern US have decreased over this same period (Jenkins et al. 2007). Despite decreases in acidic inputs to the Adirondacks, ANC has seemingly not improved much because decreased SO_4^{2-} in surface waters largely has been offset by decreases in concentrations of base cations (Stoddard et al. 1999). However, a more recent study of 188 long-term monitoring sites in the eastern United States, from 1990 to 2000, by Stoddard et al. (2003) shows evidence of ANC increases of about 1 μeq/l per year in Adirondack lakes. Slow recovery of chemical water quality in the Adirondacks may be due to historical leaching of base cations from soils, continued deposition of nitrogen compounds, or accumulation of sulfur in soil (Driscoll et al. 2003). The effects of acid deposition in the Adirondacks will be detectable for a very long time. Nevertheless, improvements in water and soil quality are expected. Scientists now have the opportunity to measure ecosystem structure and function at a time when the impacts of acid deposition may be at their worst. Only by establishing a baseline of appropriate geochemical and biotic parameters can we monitor ecosystem recovery in the Adirondack region.

Base cation depletion in soils and decreases in stream and lake water ANC are of particular concern in acid-impacted watersheds. Roy and Driscoll (2001) reported that the concentrations of base cations have decreased in 26 of 48 lakes designated as long-term monitoring sites by the Adirondack Lakes Survey Corporation. Adirondack soils naturally have low base saturation and, therefore, any process that accelerates the removal of base cations from exchange sites decreases the ability of the soil to sustain plant growth and contribute to the ANC of surface waters. In soils that are composed primarily of resistant minerals the potential for base cation supply is low and exchange sites may become occupied by hydrogen ions and aluminum, rather than calcium, magnesium, and potassium. The extent to which calcium has decreased and whether calcium is currently decreasing in soils of the Adirondack region are unclear. The amount of available calcium in soils affects the structure and function of terrestrial ecosystems and the ability of surface waters to recover from acid deposition.

Studies of the effects of acid deposition on terrestrial biota have been primarily aimed at commercially important forest plants. Acid deposition has been linked to the decline of sugar maple (*Acer saccharum*) in eastern North America (Duchesne et al. 2002, Sharpe 2002), although other factors are probably involved (Minorsky et al. 2003). Similarly, acid deposition is implicated in the high mortality of pines (*Pinus*) in southwestern China (Feng et al. 2002) and spruces (*Picea*) in Europe (Roberts et al. 1989) and North America (Fowler et al. 1989). Calcium apparently plays a large part in these effects because of its role in membrane stabilization and cold tolerance (DeHayes et al. 1999, Horsley et al. 2000). Effects of acid deposition on terrestrial animals have received little attention; however, available evidence suggests that calcium also is important here. Abundance and diversity of land snails are negatively associated with acid deposition (Wäreborn 1992, Graveland and van der Wal 1996). Snails have greater demands for calcium than other invertebrates because of the composition of their shell, and abundance and diversity of snails have been associated with calcium abundance in soil (Wäreborn 1969, Gärdenfors 1992, Graveland and van der Wal 1996, Johannessen and Solhoy 2001, Hotopp 2002). Caterpillars had a lower calcium concentration at an acidified forest than a reference forest in Ontario (Mahony et al. 1997). Acid-induced calcium depletion can negatively affect insectivorous birds (Graveland and van der Wal 1996, Graveland and Drent 1997, Hames et al. 2002), which use calcium-rich invertebrates as important sources of dietary calcium during egg-laying (Graveland and Van Gijzen 1994, Graveland 1996, Taliaferro et al. 2001). A similar connection might be postulated for insectivorous mammals, but we are unaware of any work that bears upon this topic. The effects of acid deposition on terrestrial insectivores and their prey remain unclear (Ormerod and Rundle 1998). However, changes in the physical structure of the environment and the availability of calcium are likely mechanisms for changes in terrestrial animal assemblages (Mahoney et al. 1997).

Effects of acid deposition on stream biota have been studied extensively in eastern North America and throughout northern and central Europe. Typically, these studies have compared chronically acidified stream systems ($pH < 5$) to streams with moderate to circum-neutral (5 – 7) pH values (Mulholland et al. 1992, Hermann et al. 1993, Fitzhugh et al. 1999). Densities of bacteria living on decomposing leaves are not severely changed by acidification, but bacterial productivity is negatively affected (Palumbo et al. 1987, Osgood and Boylen 1990, Mulholland et al. 1992, Hermann et al. 1993, Maltby 1996, Dangles and Guerold 2001). Surprisingly, periphyton abundance often increases at low pH, but there is a concomitant shift in species composition to filamentous green algae and acid tolerant diatoms (Planas et al. 1989, Mulholland et al. 1992, Rosemond et al. 1992, Hermann et al. 1993, Junger and Planas 1993, Ledger and Hildrew 2000). Thus, energy sources for stream macroinvertebrates are affected by stream acidification.

There have been several different mechanisms proposed to explain the negative impacts of acidification on stream macroinvertebrates. The most common effect is the toxicity of aluminum, which increases in concentration when calcium is depleted from soils (Hermann and Anderson 1986, Hermann 1987, Mulholland et al. 1992, Rosemond et al. 1992, Hermann et al. 1993), but experiments in simulated stream chambers where soil aluminum does not play a role in toxicity have shown different species to vary in their sensitivity to low pH (Bell 1971, Courtney and Clements 1998). Mayflies (Ephemeroptera) are the most sensitive macroinvertebrates to acidification, and their loss appears to be related to aluminum toxicity (Hermann and Anderson 1986, Hermann 1987, Rosemond et al. 1992, Hermann et al. 1993, Brakke et al. 1994, Ledger and Hildrew 2000). Some caddisflies (Trichoptera) are also lost from acidified systems, but there are caddisfly species that are acid tolerant (Mulholland et al. 1992, Rosemond et al. 1992, Hermann et al. 1993, Courtney and Clements 1998). In addition to aluminum toxicity, calcium depletion in acidified streams negatively affects macroinvertebrates with high calcium requirements such as snails and crustaceans (Mulholland et al. 1992, Hermann et al. 1993, Brakke et al. 1994, Rukke 2002).

Besides physiological effects of acidification on stream macroinvertebrates, changes in trophic dynamics have also been proposed. Most acidified systems are small, forested streams where allochthonous sources of energy predominate (Vannote et al. 1980, Wallace and Webster 1996, but see McCutchan and Lewis 2002). Because of the negative effects of acidification on microbial communities that colonize leaf detritus, the quality of coarse particulate organic matter (CPOM) is reduced and leaf decomposition rates are decreased due to fewer shredders and reduced feeding rates of shredders (Mulholland et al. 1992, Rosemond et al. 1993 Hermann et al. 1993, Dangles and Guerold 1999, 2001, Guerold et al. 2000, Ledger and Hildrew 2000, Dangles 2002). Also, if light is not limiting in these forested streams, algal abundance can increase favoring grazing macroinvertebrates. However, it is the grazing mayflies and caddisflies that are the most sensitive to acidification, and the lack of grazers has been used as a partial explanation for the increase in algae in acidified streams. Thus, a typical detritivore-based system may support few shredders because only a few species can tolerate the acid conditions, and because the quality of the CPOM does not support the same growth rates or production of macroinvertebrate shredders. Also, whereas algae are abundant in these systems, grazers are not that important because most are lost due to physiological limitations.

SOILS

Soils in the Adirondacks are typical Spodosols that have developed on glacial till and outwash overlying bedrock. They average less than 1 meter in depth and consist of five distinct horizons: O, A, E, B and C, which are described below. Please note that while all of the thickness and soil pH values included in the following descriptions represent data from the three sites visited on this trip, they are typical values for soils developed on this kind of parent material throughout the Adirondacks. A soil profile representative of the field area is shown in Figure 1.

The uppermost organic (O) horizon ranges from 3-10 cm in thickness and is composed of plant and animal debris in the beginning stages of decomposition. The A horizon directly underlies the O horizon. It is generally dark brown to black in color and rich in decomposed organic material, but also contains clays and other resistant minerals. A horizons vary in thickness from 3 to 16 cm at these sites. Field moist soil pH values for the O and A horizons are extremely low, ranging from 2.8 to 4.4, with A horizons being slightly more acidic (avg. = 3.1) than O horizons (avg. = 3.6).

The organic poor E horizon has a distinctive light gray color and sandy texture in Spodosols. Anywhere from 4 to 20 centimeters thick, it is the zone of maximum leaching in the soil profile and is composed of predominantly clay minerals (April et al., 1986; 2004) and very resistant primary minerals, such as quartz. Soil pH of this horizon is also very acidic, ranging from 3.4 to 3.7 (avg = 3.6). The A and E horizons are referred to as eluvial horizons in the soil profile because they are the layers where the most leaching takes place. These horizons are thicker and better developed under coniferous trees than under deciduous stands.

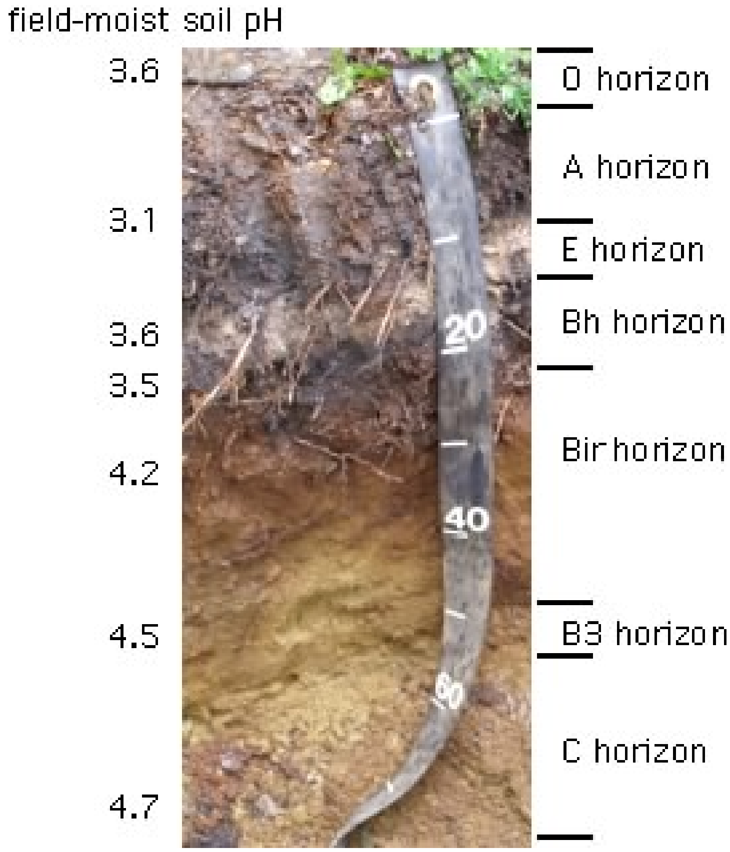

Figure 1: Typical soil profile in field area, delineating horizons and average soil pH values. Depths are in cm.

The illuviated B horizon is much thicker than the upper horizons, with an average thickness of about 40 cm. The B or "spodic" horizon is diagnostic for Spodosols and is characterized by accumulation of Al and Fe oxides and organic matter compounds which have been leached from above and re-deposited in this horizon. The B horizon is usually divided into subhorizons based on the type of material that accumulates, usually seen in the field as color differences in the soil profile. An 'h" designation (e.g. Bh subhorizon) signifies accumulation of humic organic material, giving the soil a dark brown color. "Bir" refers to accumulations of iron in the subhorizon and is indicated by soil with a reddish color. At these sites, both Bh and Bir subhorizons are present in almost all soil pits. Bh horizons are typically just below E horizons; Bir horizons are located below Bh horizons. Field moist soil pH values in the B horizon are slightly higher than in the upper horizons and increase with depth throughout the B from an average of 3.5 at the top of the horizon (Bh) to 4.5 near the bottom (B3 subhorizon).

The lowest horizon in the soil profile is the C horizon, which consists of partially weathered parent material. In this region, the parent material upon which the soil profiles have developed is till or sandy glacial outwash. The highest soil pH values are measured in the C horizon (avg. = 4.7).

SOIL MOISTURE

Soil moisture collectors, or lysimeters, are used to sample water from unsaturated soils. The chemical composition of soil moisture can give a good indication what nutrients are readily available to vegetation as well as what kinds of exchange reactions are taking place in the soil zone. In 2005, we installed soil moisture collectors at Town of Webb Site 10 (STOP 1) and Covewood (STOP 3) at different depths in the soil profiles. Two types of lysimeters were installed: suction lysimeters and passive lysimeters. Installing both types of lysimeters allows us to determine if there are any chemical differences between water that is tightly bound to soil particles (suction) and water that moves freely in the pore spaces (passive). Suction lysimeters use a vacuum to pull pore water from unsaturated soils. Suction lysimeters used here consist of a porous ceramic cup through which water is pulled from the soil, a length of PVC pipe which holds the water, and a stopper to maintain the vacuum. (Fig. 2)

Figure 2: Suction lysimeter used at Town of Webb Site 10 and Covewood – stops 1 and 3 on the field trip. (from http://www.soilmoisture.com/prod_details.asp?prod_id=1096&cat_id=16)

Passive lysimeters are devices that capture water moving downward through the soil by gravity; they do not use suction to pull water out of the soil. Two types of passive lysimeters used here are "wick" lysimeters, which employ fiberglass wicks to collect soil moisture and carry it to a collection container (Fig. 3); and simple rectangular containers of clean, pure quartz sand with a drainage tube to carry water to a collection container. Samples of soil moisture have been collected from these devices over the past three years to determine differences in soil moisture chemistry with horizon.

Figure 3: Wick lysimeters sampling soil moisture from two different horizons in an Adirondack soil pit.

As part of an NSF-funded research project investigating calcium depletion in Adirondack watersheds resulting from acid precipitation, we applied 2.25 tonnes of powdered limestone (in three separate applications of 0.75 tonnes each) over 50-meter diameter circular areas at Town of Webb Site 10 (STOP 1) and Covewood (STOP 3) between 2005 and 2007. These soils are being monitored for exchangeable cations associated with the solid phases (minerals and organic material), and pH and calcium concentrations in the liquid phase (soil moisture) in order to determine the effects of the calcium additions on the soil system. Other scientists working on the project are investigating effects of calcium depletion and calcium additions on terrestrial and aquatic biota.

ACKNOWLEDGEMENTS

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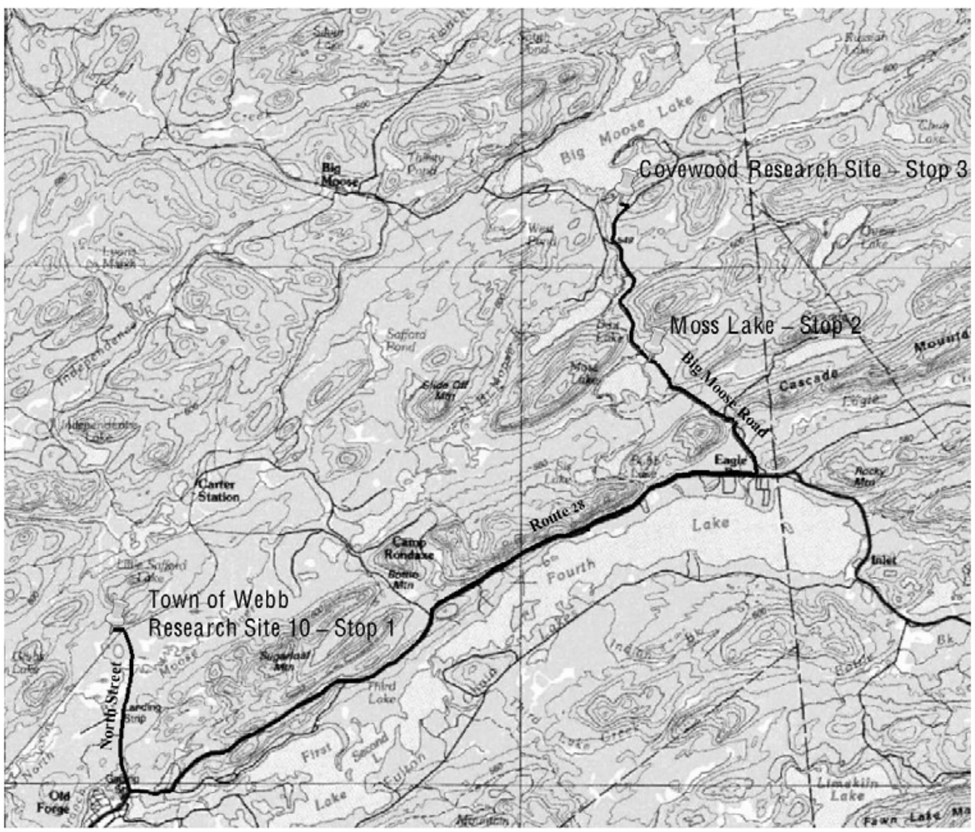

Figure 4: Location map of stops for field trip in Adirondacks.

ROAD LOG AND STOP DESCRIPTIONS

We will depart from the public parking lot located in Old Forge, NY across from the Old Forge Hardware store and the Old Forge Fire Dept, located on NYS Route 28. Head north on NYS Route 28.

Total Miles	Miles from Last Point	Route Description
0.0	0.0	Leave parking lot. Turn onto Route 28 N.
0.3	0.3	Turn left onto North St.
0.9	0.6	Note "outcrops" of glacial outwash sand on the left.
1.3	0.4	Paved road ends
1.8	0.5	Cross bridge over Moose River
2.3	0.5	Cross railroad tracks
2.4	0.1	Bear left, staying on road you are on
2.5	0.1	Pull off road to the right and park

STOP 1: TOWN OF WEBB "SITE 10"

18T: 0502060
UTM: 4843218
N43°44.519'
W074°58.464'

Note: If you wish to return to this area, permission must be obtained from the Town of Webb supervisor. During the fall months, this land is leased to a hunting club and is generally inaccessible to the public until the winter season, when the trails are re-opened for snowmobiling.

At this site, we will walk a few hundred meters into the woods to dig a soil pit and view monitoring and collection equipment used to study soil and soil moisture. This site is part of an NSF-funded research project studying calcium depletion in Adirondack watersheds exposed to acid precipitation – the "site 10" designation is our own internal label. Soils here, as in most of the Adirondacks, are typical spodosols. In this area, soils are developed on a sandy outwash material, which could be seen in roadside "outcrops" as we traveled from Old Forge into the site. Few cobbles or boulders are encountered when digging soil pits in this area.

As part of the research project, a 50-meter diameter circular area on one side of the small stream flowing through the site was treated 3 times with 0.75 tonnes (each time) of powdered limestone and an identically-sized untreated area on the opposite side of the stream was used as a control. Soil and soil moisture chemistry has been monitored at this site since 2005. Figures 5 and 6 show soil pH and exchangeable calcium with depth over time before and after application of powdered limestone. As expected, additions of calcium carbonate raised the soil pH in the upper horizons and, over time, that effect is being translated lower into the soil profile. Similarly, exchangeable calcium is increased in the upper horizons of the soils treated with powdered limestone, although that increase is only seen down to about 15 cm, even 3 years after the first application of calcium carbonate.

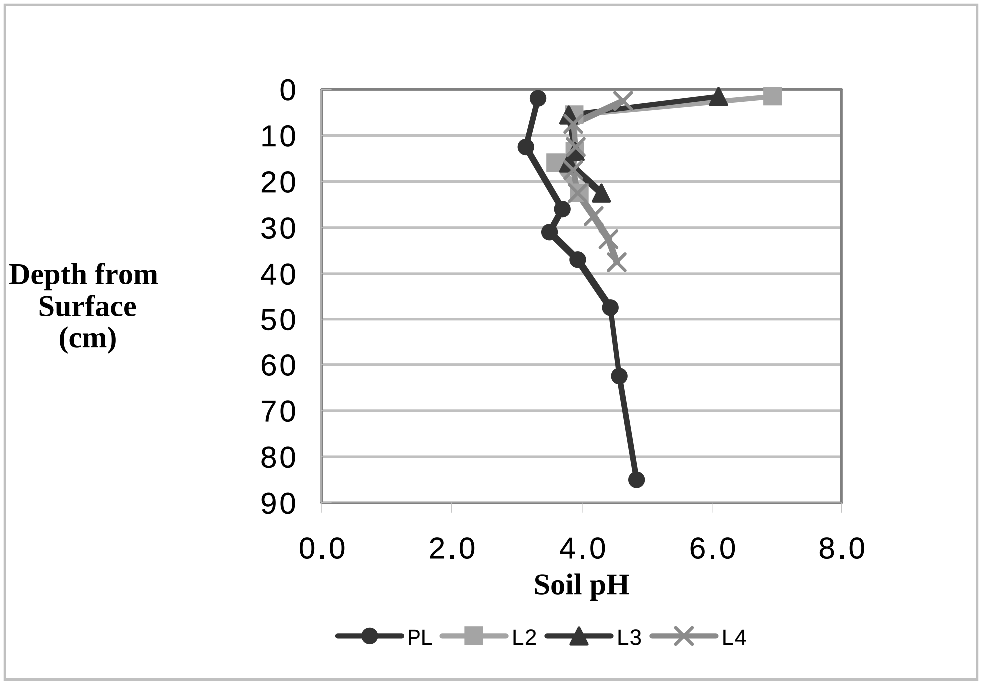

Figure 5: Soil pH in Town of Webb Site 10 soil pit. PL = pre-liming; L2 = after second application; L3 = 9 months after third application; L4 = 20 months after third application. Note that successive application of limestone raises soil pH in the upper horizons (approximately the top 40 cm).

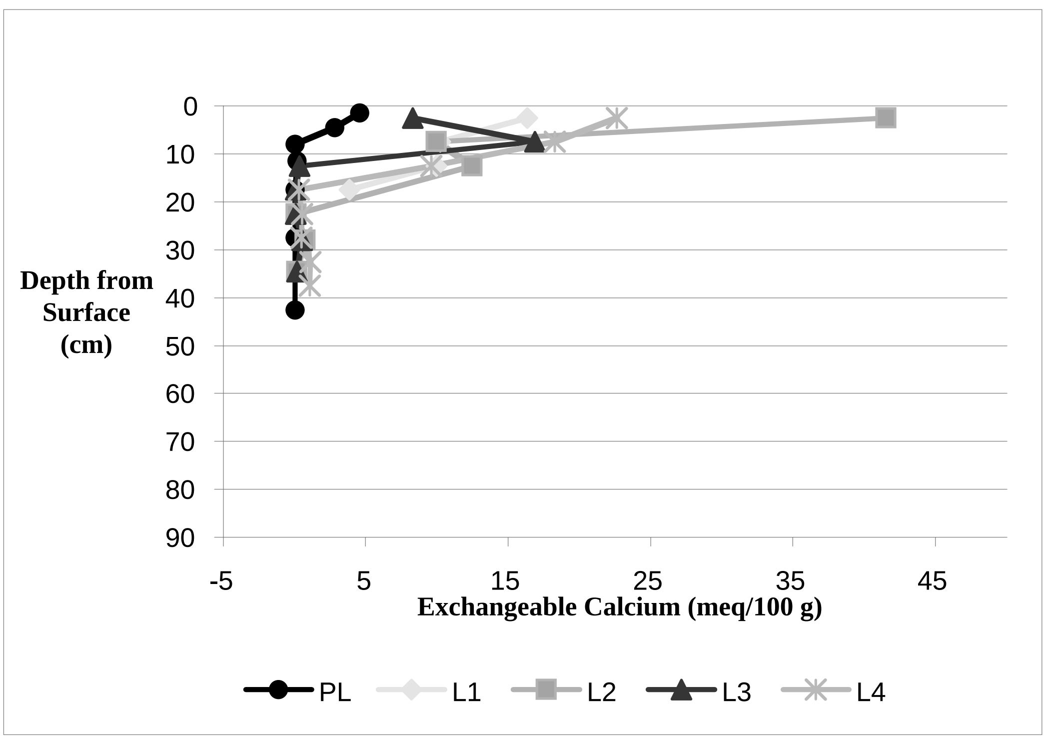

Figure 6: Exchangeable calcium in Town of Webb Site 10 soil pit. PL = pre-liming; L1 = after first limestone application; L2 = after second application; L3 = after third application. Note that successive application of limestone increases exchangeable calcium in the upper horizons (approximately the top 20 cm).

Return to vehicle. Retrace route back to intersection of North St. and NYS Route 28N.

Total Miles	Miles from Last Point	Route Description
4.7	2.2	Turn left onto NYS Route 28N
13.7	9.0	Turn left onto Big Moose Rd
15.9	2.2	Turn left into parking lot for Moss Lake trailhead and park

STOP 2: MOSS LAKE TRAILHEAD

18T: 0512455
UTM: 4848383
N43°47.304'
W074°50.712'

We will take a short walk here to the shore of Moss Lake and take a reading of water pH. The south side of the Moss Lake watershed consists of thick sand-covered bedrock and much of the north side has a thinner cover of till and bedrock close to the surface. Flow paths of water moving through the soils and underlying materials determine how well buffered the incoming acid precipitation becomes and therefore affects the lake pH. At this location, we will also view the Moss Lake National Atmospheric Deposition Program data collection site (site NY29 – Fig. 7). Precipitation pH for the past few years is shown in Figure 8.

Figure 7: NADP site NY29 at Moss Lake showing wet/dry collector, rain gage and other meteorological equipment (from http://nadp.sws.uiuc.edu/sites/siteinfo.asp?net=NTN&id=NY29).

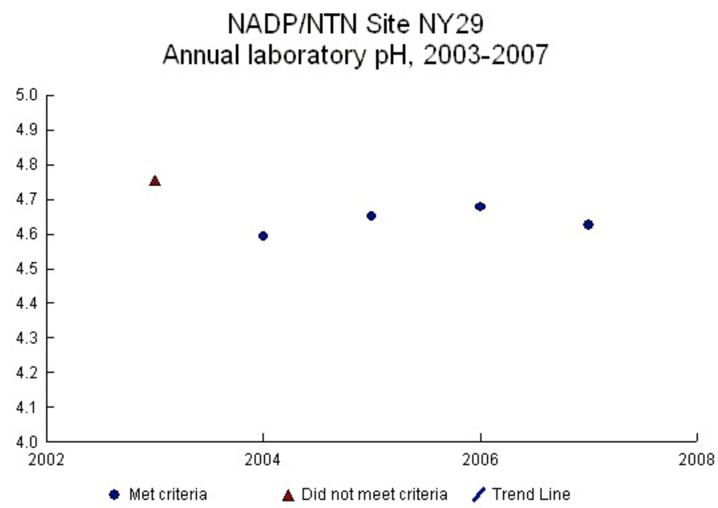

Figure 8: Mean annual pH values of precipitation samples from Moss Lake NADP site (from http://nadp.sws.uiuc.edu/sites/siteinfo.asp?net=NTN&id=NY29)

Return to vehicles. Turn left back onto Big Moose Rd.

Total Miles	Miles from Last Point	Route Description
17.1	1.2	Pass Windfall Pond trailhead on the right. We will not take the trail to Windfall Pond, but it is a nice walk. The trail passes over the outlet of Windfall Pond. The Windfall Pond watershed contains some calcium carbonate rock, which acts to buffer acidic precipitation. Outlet pH values are generally higher here than in other streams in the Big Moose area because of this buffering effect.
18.1	1.0	Turn right onto Covey Rd
18.2	0.1	Turn right at sign for Covewood Lodge
18.3	0.1	Turn right at sign for South Bay Rd.
18.4	0.1	Pull off to the left and park.

STOP 3: COVEWOOD

18T: 0511908
UTM: 4851626
N43°49.055'
W074°51.121'

Note: This land is private property and permission must be obtained from the owners of Covewood Lodge.
On the north shore of Big Moose Lake, Covewood Lodge and surrounding property presents a somewhat different surface material than previous stops. The surficial material upon which soil are developed in this area is a thinner, bouldery till, as is evident just by looking at the surrounding landscape. Numerous cobbles and

boulders are encountered when excavating soil pits around here. Depth to the saturated zone is shallower here and soils are often wetter than at Town of Webb Site 10 or Moss Lake.

The owner of Covewood Lodge uses spring water (groundwater) to supply his lodge and cabins. Several years ago, he determined that the water was very acidic (below pH of 4) and he has attempted several different methods for neutralizing the acidity. We will look at one or two these devices as well as view monitoring equipment we have set up in this area as part of the aforementioned research project.

Covewood is one of the sites where powdered limestone was applied in 2005, 2006 and 2007. Exchangeable calcium in the soils is shown in Figure 9 below. As at Town of Webb Site 10, the data show that exchangeable calcium in the soil has increased down to a depth of about 10 cm due to the addition of calcium carbonate. We have not seen any effect of the limestone additions on the groundwater yet.

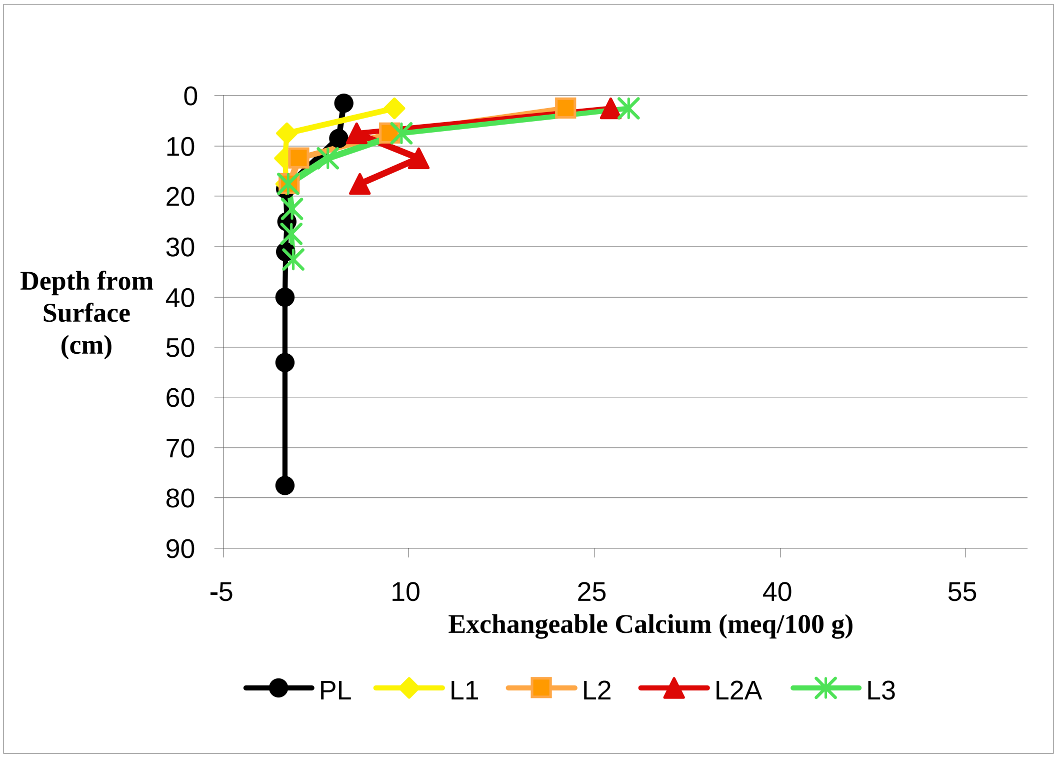

Figure 9: Exchangeable calcium in Covewood pit BIG-12. PL = pre-liming; L1 = after first limestone application; L2 and L2A = after second application; L3 = after third application. Note that successive application of limestone increases exchangeable calcium in the upper horizons (approximately the top 10 cm).

To return to Old Forge: leave Covewood property and turn left onto Covey Rd

Total Miles	Miles from Last Point	Route Description
18.6	0.2	Turn left onto Big Moose Rd
22.7	4.1	Turn right onto NYS Route 28 S
32.0	9.3	Turn left into public parking lot.

END OF FIELD TRIP

DEGLACIAL HISTORY OF THE UPPER HUDSON REGION

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INTRODUCTION

The Hudson River descends from the Adirondack Mountains through a beautiful gorge and empties into the northern Hudson lowland onto a diverse complex of ice contact, glaciofluvial and glaciolacustrine sediments. Early workers such as Ogilvie (1902), Woodworth (1905), Miller (1914, 1925), Fairchild (1917), Stoller (1922), Chadwick (1928) and others posed early explanations and some interesting contrasts for the diverse suite of glacial deposits. All provided a solid context for future workers. Readvance hypotheses were a popular paradigm in the 1960's and 1970's when Connally & Sirkin (1969, 1971) hypothesized a major readvance of the Hudson-Champlain (H-C) ice lobe through much of the northern Hudson region. De Simone (2008) recently proposed this Luzerne readvance lacks strong evidence for its occurrence and should be abandoned. This trip will address concerns over the proposed Luzerne readvance hypothesis and present an alternative interpretation of the stratigraphy of the Luzerne Mountain gorge. The interpretation is based upon a synthesis of past descriptions of the Corinth Road exposures combined with insights from borehole data analyzed from a site in the western portion of the gorge.

The southern Champlain and northern Hudson lowlands contained a continuum of glaciolacustrine lakes that fronted the retreating H-C lobe as the ice front melted back during the late Wisconsinan. Woodworth (1905) recognized the shared Quaker Springs (QS), Coveville (CV) and Fort Ann (FA) water levels in the two lowlands. Rayburn (2004), Rayburn et al (2005, 2007) and Franzi et al (2007) developed a modern chronology of deglacial events in the Champlain lowland building upon the classical work of Chapman (1937). They provide a valuable time frame for deglaciation and the sequence of Coveville and Fort Ann lake levels. De Simone (2006) recently extended their 4ft/mi tilted water planes into the northern Hudson lowland and found good agreement with major deltas for the Fort Ann, Coveville and earlier Quaker Springs and Albany II (ABII) lake levels, the latter confined to the Hudson Valley. This correlation appears to solve the old problem of aberrantly low tilts on Hudson Valley water planes that were a feature of regional syntheses for many decades (Woodworth 1905, LaFleur 1979, De Simone and LaFleur 1985, 1986, Dineen et al 1992). We'll discuss these lake levels, their ice margins and their outlets as we tour the H-C divide region.

Most recently, De Simone et al (2008) present a summary of work in progress that adds to our understanding of the deglaciation of the H-C region. In this volume, Wall expands upon his earlier (Wall 1995) analysis of IroMohawk discharge; Miller discusses the implications of the Cohoes mastodon site for the Younger Dryas with new information based upon his earlier work (Miller and Griggs 2004); Rayburn contributed a revised water planes correlation and insights into QS in the southern Champlain Valley; and, Dineen presents data on the onset of Lake Quaker Springs incorporating earlier analysis (Dineen and Miller 2006).This trip will expand that synthesis to include the Lake George basin. We will visit some different sites than the FOP 2008 trip but will link our discussion thread into that synthesis.

NORTHERN HUDSON LOWLAND GLACIAL LAKES

The Lacustrine Link Between The Hudson And Champlain Lowlands. --- We'll consider the northern Hudson lowland to represent that portion of the Hudson Valley north of the Hudson-Mohawk confluence in Waterford, NY. On the 70th annual reunion of the Northeastern Friends of the Pleistocene, Dave Franzi and colleagues took us on a tour of the deglaciation of the northern Lake Champlain Valley (Franzi et al. 2007). They estimated ice margin retreat through the region of about 0.4 – 0.5 km/year. The cause for the CV-FA transition was a flood event related to the coalescence of Lake Coveville and Lake Iroquois when the ice margin reached the northern flank of the Adirondacks. Franzi and colleagues demonstrated evidence for this at Cobblestone Hill on Altona Flat Rock (Rayburn et al. 2005).

Lake Coveville dropped catastrophically to Lake Fort Ann during the Lake Iroquois flood that likely breached the Coveville threshold. Woodworth (1905) first named Lake Coveville and proposed "The Cove" near the mouth of Fish Creek on the Schuylerville quadrangle and Chapman (1937) concurred this was the threshold for Lake Coveville. Investigation of "The Cove" (De Simone 1977, 1985) indicated it was formed by a portion of IroMohawk River discharge that flowed through the Saratoga Lake basin. Rayburn (2004) concurred and projected the Coveville level well above "The Cove" and southward down the Hudson Valley. Lake Coveville is the key link between the Hudson and Champlain lowlands as it was coeval across the divide region.

Lake Fort Ann. --- Lake Fort Ann lasted for approximately 200 years based upon varve chronology they analyzed before ice margin retreat opened the drainage route through the Gulf of St. Lawrence, FA drained, and the Champlain Sea (CS) entered the isostatically depressed valley. Varve chronology combined with radiocarbon ages (Rayburn et al., 2007) indicates that the CS began around 13,000 calibrated years BP. Lake Fort Ann did not extend into the Hudson lowland as a lake. The threshold of the lake in Fort Ann, NY, is bedrock controlled and the FA waters represented a high discharge fluvial period through the Hudson lowland that excavated the present valley occupied by the Hudson River southward from Fort Edward near Glens Falls.

Older Glacial Lakes. --- Lake Albany I (ABI) and Lake Albany II (ABII) had dropped to the next lower lake, Lake Quaker Springs (QS), while ice remained in the Hudson Lowland. Therefore, these two lakes have no extension into the Champlain Lowland. De Simone et al (2008) propose an ice margin where the drop from ABI to ABII occurred based in part upon overlooked insights provided by Schock (1963).

Lake Quaker Springs was first recognized by Woodworth (1905) who named the lake for the threshold he inferred at Quaker Springs, NY, near the better known city of Saratoga Springs. A current hypothesis (De Simone, 2006) suggests the actual dam for the lake may have been at a nick point along the Hudson near the Rensselaer-Columbia County line. The extension of Lake Quaker Springs into the Champlain Lowland has remained problematic. Previously, strandline features on the VT side of the lowland have been correlated to QS but no one has recognized any QS strandline features on the NY side. Recent work (De Simone, 2006, De Simone and Becker, 2007) indicates that QS penetrated as far north as Brandon, VT, on the east side of the lowland but probably not much farther north. This is consistent with earlier work in the Brandon area (Connally, 1970). A reasonable re-construction of ice margins enables us to resolve the apparent disparity of QS on the east side of the lowland but not on the west side. The Lake George basin was occupied by Lake Quaker Springs and features tentatively correlated to QS terminate near the northern end of the basin just south of Ticonderoga. This is consistent with Rayburn's (2004) observation that the Street Road delta approximately 6 mi north of Ticonderoga is the highest ice contact delta in Crown Point and is a CV delta. So, QS did not extend to Crown Point on the west side of the southern Champlain lowland. Thus, QS now also becomes a glacial lake coeval between the Hudson and Champlain lowlands.

Summary of Northern Hudson Lowland Lakes

Lake Albany I (AB I). --- This is recognized as a post-Erie interstadial lake occupying the lower through middle Hudson Valley and extending along the eastern lateral margin of the Hudson-Champlain (H-C) ice lobe as far north as the mouth of the Hoosic River on the Schagticoke, NY, quadrangle where there is an AB I delta. Along the western margin of the H-C lobe, the Queensbury Delta of the GlacioHudson River plots below the ABI water plane but above the AB II. This delta is south of Glen Lake on the Glens Falls, NY, quadrangle, and represents deposition into an ice marginal lake separate from AB I. If the Queensbury Delta was deposited into AB I, then using ice surface profile data and projecting to the eastern margin of the H-C lobe, the mouth of the Batten Kill would have been open. Yet, there is no AB I delta of the Batten Kill. Therefore, the Milton Delta represents the northernmost AB I delta clearly deposited into the western side of AB I. This indicates the H-C lobe position extended across the Hudson Valley near Saratoga Springs, NY, similar to where Chadwick (1928) placed the margin.

The AB I water plane projects below sea level in the lower Hudson Valley where Hell's Gate and/or the Terminal Moraine may have served as the threshold. It is also possible that a moraine dam in the Hudson Highlands held back the waters of AB I. Much more work is needed to answer the question of where the ABI threshold was located.

Lake Albany II (AB II). --- The ice retreated from the Troy North quadrangle where Schock's mapping (1963) suggests an ice front location when AB I dropped some 80ft to AB II. There is no AB II delta of the Batten Kill and Woodworth (1905) and De Simone (1983) suggest an ice margin location for the farthest northward extent of AB II just north of the mouth of the Batten Kill. The AB II water plane projects into lake sediments well south of Hudson and Catskill, NY. La Fleur's suggestion (1979) that the uplifting lake bottom served as a lake threshold may be an appropriate hypothesis.

Lake Quaker Springs (QS). --- The water level dropped some 50-60ft to establish Lake Quaker Springs. It is problematic if this drop was triggered by a high discharge event that would have come down the IroMohawk River. The retreating H-C lobe defended a northward expanding Lake Quaker Springs as the ice retreated from north of the Batten Kill confluence to a position at least north of the Forest Dale Delta and Fern Lake on the Brandon, VT, quadrangle. The ice margin at the time of lowering to CV may have been just north of the Neshobe River through the Fernville area on the east side of the valley. On the west side of the valley, there is no identified QS delta above the CV Street Road delta in Crown Point, so the Crown Point area must still have been ice covered. Chadwick (1928) proposed a Lake Bolton as the final and lowest glacial lake in the Lake George basin. Connally and Sirkin (1969) suggested there was no need for a separate Lake Bolton and we concur. Numerous strand line features just above the present shore of Lake George fall onto the Quaker Springs projected water plane. Therefore, this is the main level of the glacial lake that occupied almost the entire basin until the ice retreated to near Ticonderoga.

The QS water plane projects deep into the lake bottom sediments along the Rensselaer County-Columbia County line. Again, the exposed and up tilting lake bottom may have served as the threshold for Lake Quaker Springs. However, there is a pinching of the valley width here. Dineen (2008, pers. comm.) reports a buried nick point at the confluence of the Hudson-Battenkill channel with a southward extension of the preglacial conjoined Colonie & Mohawk channel near Coeymans, just north of the county line. A nick point may have provided a grounding line for the retreating glacier and established a condition similar to that where the proposed CV threshold developed. Further study is planned.

Lake Coveville (CV). --- The water level in the H-C lowlands dropped some 60ft to establish Lake Coveville. This occurred with the ice margin just south of Ticonderoga on the west side and just north of the Neshobe River near Fernville on the east side of the Champlain lowland. Lake Coveville fronted the retreating H-C ice lobe through the northern Champlain Valley as previously detailed by Chapman (1937) and discussed briefly already. The CV water plane projects into the lake bottom sediments south of Albany and Rensselaer, NY. However, it is possible that the CV waters were held back by a thick plug of ice marginal sediments composed of sand and gravel near Waterford, NY.

4.0FT/MI TILTED WATER PLANES THRU CHAMPLAIN TO HUDSON REGION

Currently, more than 40 strandline features in the northern Hudson through southern Champlain lowlands have been identified on maps and in the field. Deltas from tributaries were given primary importance because the topset-foreset contact positively correlates to the recognizable break in slope in the delta morphology on a topographic map. This assumes that the delta fore-slope has not been severely modified by post-depositional erosion. Elevations of some strandline features have been determined by both a Brunton MNS Multi-Navigator GPS-Altimeter-Barometer and a Garmin Foretrex 101 GPS unit that was cross-checked with the Brunton unit. John Rayburn's Brunton unit was periodically corrected for atmospheric variation by re-calibration at bench marks. Exposed topset-foreset contacts that have been measured closely match map determination of the same contact based upon delta morphology. The elevation of each strandline feature was identified on a topographic map with either a 10 foot or 20 foot contour interval. Range of error in the elevation of a feature is assigned to be +/- 10 feet or +/- 3 meters.

Shoreline features such as beaches with a typical beach profile and sand spits deposited by long-shore currents were used. Wave cut terraces, if present, were also used. River terraces must be used with more caution as there are numerous terraces along the Hudson River and its tributaries. These terraces have many elevations which may or may not coincide with any known water level. For example, the use of low terraces along the Hudson River would result in a scatter of points through which anywhere from 1-3 Fort Ann "water planes" might be drawn, none of which would be statistically valid. The terraces are merely a record of the down-cutting to lower base level by streams. All terrace surfaces below the Coveville level in the Hudson Lowland represent fluvial terraces and do not represent lacustrine strandlines.

Water Planes Tilt At 4.0ft/Mi +/- 0.2ft/Mi (0.75m/Km +/- 0.03). --- The plot by Rayburn illustrated in **Figure 1** represents an integration of the data assembled by De Simone with data from Rayburn for the entire Champlain Lowland. Historically, strandline data for the Hudson Valley were correlated to indicate water planes with abnormally low tilts compared to the surrounding Ontario, Erie, St. Lawrence and Connecticut Lowlands. Woodworth (1905) first plotted shoreline features for the H-C region and produced a plot indicating water planes with tilts less than 3ft/mi. Every worker since then has largely duplicated Woodworth's initial effort with only slight modifications. A different approach was taken by De Simone (2006) to attempt to rectify what appeared to be a long standing error in the plots of Hudson Valley strand line features. It is interesting to note that Miller (1914,1925) discussed Glacial Lake Warrensburg, a long lake that occupied the Schroon and Hudson River valleys in the southeastern Adirondacks. Miller determined the tilt for the Lake Warrensburg water plane to be 4.5ft/mi although he cautioned readers about the altitudes of the features he used due to the accuracy of the topographic maps. It would be very useful to re-map these areas and provide new data for Lake Warrensburg.

The approach was to accept the recent plot of approximately 4.0ft/mi data for the CV and FA water planes in the Champlain Valley and project these water planes into the northern Hudson Valley. The data were initially limited to the deltas deposited by tributaries to the Hudson Valley, those deposited by the GlacioHudson River as it exited the Adirondack Mountains, and ice contact deltas deposited generally in the middle of the Hudson Valley from subglacial melt water sources. The resulting hand drafted plot on simple graph paper resulted in a strong positive correlation for CV deltas along an extension of Rayburn's 4.0ft/mi plot of Champlain Valley data. This was surprising, however, the most surprising result was that higher deltas in the northern Hudson Valley also all fit along 4.0ft/mi tilted water planes that were identified as QS, AB II and AB I. The very minor deltas of small tributaries entering the lowlands are being added to this data set at the suggestion of Al Randall (2008, pers. comm.) and support the initial effort.

Comparison of this delta set with ice margins re-constructed by De Simone (see field map on trip) enabled an approximate determination of the location of the ice margin at the time of each transition from one lake level to the next and lower lake level. The synthesis adds clarity to the picture of deltas deposited by tributary streams. The GlacioHudson River produced no AB I delta in the Glens Falls region because that outlet was still blocked by ice. The GlacioHudson River flowed south following the Kayderosseras Creek Valley and deposited the Milton Delta into AB I (Stoller 1922, Miller 1925, Chadwick 1928). The

Batten Kill has neither AB I nor AB II deltas because it was ice blocked for both those lakes and only opened after the lake level dropped to QS.

The ice margin reconstruction led us to examine Schock's (1963) surficial map of the Troy North quadrangle because the data indicated there may have been a lake level transition from AB I to AB II with an ice front somewhere in the quadrangle. Schock's map and interpretation of the Newtown Road and Ballard deltas came to light in a thesis that had long been forgotten by recent workers. Schock's observation on the possible change in lake level during deposition of the 2 ice contact deltas along the same ice margin added much to the emerging picture of the CV threshold.

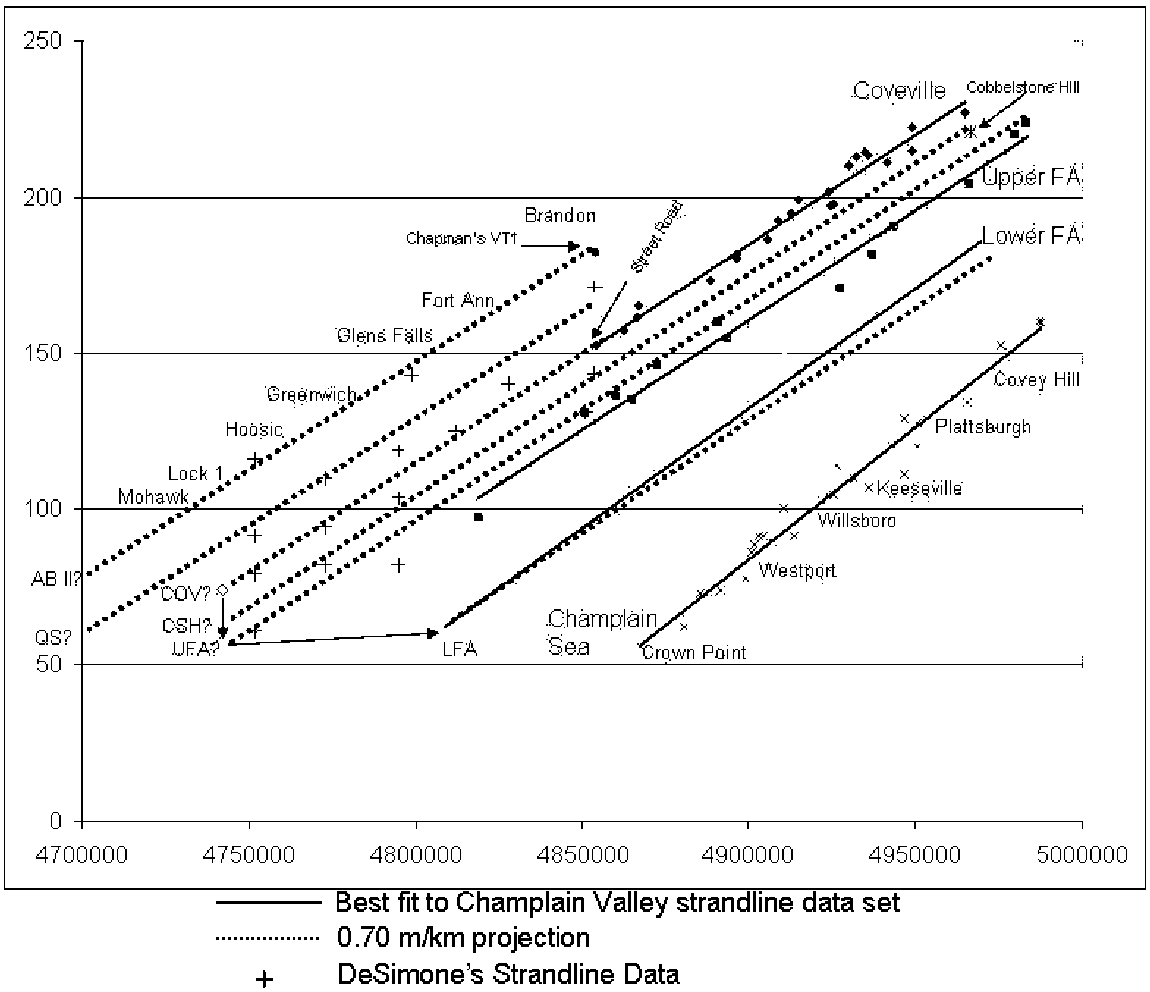

Figure 1: A plot of some of the northern Hudson through southern Champlain Lowland's strandline data combined with Chapman's data from the entire Champlain Valley completed by Rayburn and first seen in De Simone et al (2008).

LAKE COVEVILLE HALFMOON THRESHOLD HYPOTHESIS

Multiple data sources indicate there was an ice contact sediment dam across the Hudson Valley in the river reach extending from the Waterford Bridge north to Lock #1 of the H-C Canal. Halfmoon on the west and Speigletown on the east side of the valley have exceptionally extensive kame moraine deposits. This consists of subaqueous fan deposits of sand and gravel that piled up in places to form 2 deltas built into Lake Albany I and Lake Albany II on the Halfmoon side. The moraine extends as an arcuate deposit more than 5 miles long with fluvial sediment exceeding 200ft in thickness throughout much of the moraine. The moraine on the Speigletown side extends for more than 3.5 miles but its thickness is less well known. The form of the moraine on both sides of the valley is V-shaped with the narrow apex of the "V" in the vicinity of the Waterford Bridge over the Hudson River. This kame moraine represents the largest ice marginal accumulation of sediment from Albany to Glens Falls. Thus, it represents a deposit formed by a significant still stand or even a surge of the H-C ice lobe.

A plausible reason for an extended still stand lies with the longitudinal profile of the Hudson-Battenkill channel near Lock #1 where there is a buried nick point (Dineen 2008, pers. comm.). A step in the valley profile here would have provided a grounding line for the retreating H-C ice and enabled the ice to maintain this position. Schock (1963) reported in his investigation of the Troy North quadrangle that the Newtown Road and Ballard deltas, both ice contact deltas, were graded to different lake levels. Indeed, the higher Newtown Road delta is AB I while the lower Ballard delta is AB II. Schock dismissed the possibility that the lake lowered while the ice stood at this front. However, this may be the best explanation for the multiple kame delta elevations. A drop in lake level would have promoted grounding of the ice and ensured a continuous source of sediment to a stable ice margin via the dirt machine (Lowell, 2008). To the south, the Prospect Hill subaqueous fan extends to an elevation of 200ft and there are remnants of ice contact sediment in Pleasantdale on the east side of the valley. This would be consistent with the 200-220ft elevation dam spanning the Hudson necessary to retain Lake Coveville. The earthen dam possibly extended for a 2.5 mile stretch of the river from Lock #1 south to the Waterford Bridge. While holding back Lake Coveville, this earthen dam undoubtedly was continuously being eroded but it may have been substantial enough through its thickness to retain the lake waters for the 200 year duration of Lake Coveville.

Dave Barclay (2008 pers. comm.) notes that the lowered base level and grounding of the H-C lobe at the proposed Halfmoon threshold would likely have resulted in a surge or readvance of the ice. The basal lacustrines exposed at Lock #1 within the proposed threshold offer some support for this surge. There, thick basal varves contain numerous chunks of lodgement till that dropped, presumably in a frozen state from the ice into the accumulating lake bottom sediments. The sediments are very compacted and exhibit a "leaned upon" structure as described by LaFleur. Thus, there may be evidence the ice responded to the lowered lake level from ABI to ABII by surging slightly over the basal lacustrines at the proposed threshold.

Once Lake Coveville was established, Wall notes there would have been a difference in base level between the present path of the Mohawk River and its northerly distributaries, a complex of channels studied by Stoller (1911, 1916, 1918), La Fleur (1965, 1975, 1979) and others (Dahl, 1978, De Simone, 1977, Hanson, 1977). Wall notes in his description of the Cohoes Falls stop that a dam somewhere south of the IroMohawk distributary channels – south of Mechanicville – and north of the present Hudson-Mohawk confluence would have furnished a difference in base level between the present path of the Mohawk River and the other distributary channels that would have entered Lake Coveville. The lower base level below the CV dam at Halfmoon would have provided a hydraulic reason for the IroMohawk River to favor and dissect this pathway to the Hudson and abandon the northerly distributaries.

Historically, the Dutch name "Halve Moon" was applied to the large arcuate re-entrant of the Hudson River that exists today just below Lock #1. Historical descriptions of the Hudson River in the reach where the inferred Coveville dam existed offer interesting anecdotal data on the river's profile. There were rapids above Halfmoon Point (Waterford) and at Lock #1 but the river was deeper in between these two sections. This might reflect the buried step in the river profile or even a buried plunge pool at the step. The long history of the name Halfmoon indicates this is the most appropriate name for the hypothesized threshold for Lake Coveville.

DEGLACIATION OF LAKE GEORGE BASIN

Early Local Lakes. --- Chadwick (1928) discussed the retreat of the Lake George ice and most of the story he presented seems to hold true with the addition of a few details. Early ice retreat is marked by the series of kame terraces southwest of Lake George Village at 780ft, 740ft and finally at 590ft. Initially, drainage was probably across the divide above present Lake Vanare and into Glacial Lake Corinth (early Glacial Lake Warrensburg). Later, drainage was back against the ice front and southward against the H-C ice and the accumulating Glen Lake kame moraine. It is likely that during this interval of Lake George ice retreat, the H-C ice retreated from Corinth into the Luzerne Mountain gorge. As the H-C ice retreated from Corinth, the level of Lake Corinth/Lake Warrensburg would have dropped as the ponded water found a lower route along and/or under the H-C ice in the gorge. This would have abandoned the Corinth col and ended the Kayderosseras-Hudson flow. At some point in the H-C ice retreat, the 670ft Hartman terrace was deposited at the lower end of the Luzerne Mountain gorge. The Hartman terrace likely accumulated just prior to Lake George ice retreating enough to deposit the 590ft kame terrace that Chadwick attributed to a local Lake Caldwell.

Glacial Lake Caldwell And Subsequent Falling Waters To Glacial Lake Fort George. --- Continued retreat of the Lake George ice was probably accompanied by a falling water level as there are no kame terraces between 590ft and 420ft. Drainage had to be across the Glen Lake kame moraine and into the local Queensbury lake where the Queensbury kame delta accumulated. Meltwater from both the Adirondack-Hudson River and from Lake George ice likely contributed to the Queensbury delta at 470ft. Continued retreat of the H-C ice allowed the local Queensbury lake to become confluent with ABII and the Adirondack-Hudson River deposited the West Glens Falls delta. This last 440-420ft Lake Fort George of Chadwick was the last separate glacial lake in that basin.

Lake Bolton Confluent With Lake Quaker Springs. --- Eventually, Lake Fort George found an outlet through the Glen Lake kame moraine and deposited the sandy Patten Mills delta into Lake Quaker Springs. At this time, Chadwick’s Lake Bolton is actually confluent with Lake Quaker Springs in the Hudson Valley. Lake George ice then fronted Lake Quaker Springs as the ice retreated northward through the basin to just south of Ticonderoga. Lake Quaker Springs dropped to Lake Coveville with the ice in the northern end of the Lake George basin. Lake Coveville dropped to Lake Fort Ann and the waters in the Lake George basin became isolated from Lake Fort Ann. This scenario is only slightly different from the well illustrated story of Chadwick and owes much to Chadwick’s thoughtful analysis of the sequence of events.

LUZERNE MOUNTAIN GORGE STRATIGRAPHY AND DISCUSSION OF THE LUZERNE READVANCE HYPOTHESIS

The Luzerne Hypothesis. --- Connally & Sirkin (1969) initially outlined the Luzerne readvance hypothesis as an event that terminated in the Glens Falls, NY, region and occurred during the duration of Glacial Lake Albany. The Luzerne readvance hypothesis was more formally proposed 2 years later (Connally & Sirkin 1971) as a 20-35 mile readvance of the Hudson-Champlain lobe into Lake Albany. A “type section” was designated among a series of roadside exposures along Corinth Road west of Glens Falls within a steep gorge of the Hudson River as it exits the Palmertown Range.

Connally and Sirkin (1969, 1971) described an upper till overlying lacustrine sediments with a lower till deformed till in 2 exposures. A 17ft thick basal “gray-black bouldery, silty-clay till containing clasts of dark-gray, contorted lacustrine sediment” is overlain by a 4ft thick “thinly-laminated to thin-bedded sand”, in turn overlain by 5-12ft of “moderate-olive-gray till, very compact, very bouldery, with a sandy-loam matrix and many limestone and shale clasts,” finally topped by 4-10ft of “moderate-olive-gray till and colluvium overlain by spoil and vegetation.” To the east of this section, another section reports 6ft of till similar to the basal till above but underlain by 10ft of dark gray, laminated and rhythmically bedded lacustrine clay, silt and fine sand, greatly contorted. A third section still farther east reports 20ft of oxidized pebbly sand beneath the gray-black till with 4-20ft of light brown till containing a sandy matrix and angular boulders showing crude stratification. The 3 described sections span an approximate horizontal distance of 3000ft and elevations are provided. Connally and Sirkin proposed the upper brown till in their western and eastern sections represented a late Wisconsinan readvance they termed the Luzerne readvance.

In a search for exposures that may aid in understanding the Corinth Road stratigraphy, no new exposures were observed. Further, the old Corinth Road exposures are vegetated over but still reasonably accessible.

Prior Workers Did Not Interpret The Sections As From A Readvance. --- Previous workers did not attribute the observed sediments as representing a readvance (Chadwick 1928, Hansen et al 1961) and Woodworth (1905) dismissed the notion of a readvance in the stratigraphy of the Hartman terrace kamic gravels exposed a short distance to the west at considerably higher elevation and stratigraphically above the proposed readvance till. Woodworth also discussed a thick sequence of glaciofluvial gravel and sand with no apparent tills but the location of this exposure has never been reconciled with current exposures along the lower Luzerne Mountain gorge.

Hansen et al (1961) did not measure sections because the focus of their study was the decollement structure present between the lower lacustrine unit and the overlying till. They observed this at 3 exposures along Corinth Road and provided a cross section with exposure locations and elevations along a 2000ft stretch of Corinth Road. Hansen et al observed the basal lacustrines were rhythmically bedded and composed of dark gray clayey laminae alternating with light gray silty laminae. Several beds of moderately sorted medium to very fine sand occur within the rhythmites and ice rafted clasts were "plentiful." Above the lacustrines, they describe a compact, gray stony till with rounded to sub rounded clasts that grades upward to a yellowish brown, compact till at their outcrop #1. In their outcrop #2, they describe the upper part of the till as becoming loose and sandy, easily disaggregated. In their outcrop #3, the upper part of the till is described as silty, hard and compact. It should not be overlooked that Hansen et al consistently report their lower gray till as grading upward to a brown facies in all 3 exposures. The upper brown till facies is compact and silty in 2 sites but loose and sandy in their middle outcrop.

One problem emerges with Connally and Sirkin's correlation of their 3 exposures with those of Hansen et al. Connally and Sirkin compare the till fabrics of their proposed upper readvance till with the brown facies fabrics examined by Hansen et al. The problem here is that we do not believe Hansen et al observed the true upper diamicton facies that Connally and Sirkin propose as their readvance. The readvance diamictons are stratigraphically above a lacustrine sand unit that separates the lower and upper diamictons. Hansen et al did not discuss any interbedded lacustrine sand unit between their lower gray till and upper brown till. Hansen et al never looked high enough in the exposure to observe the sand unit and the overlying diamicton. This was not their focus. They reported fabrics in the lower gray facies and upper brown facies all within the same lower till unit observed by Connally and Sirkin. So, the comparison of till fabrics reported by Connally and Sirkin is invalid. Further adding to the problem is it is not clear these workers observed the same 3 exposures as there are at least 6 exposures along Corinth road and the location data for both Connally and Sirkin and Hansen et al are inadequate to verify which exposures match each other.

Subsequent Workers Suggest Alternative Interpretations. --- Bob La Fleur took students to the site to measure, describe and interpret the Corinth Road sections as part of a field lab in glacial stratigraphy. I was among those students as were some of the trip participants. We cleaned the type sections, measured, described and collectively decided there was insufficient evidence of a readvance. The type sections can be alternatively interpreted.

The best efforts of these students have been synthesized here to offer a more complete picture of the entire Corinth Road stratigraphy. Cro (1984) examined the stratigraphy, texture, fabric and mineralogy of 3 exposures as part of a thesis project. Both LaFleur and De Simone assisted Cro during the early stages of his field project in the summer of 1983. Cro dismisses any comparison of fabrics in the brown till facies as invalid due to the alteration of till fabric in the ablation environment. Thus, Connally and Sirkin's assertion that the E-W oriented fabric in their brown proposed readvance till is not valid if this till is an ablation till or sediment flow deposit. However, Cro compared the gray lower and brown upper facies that Hansen et al described in their outcrop #3 and did not observe the same upper diamicton of the proposed readvance. So, there is no direct comparison of textures possible.

Both Schuster (ca 1986) and Hixon (1988) carefully measured the same easternmost exposure observed by Hansen et al and by Connally and Sirkin. The difference between the observations of the earlier workers and the later workers is that the later workers could not observe the lower portion of the lower gray till of Hansen et al nor the underlying deformed lacustrines. This is likely due to slumping and cover of the lower portion of these exposures. The descriptions of Schuster and Hixon are in close agreement. This adds confidence to the more complete study of Hixon who measured and described 5 sections along Corinth Road. LaFleur's notes on the location of these 5 sections further aids in their map location and in the correlation of Hixon's 5 exposures with the exposures observed by all earlier workers. **Figure 2** presents a synthesis of the 6 total stratigraphic sections described by Hixon, Schuster, Cro, Connally and Sirkin, and Hansen et al. Connally and Sirkin's middle exposure is not shown as there is some uncertainty as to exactly where it fits with the other measured exposures.

Corinth Road Stratigraphic Units of Figure 2. ---
Hartman Gravel: Glaciofluvial sand and gravel of the 670ft Hartman terrace as originally discussed by Woodworth (1905) and later by Chadwick (1928). The sediments are horizontally bedded brown sands and gravels coeval with the lower level of Glacial Lake Corinth. These deposits represent a fluvial gravel and sand unit that accumulated in a kame terrace setting. Exposures are now insufficient to confirm if the gravel is predominantly sourced from the H-C ice or is inwash from the Adirondack-Hudson River. The Hudson-Champlain lobe dammed the lake at this level (Chadwick 1928).

Hartman Till: Brown, highly compacted silty diamicton with approximately 10% clasts ranging from angular to rounded but less well rounded than the clasts in lower till units. Cro counted clasts and determined their lithologies. Adirondack gneiss in the brown tills average 56% of the clasts versus 30% in the gray tills. Shale derived from the Hudson-Champlain lowland averages 27% in the brown tills versus 49% in the gray tills. The matrix contains abundant frosted quartz grains with minor garnet, micas and mafic minerals. The lower contact of this till with the lacustrines shows no deformation, only a truncation surface.

This till is interpreted to represent the late Wisconsinan lodgement till. An alternative interpretation is that this till could be from a readvance of the H-C lobe. If this was a readvance till and not the lower diamictons proposed by Connally & Sirkin, then the Hidden Valley moraine could be the limit of the readvance in the southeastern Adirondacks. However, in view of the lack of evidence elsewhere for the proposed readvance of Connally & Sirkin as discussed below and by De Simone (2008) and De Simone et al (2008), it is *not* preferred to consider a readvance as the best interpretation.

Luzerne Lacustrines: Below the Hartman till both Hixon and Cro measured sections composed of interbedded silty clay, silt and sand, horizontally bedded, brown in color, and undeformed. In the 2 nearby sections measured, there was a brown, compact, sandy silt matrix diamicton interbedded with the finer grained lacustrines. The nearly 5 foot thick diamicton does not incorporate any of the underlying lacustrine sediment into its base and there is no deformation at the lower contact. This diamicton is interpreted to represent a sediment flow deposit that originated on the steep upper slopes above the site. The lateral extent of the diamicton is unknown but is shown on the cross section as a discontinuous lens.

Hixon Till: This is the till measured by Hixon in his exposure #3 and is correlated to be the same section seen by Hansen et al as their outcrop #1. Hansen et al describe this till as a dense, compact, gray stony till that grades upward into a dense, compact brown stony till. Clasts range from rounded to angular. Hixon notes the base of the till appears to incorporate gray silt and clay from the underlying lacustrine unit. Hansen et al discuss the folding and thrusting observed along the contact between the till and the underlying lacustrine sediments. The gradational contact between the lower gray facies and the upper brown facies as described by Hansen et al and shown by Hixon indicates the gray and brown portions are facies of a single till unit. Rounded clasts are more abundant in the gray till facies while angular clasts are more abundant in the brown till facies as noted by Cro at other sections. Cro also observed the fabric differences between the gray and brown till facies and decided they were the result of the ablation process altering the brown till fabric during deposition. The brown facies incorporates more Adirondack lithologies and angular clasts as the sediment comprising the upper till facies is derived from sediment within and on the surface of the glacier. In contrast, the lower gray till facies contains more rounded clasts and more Hudson Valley lithologies as the sediment comprising the lower till facies is primarily derived from at or near the base of the glacier.

The age of this middle till unit is problematic. Its extent along Corinth Road is also unknown. The unit was observed and measured at only one section but clearly lies at a very different elevation from the till units above and below it. The Hixon till is nearly 25 ft thick where measured and the underlying lacustrine sediments are sheared along the contact with the till. Further, the underlying lacustrines are highly compacted. It is inferred the Hixon till may be middle Wisconsinan or older in age but this based solely upon stratigraphic considerations and no other supporting data.

<u>Cro Lacustrines:</u> Cro's exposure #2 contains no till, only a thick sequence of interbedded silt, clay and sand that is highly compacted and brown in color. The clays display incipient fissility and the sands support undercutting as evidence of their dense structure. These are correlated to exposures to the east based upon stratigraphic position and elevation and similar sediment textures. Hixon's exposure #2 and his exposure #1, the latter examined in even greater detail by Schuster, both contain a diamicton unit in addition to lacustrine sediments. Schuster described graded bedding in the interbedded sands of the deep water Cro lacustrines and noted the presence of flutes or load casts on the undersides of sand beds. Thus, the graded sand beds represent turbidites deposited in very deep water.

Schuster did not observe the diamicton at the top of the easternmost exposure, #1 of Hixon. Hixon described the diamicton as a poorly sorted brown sandy and gravelly unit, consistent with the upper or readvance till of Connally & Sirkin's exposure #3. Neither Hixon nor Connally & Sirkin reported any deformation at the contact between this diamicton and the underlying turbidite sands. Hixon's exposure #2 is correlated with Connally & Sirkin's exposure #1. The upper diamicton above a thin but similar deep water sand unit shows no deformation at the lower contact. Here, however, the diamicton is comparatively compact, brown in color and more silty than to the east. These two upper diamictons are the proposed late Wisconsinan readvance tills of Connally & Sirkin. However, their stratigraphic position, elevation and the nature of the sediment and lack of lower contact deformation suggests an alternative explanation. It is inferred here that both these diamictons are sediment flow deposits composed of reworked till and lacustrines from higher up on the slopes that were carried into the deeper lake waters contemporaneous with deposition of the Cro lacustrines. It is also possible the easternmost relatively uncompacted diamicton represents the youngest unit along the entire Corinth Road section and that it was deposited during the Holocene by sediment flow or slump processes. The lateral extent and thickness of both of Connally & Sirkin's proposed readvance deposits is unknown. It is certain, however, that they occur at an elevation that may best place them into the Cro lacustrine unit.

<u>Hansen Till:</u> This is the lower till observed most frequently by past workers. It was visible at Hixon's exposures #2 and #1, Hansen et al's outcrop #2 and #3, and Connally & Sirkin's #1 and #3. Connally & Sirkin's exposure #2 cannot be matched to others along the Corinth Road section. Thus, it is left out of this discussion but its possible location is indicated on the cross section.

The Hansen till consists of a lower gray facies and an upper brown facies, similar overall to the Hixon till discussed above. Cro discussed these as his upper till C and lower till D. The lower gray facies is dense, compact and enriched in rounded shale clasts. Cro determined this gray till facies contained 49% shale clasts and 30% gneiss clasts. X-ray diffraction analysis of the clay fraction by Cro determined that the gray facies contained more illite than the other amphibole enriched brown till facies.

The lower contact between the Hansen till and the underlying basal lacustrine unit shows shear deformation with folding and thrust faulting consistent with emplacement of the till as a lodgement facies. Hixon noted the lower gray facies appeared to incorporate gray clay and silt from the underlying lacustrine unit.

The contact between the lower gray facies and the upper brown facies was noted as gradational by both Hansen et al and Hixon. Overall, the lodgement gray facies contains approximately 30% clasts while the upper brown ablation facies contains approximately 15% clasts. Cro interpreted the gray and brown facies to represent a lodgement-ablation till couplet. Indeed, as discussed for the Hixon till above, the differences in clast lithologies, clay mineralogy, clast angularity and fabric of the gray and brown units is best explained by them representing a facies change. A basal lodgement till composed predominantly of more rounded clasts of more valley shale lithology with high illite content changes in a gradational manner to an upper facies predominated by more angular clasts with more Adirondack gneiss lithology and a higher amphibole content versus illite.

Basal Lacustrines: Only a thin layer of these black and gray horizontally bedded silt and clay rhythmites was discussed by Hansen et al beneath their decollement structures at the 2 easternmost and lowest elevation exposures along Corinth Road. Hixon did not observe the basal lacustrines as they were likely covered by slumping during the years between their observations. Hansen et al describe the rhythmites as composed of dark gray clay rich laminae alternating with yellowish gray silt rich laminae. Some interbeds of medium to very fine sand also were noted. Ice rafted clasts were cited as plentiful and there was deformation around the clasts due to their emplacement into the deep waters of the lake. The upper portion of the basal lacustrines showed evidence of shear deformation due to ice override and deposition of the overlying Hansen till.

Correlation Of Western and Eastern Luzerne Mountain Gorge Stratigraphic Units. --- LaFleur (1990-91, private files) examined data from numerous borings drilled to evaluate a site in Corinth, NY, at the western end of the Luzerne Mountain gorge. His interpretation of the stratigraphy in Corinth is summarized in **Figure 3.** LaFleur recognized 3 tills and interbedded lacustrines with a basal fluvial gravel unit over an elevation range of more than 200ft. We propose the multiple till sequences along the western and eastern ends of the Luzerne Mountain gorge are correlative and consistent in their preservation of 3 tills. The uppermost till in both locations is interpreted to be the late Wisconsinan till. Wider correlations may be possible as the underlying gray till and "black & white" lacustrines along the Luzerne Mountain gorge are similar in character to the Hell Hollow till and related lacustrines at West Milton, NY (LaFleur 1975, 1979). At West Milton, an erosional unconformity separates the older Hell Hollow till from the upper late Wisconsinan or Mohawk till.

Additional Data Does Not Support The Luzerne Readvance. ---
Detailed surficial geologic mapping reveals no evidence of a readvance throughout the area where it is proposed; A readvance of 30 miles is proposed but no evidence of overridden sediment has been cited. Detailed surficial geologic mapping over the entire area of the proposed readvance in southern Washington County revealed no observed evidence to support a readvance. Connally & Cadwell (2002) state the readvance "forced drainage eastward into the Batten Kill drainage system" and state "there is a plethora of stagnant ice features probably left by wasting ice from the Luzerne readvance." These statements are not supported by the results of detailed surficial geologic mapping (De Simone 2006, 1992, 1985, 1977; De Simone & La Fleur 1986, 1985; De Simone & Newton 1994; Dethier & De Simone 1996; Dineen et al 1988). The distribution of glacial deposits tells a very different story, one of systematic stagnation zone retreat of an active Hudson-Champlain ice lobe from the northern Hudson lowland and from the adjacent Taconic foothills east of the lowland. Numerous exposures were studied during this extensive mapping effort and all available well logs were examined. None of the data support readvance of glacial ice.

Problems with the hypothesized extent of the Luzerne readvance; Connally & Sirkin (1971) stated the extent of recession prior to readvance can only be estimated. However, they cited one piece of evidence, that Woodworth (1905) inferred the lake sediment on the floor of Wood Creek showed signs of being "ice worn." Wood Creek is approximately 12-13 miles east of the proposed readvance type section. Detailed mapping of the Wood Creek area indicates the channel is part of the Fort Ann Outlet Channels complex. As such, the sediment on the floor of the channel was scoured, not by ice, but by outflow from Lake Fort Ann. Based upon this one citation, the Luzerne readvance was hypothesized to have an extent of 20-35 miles.

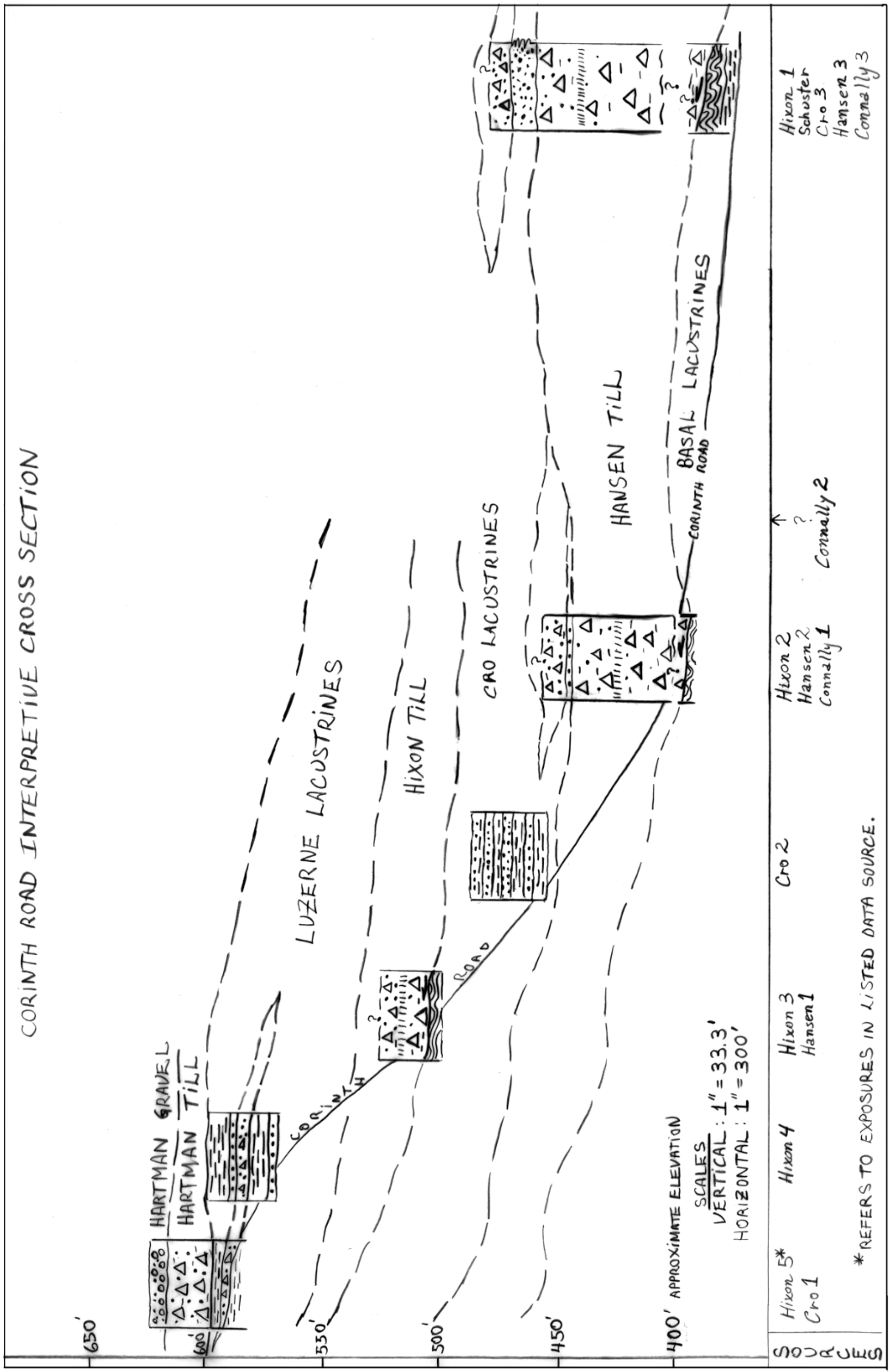

Figure 2: A stratigraphic synthesis of the Corinth Road exposures indicates there may be as many as 3 tills separated by lacustrines over an elevation range of nearly 300ft. The uppermost or Hartman till is inferred to be late Wisconsinan and is not interpreted to represent a readvance or surge of the H-C lobe. The lacustrine units contain diamictons interpreted to represent sediment flows or slumps.

Luzerne Mountain Gorge

W Corinth Section	Corinth Road Section E
SG-S-C Unit III	Hartman Gravel
Till III	Hartman Till
Clay-Silt Unit II	Luzerne Lacustrines
Till II	Hixon Till
Clay-Silt Unit I	Cro Lacustrines
Till I	Hansen Till
Basal Gravel	Basal Lacustrines
LaFleur, 1990-91, private files	De Simone & LaFleur, this volume

Figure 3: Proposed correlation of the Luzerne Mountain gorge stratigraphy along Corinth Road on the east with the borings data from Corinth on the west. The agreement between the west and east ends of the gorge suggests this region of the Adirondack Hudson Valley may sporadically preserve evidence of older glaciations.

Glaciological problem with the hypothesized readvance; As most recently depicted (Connally & Cadwell 2002), the Luzerne readvance of Hudson-Champlain ice covered much of Washington County, NY. Ice overtopped Taconic upland summits exceeding 1400ft at Willard Mountain immediately adjacent to the Hudson lowland and overtopped high Taconic Range summits along the NY-VT border with elevations above 2000ft. The readvance stopped along the western crest of Mt. Equinox (3840ft) and Mother Myrick Mtn. (3361ft) west of Dorset & Manchester, VT. This Taconic summit ridge exceeds 3000ft in elevation at several locations.

However, the readvance is not postulated to have brought ice up the Mettawee River valley oriented approximately parallel to ice flow. The present drainage divide between the north-flowing Mettawee River and south-flowing West Branch Batten Kill occurs in Dorset village at an elevation of only approximately 930ft.

Wouldn't readvancing ice have flowed up the Mettawee Valley and into the Vermont Valley through the Manchester region? De Simone's detailed surficial geologic mapping and examination of subsurface data in Dorset (2007), Manchester (2005, 2004) and Arlington (2001, De Simone and Baldivieso 2001) , VT, does not support a Luzerne readvance.

Summary of Arguments Against The Luzerne Readvance

*The type sections do not expose an acceptable readvance diamicton. The readvance diamictons have here been interpreted to be sediment flow or slump deposits within lacustrine units in a thick 3 till sequence.

*The proposed extent of the readvance is not supported by mapped evidence.

*The proposed extent of the readvance leaves out areas that would have been covered by advancing ice; These areas reveal no evidence to support a readvance.

*The timing of the proposed readvance has not been matched to the paleoclimatic record.

The hypothesis should be abandoned.

SUGGESTED CRITERIA FOR A READVANCE PROPOSAL

Our current understanding of the glaciological settings for ice surges coupled with a well documented paleoclimatic record indicates that any readvance hypothesis should meet certain minimum standards of evidence. A distinction should be made in the hypothesis between a surge confined to a valley and likely the result of local glaciological conditions and not driven by significant climatic changes versus a larger scale readvance. The larger readvance would have ice override both valley and upland sites and would be driven by significant regional or global climate change.

A set of criteria are suggested for future readvance hypotheses (De Simone 2008). Also, these same criteria might be applied to other previously proposed readvance hypotheses in an attempt to determine if these hypotheses are valid. Here are these criteria:

*Deformed sediment: A readvance diamicton should contain deformed sediments derived from the deposits beneath the advancing ice. This would include folded and contorted beds of any underlying lacustrine sediments if the ice was readvancing into a glacial lake basin. Ripped up clasts of deformed sediments might be mixed with a diamicton facies resembling basal till. A deformed till unit may be present that would represent a true readvance till. This till would likely contain fragments of the underlying sediments advanced over as recognizable or barely recognizable facies within the till.

*Evidence of ice override along the basal contact: The basal portion of the readvance diamicton should contain structures indicative of emplacement by advancing ice. The contact with the underlying sediment must also show deformation associated with ice override.

*Mappable extent: The readvance diamicton should be aerially widespread and be identifiable as a mappable unit of deformed till. A readvance diamicton identified from only one location might be of only local significance and should not be used to hypothesize a widespread readvance.

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TRIP STOPS

1. Crown Point CV Delta:

Location: Crown Point quadrangle. Proceed along Rte 9N from north of Crown Point to the intersection with Street Road. The main entrance to this pit is on the west side of the road. However, if you loop back and turn west toward the town transfer station, there is another upper entrance to this pit. The main or lower entrance is preferred for easy parking of large vehicles. Note that you can drive up to the high level of the pit from below along the access road in the north face of the pit. It is a wide and safe route but not for a bus.

Parking: N 4861517, E 0625137, lower or main entrance parking area.

Ownership & permission: Gerald Huestis is the owner and the pit is operated by his 2 sons. Mr. Huestis does not need prior arrangement for access to his excavation but it is good to call and introduce yourself. 518-623-3671.

Description: The lower portion of the pit exposes the bottomset sand beds of the delta. The upper portion of the pit beautifully reveals the topset-forest contact at 525ft within interbedded gravels with some sand. There is no ice contact deformation evident in the exposure and supports the notion this is an open water delta of the small brook exiting the Adirondacks. The valley wall portion of the Street Road delta has numerous kettles and suggests the delta buried recently stranded ice.

Discussion: Rayburn (2004) mapped Champlain Valley strandlines as far south as Ticonderoga, NY, and Brandon, VT. On the New York side of the valley, no strandlines higher than Lake Coveville were found; across the valley in Vermont, there are higher deltas in Brandon, VT. The highest one is Chapman's (1937) Forest Dale delta that was originally thought to be CV but it is better correlated with QS at 560-570ft. The northern termination of QS strandlines perhaps at the Fernville kame moraine at the north end of Forest Dale suggest that this may have been the ice margin position when the lake dropped from QS to CV. It is interesting that there is no QS delta or obvious kame terrace above the Street Road delta. But, the vicinity is tightly constrained by the surrounding Adirondack Mountains where the brook spills out onto the lowland and there may have been little or no time passage between ice recession to the kettled area next to the valley wall and the drop from QS to CV.

However, south of the Street Road delta on the Ticonderoga and Putnam quadrangles, there are lake sediments that correlate to a higher lake and suggest the ice margin was near or at the Street Road delta at the time QS dropped to CV.

Just north of the Street Road delta is are a complete series of well defined CV and FA strandlines in the upland west of the Town of Crown Point. Most of these strandlines were recognized by Connally and Cadwell (2002), but correlated to different levels than stated in Rayburn (2004). A small delta fan at 525ft (161m) is at the Coveville level. Terraces at 440ft (135m), 350ft (108m), and 185ft (56m), are at the highest Fort Ann, lowest Fort Ann, and highest Champlain Sea levels respectively (Rayburn, 2004). The Town of Crown Point is built on the highest Champlain Sea delta terrace. The foreset-topset contact in the Coveville level Street Road delta is at about 525ft (161m). These strandline measurements are in good agreement with the 0.70 m/km isostatic rebound estimate for the entire valley data set.

By projecting this rebound trend southward into the Hudson Valley we hope to unravel the earlier deglacial story hidden among the Lake Albany strandlines, lacustrine sediments, and ice marginal deposits.

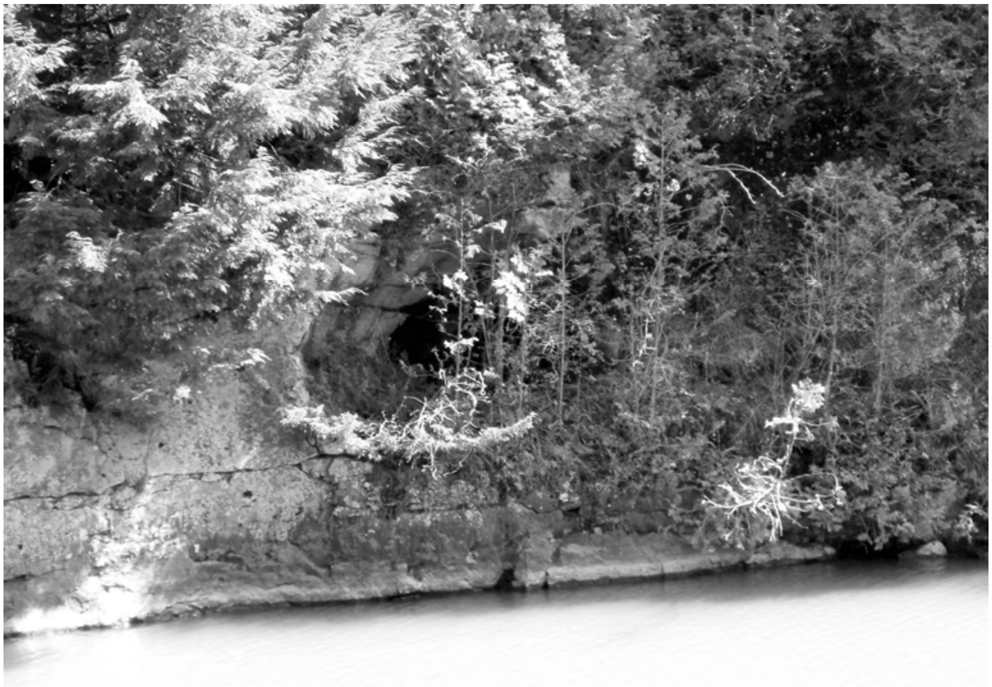
Topsets over Foresets in Street Road Delta.

Bottomsets of Street Road Delta.

2. Dresden Station Subaqueous Fan:
Location: Putnam quadrangle along Rte 22.
Parking: N 4836666, E 0627970 at entrance to the pit.
Ownership & permission: Richard Dedrick Trucking & Excavating of Putnam Station, NY. Call Richard @ 518-547-8432.
Description: Striated and grooved bedrock ridges with steep flanks are capped with subaqueous fan gravel and sand beds that fine upward to lacustrine sand. The top of the deposit has been terraced at an elevation of 190ft, likely a FA terracing event. There is no till atop the bedrock.

3. Chapman's Potholes:
Location: Fort Ann quadrangle on Flat Rock Road, a small loop of old Rte 4 between the H-C Canal and present Rte 4. The junction with Rte 4 is approximately 1.5mi north of the Rte 4 – Rte 149 junction in Fort Ann Village. The road is between the village of Fort Ann and the Rte 22 junction north of the village.
Parking: N 4809487, E 0624274; park at the edge of the railroad tracks and proceed across the tracks and up onto the outcrop to the south. CAUTION ! Poison ivy and oak present.
Description: There is a pothole visible in the opposite wall of the canal. Also, there appear to be potholes present on the top of the outcrop in the brush near the south end of the outcrop and along the south face of the outcrop. These potholes have been largely covered by organic litter.
Location: Fort Ann quadrangle on the north side of Rte 4 approximately 0.4mi south from the junction with Flat Rock Road. *This is a very heavily traveled 55mph passing zone along Rte 4 and* CAUTION *is urged!*
Parking: N 4809022, E 0623199; park off the north side of the road where there is a little used turnoff into the grassy area.
Description: There are 4 large potholes visible in this outcrop of granitic gneiss. All are semi-circular in form and extend down below the present graded ground surface.
Discussion: Chapman (1937) described the location of numerous potholes along the Hudson-Champlain Canal north of the village of Fort Ann. These potholes have been interpreted to be the result of the outflow of Lake Fort Ann as that discharge passed through the primary Fort Edward channel of the Fort Ann outlet channels. More specifically, the potholes are concentrated in the area where the channel is constricted by the steep granitic gneiss flank of Battle Hill to the north and the predominantly carbonate bedrock platform to the south. Pothole formation has been attributed to mechanical abrasion by stream bed load into the rock bottom of a channel. Eddying in very powerful flow may be the mechanism of formation for the potholes observed in this Fort Ann outlet.

Pothole on Flank of Battle Hill in Gneiss



Pothole in Canal Wall

4. Glen Lake Kame Moraine – Kame Terrace:

Location: Putnam Mountain quadrangle. Proceed approximately 2.8mi east on Rte 149 from the junction of Rte 9 and Rte 149, near I-87 Exit 20. Continue past the Bay Road junction east for an additional 1.4mi to the Rte 9L junction. From this point, there are several gravel and sand excavations with access from Rte 149. The first is approximately 0.8mi to the east with another smaller pit in sand another 0.1mi east behind the self storage facility. Continue along Rte 149 from the Rte 9L junction a total of 1.6mi to Patten Mills Road and turn south on this road. Proceed 0.8mi to the entrance of the Jointa Galusha excavation. There is another excavation adjacent to the Washington County closed pit on Tripoli Road. This excavation can be accessed by continuing on Rte 149 to the next intersection with Tripoli Road on the south and Hadlock Pond Road on the north.

Parking: N 4805293, E 0613466 at the entrance gate. After checking in at the scales office, drive to the left and park before the sorting operation if it is busy or drive down to the edge of the pond at the bottom of the pit if there is no activity.

Ownership & permission: Jointa Galusha, John Davidson @ 5158-792-5029.

Description: Bedrock ridges in the pit trend approximately 070° and striations. Although faint in the gray gneiss and black amphibolite, average 220° and range from 210-230°.So, the H-C ice lobe flowed subparallel to the orientation of the bedrock ridges in this area during at least the later thin ice stages.

In the bottom of the excavation, well sorted sand and well sorted gravel occur. The sand beds show some ripples and deformation likely the result of loading and de-watering of the sediment. Above, there are interbeds of gravel and sand. Some cobble gravel beds are free of matrix Clast supported gravel beds might exhibit grading upward to sand beds. The sequence is repeated a few times in these thick beds. Near the top of the excavation, sand beds predominate and the sediment appears more lacustrine. Small dunes are evident on the top surface of the feature.

Discussion: The Glen Lake kame moraine-kame terrace is an extensive deposit that covers portions of 4 quadrangles – Putnam Mountain, Hudson Falls, Lake George and Glens Falls. The deposit is the most extensive ice marginal accumulation of sediment north of the Hudson-Mohawk confluence. Likely, the H-C lobe maintained a position here along its western-northern lateral margin while the front of the lobe and eastern margin retreated rapidly. This was possible due to the Glen Lake area being close to the ice source coming from the Champlain Valley. The long duration of the ice margin here would account for the extensive and thick accumulation of sediment.

The kame terrace generally tops at approximately 520-490ft along its entire length of approximately 14 miles to the base of the Luzerne Mountains. It is a recessional moraine and there is no evidence of readvance observed within exposures as of this writing. It is very possible the deposition in this deposit was time transgressive with some portions of the moraine accumulating before other portions. However, the moraine is apparently graded to a persistent base level along what was the northern and western edge of the H-C lobe. An ice marginal lake may have persisted in the area of the kame moraine, a lake dammed by the ice and extending northward into the Lake George basin and adjacent basins. If true, then 500ft would be an approximate maximum elevation of the highest level of Glacial Lake George.

Ice recession from the base of the Palmertown Range to the south allowed the GlacioHudson River to have its first access to the valley and it deposited the West Glens Falls open water delta into AB II. Eventual ice recession from the kame moraine allowed Glacial Lake George to lower and this outflow deposited the Oneida Corners fan delta into Lake Quaker Springs. Contemporaneously, the GlacioHudson River built its major Glens Falls delta into QS.

Ice margins (De Simone, 1985, De Simone & La Fleur, 1985, 1986) through this area are similar to those of Chadwick (1928) and La Fleur (1979).

5. Corinth Road Multiple Till Exposures:

<u>Location;</u> Proceed west along Corinth Road from Exit 18 of I-87 approximately 3 miles, past the Queensbury sewage treatment plant on the south side of the road. Continue another ½ mile past roadside outcrops of bedrock at the foot of the Luzerne and Palmertown Mountains. As the road grade begins to rise, look for any of several parking areas along the north side of the road where the overburden exposures are visible. It’s a good idea to drive all the way to the top of the hill, past the house on the bluff and note the highest exposure below the house. This is the Hartman terrace. You can turn around and return to a convenient pullout and park to examine the exposures.

<u>Parking;</u> The best parking area is about ½ of the way up the hill where there is a large pullout that still has crumbly asphalt and plenty of room.

<u>Ownership and permission;</u> You only require permission to access the bluff exposure below the house at the elevation of the Hartman terrace.

<u>Description and Discussion;</u> Please see the text for a thorough analysis of this site. Note that the exposures have been overgrown for many years and may not be in good shape at the time of your – or our – visit.

TRANSPRESSIONAL DEFORMATION IN TACONIC SLATES AND ITS RELATION TO BASEMENT ARCHITECTURE

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INTRODUCTION

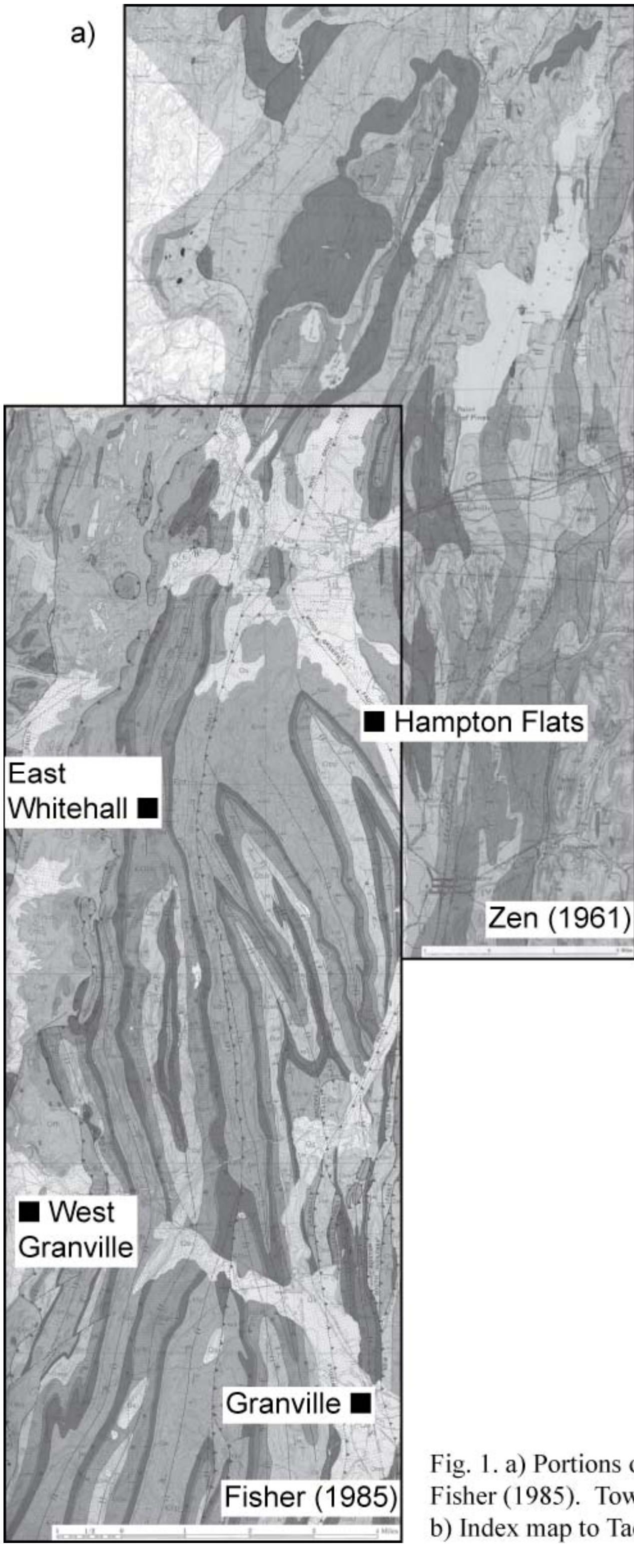
The Taconic allochthon is a north-northeast-trending tectonostratigraphic unit in westernmost New England and easternmost New York State. The allochthon, which consists of Neoproterozoic rift clastics, Cambro-Ordovician slope-rise deposits, and Ordovician flysch, was emplaced during the Middle to Late Ordovician Taconic orogeny (Rowley and Kidd 1981, Stanley and Ratcliffe 1985, Karabinos et al. 1998). The strata were folded and cleaved during orogenesis. In the northern Giddings Brook thrust sheet, the penetrative deformation, low-grade metamorphism, and dominantly fine-grained nature of the strata resulted in the formation of slate that has been extensively quarried (Dale 1899).

Nearly thirty years ago, Rowley et al. (1979) led a field trip for the New York State Geological Association through part of the slate belt in the northern Giddings Brook thrust sheet. The mapping presented in that field guide and the bedrock geologic map of the Glens Falls-Whitehall region compiled by Fisher (1985) show that the folds and slaty cleavage in the section of the Giddings Brook thrust sheet between Granville/West Granville and Hampton Flats/East Whitehall are not parallel to the overall trend of the Taconic allochthon (Fig. 1). Although the allochthon trends north-northeast, the folds and slaty cleavage in this area trend north-northwest. The primary goal of this field trip is to examine evidence supporting the hypothesis that the strata in this area underwent transpressional deformation. The field trip (Fig. 2) will begin in the classic area mapped by Zen (1961) to the north of Hampton Flats where the folds and slaty cleavage trend north-northeast. Inspection of the style of deformation of the strata to the north of Hampton Flats will provide a framework for understanding that of the strata in the area between Granville/West Granville and Hampton Flats/East Whitehall. Continuation of the field trip to the south to examine the area of anomalously north-northwest-trending folds and slaty cleavage will include stops that permit discussion of not only transpression but also the magnitude of tectonic volume change in slate, which remains a controversial topic.

STRUCTURAL DOMAINS

The northern Giddings Brook thrust sheet is divided into two structural domains on the basis of the (1) trend of the regional-scale D_2 folds (F_2), (2) strike of slaty cleavage (S_2), (3) dip of slaty cleavage (S_2), (4) rake of the mineral lineation (L_2), (5) state of strain, (6) strain symmetry, and (7) magnitude of finite strain. The last three characteristics refer only to the strain related to S_2 formation and are determined primarily from strain fringes around pyrite framboids and subspherical carbonaceous material. S_2 was superposed on the F_2 folds during the final stages of fold development and is approximately axial planar to the folds. Thus, the characteristics used to define the structural domains reflect only a portion of the deformation history undergone by the strata. D_2, however, is the main deformation event in the northern Giddings Brook thrust sheet, and S_2 is the dominant cleavage in the strata.

Domain 1 lies to the north of Hampton Flats, and domain 2 lies between Granville/West Granville and Hampton Flats/East Whitehall. Exposures that display characteristics of both domains 1 and 2 also exist, and these define a minor structural domain referred to as domain 1/2. Domain 1/2 lies on the eastern edge of domain 2 and is not considered on this field trip.



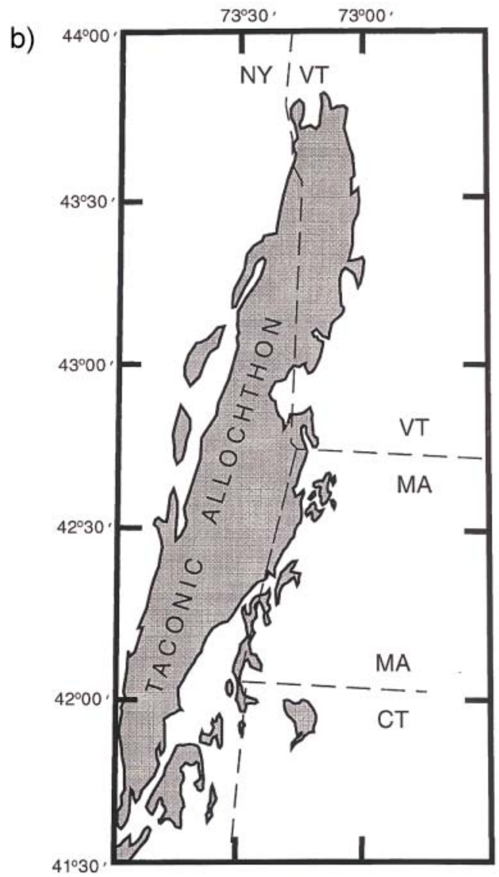


Fig. 1. a) Portions of bedrock geologic maps of Zen (1961) and Fisher (1985). Towns referred to in text indicated by squares. b) Index map to Taconic allochthon.

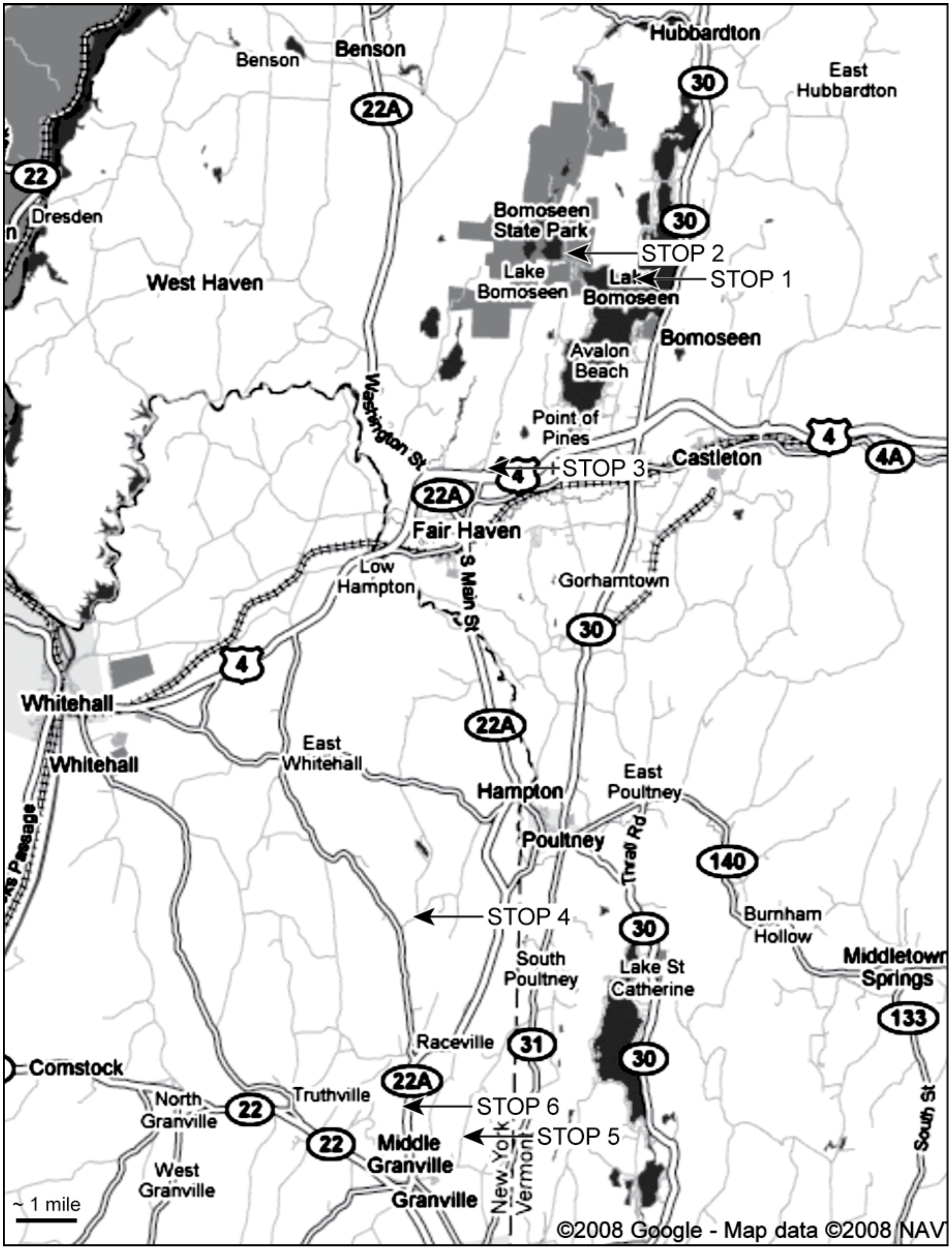


Fig. 2. Road map of area traversed during field trip. Stop locations shown by arrows.

Domain 1

Domain 1 is characterized by north-northeast-trending F_2 folds. The mean strike of S_2 is also north-northeast, and the mean dip is about 30°E. The mean trend of L_2 is east-southeast such that the rake of L_2 on S_2 is about 90°. Fibers in strain fringes are curved in thin sections oriented perpendicular to S_2 and parallel to L_2. The sense of curvature is consistently clockwise when the thin sections are viewed to the north-northeast and the fibers are traced toward the core object. This indicates top-to-the-west-northwest non-coaxial flow, which is consistent with the west-northwest vergence of the F_2 folds. In thin sections oriented parallel to S_2, the trace of the fibers in the strain fringes is straight and parallel to L_2. This evidence for plane strain in combination with the curved shape of the fibers in S_2-perpendicular and L_2-parallel thin sections gives a monoclinic symmetry for the strain related to S_2 formation. The strain magnitude is quantified using the length of the fibers in the strain fringes. Fibers were measured and $1 + e_1$ was determined for five sites. Site means for $1 + e_1$ range from 2.1 to 2.4, and the mean value for $1 + e_1$ for the five sites is 2.3. For additional information on domain 1, see Chan (1998).

Domain 2

Domain 2 is characterized by north-northwest-trending F_2 folds. Some folds can be traced continuously to the north and south of domain 2 where the fold axis changes trend from north-northwest to north-northeast. Both the strike and the dip of S_2 differ from that observed in domain 1. The mean strike of S_2 is north-northwest, and the mean dip is about 45°E. L_2 also differs in orientation, the mean trend being southeast. The change in orientations of S_2 and L_2 results in a rake of L_2 on S_2 of about 50° from the south-southeast. Like domain 1, fibers in strain fringes in S_2-perpendicular and L_2-parallel thin sections are curved and the sense of curvature is consistently clockwise when the thin sections are viewed to the north-northeast and the fibers are traced toward the core object. In S_2-parallel thin sections, the long axis of the strain fringes is parallel to L_2. At some sites, the strain fringes bracket the core object and show no evidence of fiber growth perpendicular to the long axis of the strain fringes. At other sites, the strain fringes completely surround the core object, and the fibers display radial growth. The former morphology indicates plane strain and the latter flattening strain. In addition, at several sites the fibers are curved in S_2-parallel thin sections. The sense of curvature is consistently clockwise when the thin sections are viewed down into the ground and the fibers are traced toward the core object. The fibers at these sites, therefore, are curved in three dimensions, indicating triclinic strain symmetry, and the sense of shear has thrust and left-lateral components. The strain magnitude was determined for seventeen sites from measurements of fiber length. Site means for $1 + e_1$ range from 1.2 to 1.65, and the mean value for $1 + e_1$ for the seventeen sites is 1.4. Site means for $1 + e_2$ range from 1.0 to 1.3, and the mean value for $1 + e_2$ for the seventeen sites is 1.05. If only the seven sites displaying flattening strain are considered, the mean value for $1 + e_2$ is 1.1. For additional information on domain 2, see Underwood (2006).

DISCUSSION

The orientation of structural elements and the strain are relatively homogeneous within each structural domain. In domain 1, the approximately down-dip L_2 is internally consistent with the evidence for plane strain and monoclinic strain symmetry. S_2 is interpreted as having formed within a thrust-sense shear zone. In domain 2, the moderately raking L_2 and the evidence for flattening strain and triclinic strain symmetry are consistent with the predictions of models of transpression in which the shear zone is inclined rather than vertical (Sanderson and Marchini 1984, Jones et al. 2004). S_2 is interpreted as having formed within a shear zone characterized by left-lateral and thrust components of motion.

A straightforward interpretation of these observations and inferences is that the bulk shortening direction was similarly oriented for both domains 1 and 2 during S_2 formation and the orientation of the shear zone differed. The two-dimensional character of the deformation in domain 1 indicates that the strike of the shear zone is north-northeast, i.e., parallel to the strike of S_2. In domain 2, the strike of the shear zone is more difficult to determine. For a bulk shortening direction parallel to the trend of L_2 in domain 1, a shear zone striking northwest will behave as an oblique ramp and have components of left-lateral and thrust motion. A northwest-striking oblique ramp is consistent with the geologic history of the region. Rankin (1976) and

Thomas (1977, 2006) have shown how the overall architecture of the Appalachians was influenced by the Neoproterozoic rift system that developed during the initial stages of opening of the Iapetus Ocean. Northwest-striking transfer zones are an important element of the rift system.

Paleogeographic reconstructions place the Neoproterozoic rift system outboard of the Mesoproterozoic basement massifs in the Appalachians. Although the Taconic allochthon lies to the west of the Mesoproterozoic Green Mountain and Berkshire massifs, the strata composing the allochthon were deposited on the slope and rise of the Iapetus Ocean to the east of the massifs and transported across the massifs during the Taconic orogeny (Stanley and Ratcliffe 1985, Bradley 1989). Basement heterogeneities can cause stress concentrations that affect the deformation of the overlying strata (Schedl and Wiltschko 1987, Wissing et al. 2003). Thus, reactivation of an Iapetan transfer zone as an oblique ramp during emplacement of the Taconic allochthon is a reasonable working hypothesis to explain the characteristics of domain 2.

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ROAD LOG

Miles from last point	Total miles	
0.0	0.0	Meet in the parking lot of the Shaw's supermarket in Fair Haven, Vermont. The supermarket is on the west side of VT 22A 0.25 miles south of US 4. If traveling to Fair Haven along US 4, take Exit 2 and go south along VT 22A to find the supermarket. Be aware that a Shaw's supermarket search using Google Maps and MapQuest may result in directions to a location north of US 4. Leave parking lot and proceed south on VT 22A.
0.1	0.1	Turn left onto Fourth St.
0.6	0.7	Turn left onto Dutton Ave. Dutton Ave. may be labeled Dutch Ave. on Google Maps and MapQuest maps.
0.3	1.0	Cross over US 4.
0.2	1.2	Continue straight on Scotch Hill Rd.
3.9	5.1	At Y in road, stay to the right.
0.2	5.3	Turn left onto Cedar Mountain Rd. at entrance to Bomoseen State Park. Cedar Mountain Rd. is a one-lane, dead-end, gravel road.
1.3	6.6	STOP 1 Park at entrance to abandoned quarry. Do not block driveway to house. Walk north along the path into the quarry. A path leads off to the right from the main path into the quarry. Continue along the main path and climb up slate pile on the east side to view quarry walls.

This is a well-known exposure of the Cedar Mountain syncline. The strata belong to the Cambrian Mettawee (Middle Granville) Slate Formation. The main cleavage in the strata is the slaty cleavage (S_2), which is axial planar to the fold. The west wall of the quarry exposes both a profile view of the fold and a folded S_0 plane with a well-expressed S_0-S_2 intersection lineation parallel to the fold axis. A crenulation cleavage (S_3) is also present and is readily observed in blocks in the pile of waste slate.

The quarry lies in the transition region between Taconic rocks to the west, in which S_2 is the dominant cleavage and S_3 is either absent or only weakly developed, and Taconic rocks to the east, in which S_3 is the dominant cleavage. Because S_3 pervades the strata in the quarry, the exposure is not mapped as part of domain 1. Structural orientations, however, are consistent with this domain. Both the north-northeast trend of the fold axis and the shallow, approximately 25E dip of S_2 are characteristic of domain 1.

In addition to the fold and cleavages, reduction spots are a highlight of the exposure. The green spots can be ellipsoidal or irregular in shape. Reduction bands subparallel to S_0 are also present.

		Turn around and retrace route along Cedar Mountain Rd.
1.2	7.8	Turn right at stop sign.
0.1	7.9	At Y in road, stay to the left.
0.1	8.0	STOP 2 Park temporarily in grassy area along side of road. Obtain permission from homeowner to park in the parking area of one of the houses. The outcrop is the large cliff face just to the east of Glen Lake.

This exposure of the Scotch Hill syncline is a frequent stop on field trips to the northern Taconic allochthon. Please do not hammer on the outcrop face and be on guard for poison ivy.

The folded strata belong to the Ordovician Poultney Formation. The exposure is mapped as part of domain 1 on the basis of the orientations of the fold axis and S_2. Detailed information on the S_2-related strain is not available because strain fringes have not been sampled. The north-northeast trend of the fold axis is shown by the S_0-S_2 intersection lineation, which is best observed on west-facing, vertical S_0 surfaces on the short limb of the fold. The mean orientation of S_2 is 025, 25E.

Two sets of veins can be observed on the outcrop face. Veins in one set intersect S_0 at an angle to the S_0-S_2 intersection lineation, and veins in the other set intersect S_0 parallel to the S_0-S_2 intersection lineation. Cross-cutting relations are visible on both limbs of the fold and show that veins in the latter set are younger. These veins, which are coaxial with the fold, are interpreted to have formed in response to flexural folding. On the short limb of the fold, they display Z shapes. The vein asymmetry is not as clear on the long limb of the fold, but there is a very wide vein that displays an S shape. To observe this vein, go around the small shed and climb to the base of the exposure behind the shed. The top half of the vein is visible in the main outcrop face, and the bottom half is visible in the continuation of the exposure in front of the main outcrop face. Despite being pre-folding, veins in the older set lie in nearly the same orientation on the long and short limbs of the fold. This is interpreted to be a result of rigid rotation of S_0 combined with shearing within layering in opposite senses on the two limbs during flexural folding.

		Continue south along West Castleton Rd./Scotch Hill Rd.
3.9	11.9	STOP 3
		Park along side of road.
		There is an outcrop on the west side of the road. More illustrative outcrops are present to the east along either side of the Exit 3 off ramp from US 4.

These exposures of the Cambrian Browns Pond Formation lie on the long limb of the Scotch Hill syncline. Data obtained from strain fringes have been used in combination with the orientation of outcrop-scale structural elements to place the exposures in domain 1.

Features observable in the exposures include calcite veins, which form two main sets. The veins that strike approximately north-south are more easily observed in the outcrops along the exit ramp, and the veins that strike approximately east-west are more easily observed in the outcrop along the road. Veins in both sets are deformed and are inferred to predate S_2 formation. The thickness and orientation of the veins in the north-south striking set differ depending on whether the vein lies in limestone or slate. In relatively coherent beds of limestone, the veins are thick and about perpendicular to S_0. In slate, the veins are thin and lie at a low angle to S_2 as a result of rotation during S_2 formation. Boudinage is visible in many of the veins in the slate layers. In contrast, veins in the east-west striking set are folded.

		Continue south along Scotch Hill Rd./Dutton Ave./Main St.
1.0	12.9	LUNCH STOP
		Park in parking space adjacent to the town green of Fair Haven, Vermont.
		Continue south along Main St.
0.3	13.2	Continue straight onto VT 22A over railroad tracks. Do not turn right onto VT 4A.
1.6	14.8	Cross into New York State.
3.1	17.9	Turn right onto CR 18.
2.1	20.0	Turn left onto Hills Pond Rd. Hills Pond Rd. is a one-lane, gravel road.
1.0	21.0	Hills Pond Rd. becomes paved.
1.5	22.5	Turn right onto South Rd.
0.1	22.6	STOP 4
		Park along side of road.
		The outcrop is on the north side of the road directly opposite a driveway.

This is a small exposure of the Cambrian Mettawee (Middle Granville) Slate Formation within domain 2. The outcrop lies on the long limb of a north-northwest-trending anticline. The Cambrian Browns Pond

Formation is exposed to the north of the road and yields strain fringes that are curved in S_2-parallel thin sections. This is one of three locations in domain 2 where evidence for triclinic strain symmetry has been found. The strain fringes also record flattening strain.

This stop is in the heart of domain 2, and the overall north-northwest-trending structural grain of domain 2 can be easily seen on Google Earth images of the area directly north of this stop. Although the images have not been analyzed in detail, the prominent lineaments are probably the Browns Pond Formation and sandstone layers in the Hatch Hill Formation.

		Continue west along South Rd.
0.3	22.9	Turn left onto CR 21.
2.3	25.2	Turn left onto NY 22A.
0.1	25.3	Cross over Mettawee River.
0.6	25.9	Veer right onto Raceville Rd.
0.1	26.0	Turn left onto Butler Rd. Butler Rd. is not marked.
0.1	26.1	Turn right onto Cross Rd. Cross Rd. is a gravel road.
0.9	27.0	Turn right onto New Boston Rd.
0.6	27.6	Continue straight onto Stoddard Rd.
1.0	28.6	STOP 5
		Park along side of road.
		One outcrop is on the east side of the road opposite the drained beaver pond, and the other is on the island in the drained beaver pond. The latter outcrop can be accessed using an ATV track at the south end of the pond.

Exposures along this road lie in the core of a syncline in domain 2 and provide excellent examples of deformed graptolites. The short limb of the fold is exposed in the outcrop along the road, and the core is exposed on the island in the drained beaver pond. The graptolite-bearing strata belong to the Ordovician Mount Merino Formation. The graptolites appear as white "films" on S_0 surfaces because during S_2 formation strain fringes developed adjacent to the periderm, which also fractured and infilled with fibrous material. At high magnification, chocolate-tablet structure can be seen in the periderm of graptolites from the exposure along the road. This contradicts evidence for prolate strain from measurements of the thecal spacing of the graptolites (Goldstein et al. 1998). The thecal spacing, which indicates large volume loss, is inferred to record the total strain undergone by the strata since deposition. Estimates of volume change using markers that record only the strain related to S_2 formation support approximately constant-volume deformation.

		Continue south along Stoddard Rd.
0.5	29.1	Turn right at T intersection onto Fox Rd./Depot St. Fox Rd. is not marked.
0.8	29.9	Turn right onto Main St. Main St. is CR 24.
0.1	30.0	Turn right onto NY 22A.
1.2	31.2	STOP 6
		Park in small parking area on the east side of NY 22A.
		The outcrop is directly opposite the parking area.

This road cut exposes the Ordovician Poultney Formation. The outcrop lies on the long limb of a north-northwest-trending anticline in domain 2, and the strain fringes indicate triclinic strain symmetry and flattening strain.

The outcrop provides good examples of clastic dikes and sills (?), which formed before lithification, and brittle faults, which formed after S_2 formation. The clastic dikes lie at a high angle to S_0 and S_2 and are folded. Some of the sandstone layers may be clastic sills. The distance between two relatively thick sandstone layers, for example, changes, indicating that they are not parallel to S_0. In addition, some sandstone layers do not show evidence for internal layering. The brittle faults are generally steeply dipping, and the direction and sense of slip along the faults are well shown by slickenfibers and steps. Normal and strike-slip motions are observed. Post-S_2 faults with similar kinematics have been mapped in other exposures in domain 2.

THE MIDDLE ORDOVICIAN SECTION AT CROWN POINT PENINSULA, NEW YORK

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INTRODUCTION

The Chazy, Black River and Trenton Groups are a well studied sequence of fossiliferous limestones, dolostones and sandstones in the Champlain Valley of New York, Vermont and southern Quebec. These rocks record shallow water cyclic sedimentation in the foreland basin of Laurentia prior to and during the initial stages of the Taconic Orogeny. This field trip reviews the stratigraphy of these units, as well as the evidence for reconstructing depositional environments and cyclic sea level changes.

THE CROWN POINT HISTORICAL SITE

This field trip takes place at the scenic Crown Point Historical Site (Figure 1), a *no-hammer, no-collecting locality*. Approximately 120 meters of section is exposed in a combination of excavations, ridges and shoreline exposures that dip ~8° northwest. The overall stratigraphy (Figure 2) from the Crown Point Formation (Chazy Group) to the Glens Falls Formation (Trenton Group) records the Middle Ordovician transgression seen throughout the Appalachians. Correlation of this stratigraphy exposed at Crown Point to the type Trenton in central New York State is provided by *Prasopora spp.* (Mehrtens and Barnett, 1979). Conodonts collected from strata to the north and east (Harris, pers. comm. and Harris, *et al*., 1995; Roscoe, 1973) provide additional age control. Detailed sedimentologic studies of these units have been published by Bechtel and Mehrtens (1995), Mehrtens and Cuffey (2003) and MacLean (1986) and Selleck (1988).

TECTONIC SETTING

The Middle-Upper Ordovician stratigraphic sequence of westernmost Vermont and northeasternmost New York consists of several rock units (Chazy, Black River and Trenton Groups) which record sedimentation in a progressively deepening foreland basin located to the west of the accretionary prism-volcanic arc terrain of the Taconic Orogeny. To the west of this foreland basin lay the Grenville Province metamorphic basement and overlying Cambrian sandstones of the Adirondack Mountains and Champlain Valley. To the east lay uplifting Taconian lands along the deforming Iapetus margin. To the west of this foreland basin lay the Grenville Province metamorphic basement and overlying Cambrian sandstones of the Adirondack Mountains and Champlain Valley. To the east lay uplifting Taconian lands along the deforming Iapetus margin. As described by Rowley and Kidd (1981) and Stanley and Ratcliff (1985) the eastern margin of Laurentia was an eastward dipping continental margin beneath the Iapetus Ocean and associated island arc terrain. The transgressive sequence recorded by the Middle-Upper Ordovician stratigraphy in Vermont, New York and southern Quebec is interpreted to be the result of a combination of factors related to the Taconic arc-continent collision, including foundering produced by passage of the peripheral bulge through the foreland basin (Jacobi, 1981), thrust loading, or a combination of the two processes (Bradley and Kusky, 1986). The global Ordovician eustatic sea level rise (Mussman and Read, 1986) was also a factor in producing the transgressive sequence along the Iapetus margin.

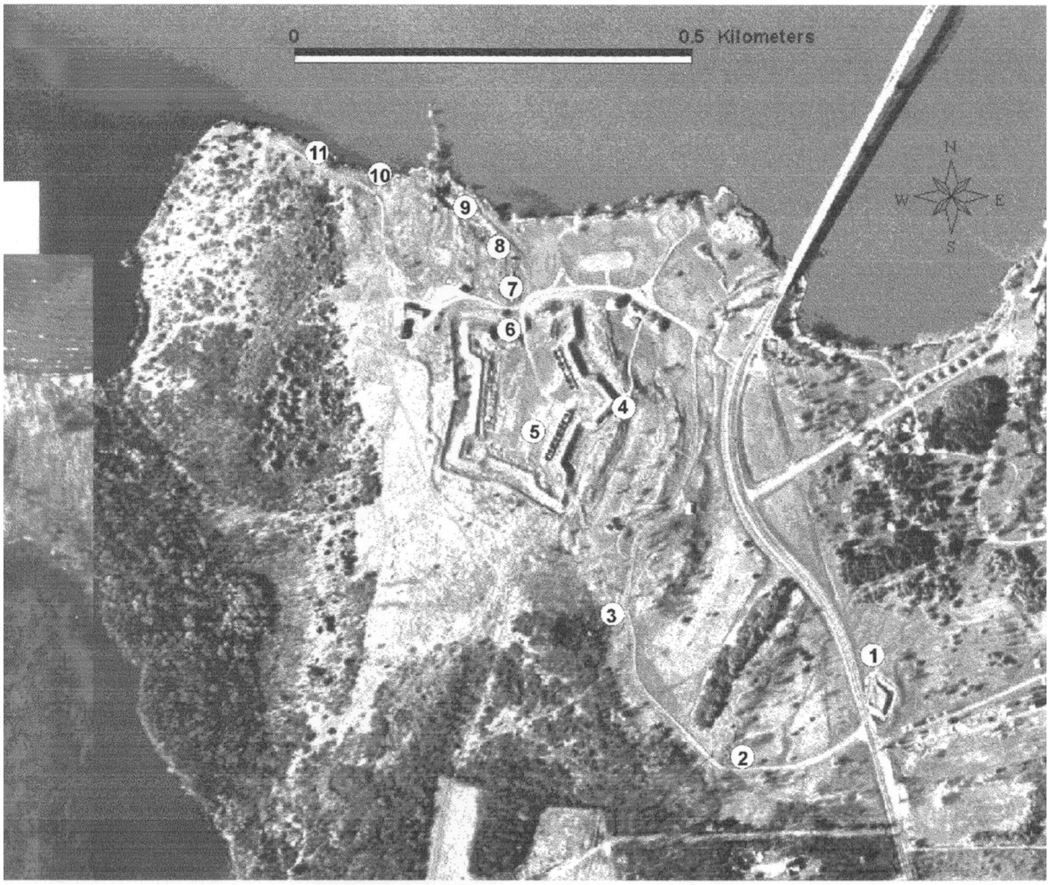

Figure 1. Locality map for the Crown Point Reservations. Numbers refer to field trip stops.

Important to our understanding of the Middle-Upper Ordovician stratigraphy in the Champlain Valley region is the role of syn-depositional block faulting in controlling facies and thickness patterns in the sedimentary sequence. Evidence for active fault movement during deposition of the Middle Ordovician units, a phenomenon described for the Middle Ordovician of central New York by Cisne *et al.* (1982), and southern Quebec (Mehrtens, 1988), includes karstified limestone horizons within the Black River Group (Bechtel and Mehrtens, 1995), clast composition evolution within fault breccias such as the Lacolle Formation (Mehrtens and Gleason, 1988), and condensed stratigraphic sequences, for example, Lower (?)-Middle Ordovician Providence Island Dolostone overlain by the Black River Group, with no intervening Chazy (Bechtel and Mehrtens, 1995). There is also biostratigraphic evidence for an unconformity between the Lower Ordovician part of the Beekmantown Group (Providence Island Dolostone) and basal Chazy (Speyer and Selleck, 1986). Today, the Champlain Valley contains excellent exposures of the Cambro-Ordovician stratigraphic sequence (Welby, 1982) that include autochthonous Middle Ordovician limestones, dolostones sandstones and shales which have been overridden by a series of major thrust faults. The Champlain Thrust, for example, has emplaced siliciclastic and carbonate rocks approximately 80 km westward onto authochthonous black shales (Stanley, 1987).

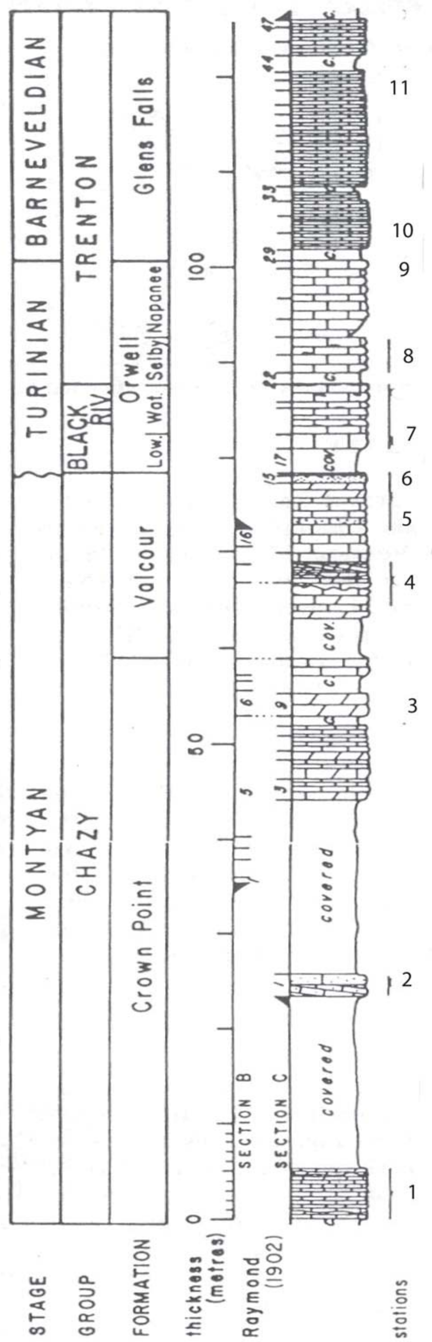

Figure 2. Stratigraphic column for the units visited during this field trip. Numbers refer to field trip stops. Modified from Baldwin and Harding (1993).

CHAZY GROUP

The stratigraphy of the Chazy Group in eastern New York and adjacent Vermont was described by Oxley and Kay (1959) and is summarized in Fisher (1968). Welby (1962) includes a summary of stratigraphic relationships between exposures in New York and Vermont and Hoffman (1963) presents the stratigraphy for the Middle Ordovician units in southern Quebec. Speyer and Selleck (1986), present regional correlations within the Chazy Group in the Champlain Valley.

The Chazy Group is Middle-Upper Ordovician (Whiterockian Series, Chazyan Stage) in age (Figure 2). Based on conodonts (Harris, *et al*., 1995) the Chazy spans the upper half of the *Phragmodus polonicus* through *Ca. sweeti* zones (North Atlantic lower latitude province) or the base of the *Pygodus serra* through *Pygodus ans* Zones (North American cosmopolitan province). In terms of graptolites, the unit extends from the base of the *Diplograptus decoratus* to mid-*Nemagraptus. gracilis* zones.

The Chazy Group varies in thickness within the Champlain Valley from 250 meters in the north (Isle la Motte region) to 90 meters at Crown Point, to less than 15 meters at Ticonderoga and to zero at Whitehall (Oxley and Kay, 1959). In the type area in the northern Champlain Valley, the Chazy can be divided into three formations (Figure 2) as well as several members. Organic buildups (called variously biostromes, mounds, and reefs) occur in all three units, however the composition of the framebuilders varies stratigraphically; Pitcher (1964) summarizes the stratigraphic trends in faunal composition. Organic buildups are absent here at Crown Point, although they can be found across the lake and to the north, in Vermont. The nature of the basal contact of the Chazy Group with the underlying Lower Ordovician Beekmantown Group also varies within the Champlain Valley. In the south, basal Chazy horizons are in unconformable contact with tilted, eroded Beekmantown Group strata. In the northern Champlain Valley, basal Chazy horizons (the Scott Member of the Day Point Fm.) are in apparent conformable contact with the Providence Island dolostone of the Beekmantown Group (Speyer, 1982). The nature of the upper contact of the Chazy Group with the overlying Black River Group is lithologically abrupt, but evidence for significant erosion is generally absent (Bechtel and Mehrtens, 1995). This contact is covered at the Crown Point Historical Site.

The basal Chazy unit, the Day Point Formation, is not exposed at Crown Point and will not be seen on this field trip. Mehrtens and Cuffy (2003) described the depositional environments and reef succession in this unit. They documented the growth of bryo-mounds that typically attained one meter in height and which grew seaward of cross-bedded sand shoals. The mounds frequently exhibit internal zonation; early colonizing *Champlainopora chazyensis* often grows on brachiopod shells and are in turn encrusted by *Batostoma chazyensis*. Cycles of sea level are documented by recurring sand shoals as well as variation in upward and outward growth of the bryo-mounds.

The Crown Point and Valcour Formations will be examined on this field trip; in general both units consist of fossiliferous bioclastic wackestones, packstones and grainstones, with varying degrees of fabric-selective dolomitization. Shaley, nodular limestones are present in the stratigraphy, but are rarely exposed at the surface. Environments of deposition vary from subtidal, storm-dominated shelf settings to more nearshore sand shoals and tidal flats. Intervals of penecontemporaneous cementation and karstic erosion may mark intervals of subaerial exposure.

Cements within the Chazy Group at Crown Point typically consist of an early equant to prismatic low-Mg calcite followed by later coarse calcite spar. Dissolution and chert replacement of aragonite bioclasts is common. Dolomitization in the Crown Point limestones is widespread, and is highly fabric selective in some facies. Variations in primary mineralogy (low-Mg calcite vs. aragonite) appears to have controlled the dolomitization of some bioclastic materials; grain size, sorting, porosity, intensity of borrowing and distribution of early cements (and thus permeability of materials during burial diagenesis) seem to best explain the highly variable patterns of dolomitization (Selleck, 1988).

BLACK RIVER GROUP

The Black River Group in the Champlain Valley is a relatively thin unit (25-30 meters) that consists of massively-bedded wackestones to packstones which represent deposition in lagoonal to shallow subtidal environments. The gradual deepening that characterizes this unit (and which continues into the overlying Trenton Group) is punctuated by cyclic sea level changes that occur on the macroscopic (meter) as well as microscopic (centimeter) scales, the latter visible only in thin section. The lithologic variation within this unit over its outcrop area of New York, Ontario, Quebec and Vermont has contributed to the proliferation of stratigraphic names, however the Pamelia, Lowville (House Creek and Sawyer Bay Members) and Chaumont Formations can be recognized in the Champlain Valley. Bechtel (1993) summarizes the evolution of nomenclature applied to this unit. Based on conodonts (Harris, *et al*., 1995), the Black River Group is Middle-Upper Ordovician (Mohawkian Series, Blackriverian Stage).

The lower contact of the Black River with the underlying Valcour Formation of the Chazy Group is covered at Crown Point, however the basal beds of coarse-grained subarkose sandstones are exposed and will be visited on this field trip. The upper contact of the Black River with the overlying Trenton is likewise covered.

<u>Cement Stratigraphy of the Black River Group</u>

There are multiple types of cements present within the Black River limestones which record a complex diagenetic history. The general cement stratigraphy pattern records early nonluminescent cement associated with precipitation in oxidizing waters of the shallow meteoric phreatic zone. With increasing reducing conditions, bright and dull luminescent cements represent precipitation under shallow burial conditions. Ferroan calcite with dull to nonluminescence represents precipitation in a late burial situation from high temperature burial fluids. Early marine Black River Group micritic cement is ferroan and very dull luminescence, representing deposition in a reducing, lagoonal environment. Subsequent cementation took place in the shallow meteoric phreatic zone, with nonluminscent cements with bright rims representing oxidizing conditions becoming slightly more reducing with burial. These observations are consistent with those of Mussman, *et al*. (1988) who interpreted such patterns to be related to a cratonward-dipping meteoric water lens beneath tidal flats. Tectonic uplift would lead to stagnation of the aquifer and increasingly reducing conditions. Within this general pattern, however, there are many variations in the Black River limestones which record frequent base level changes associated with sea level fluctuations and block fault movements in the Taconic foreland basin. These base level changes have produced numerous firmgrounds (at all Black River localities in the Champlain Valley) as well as beachrock (at Arnold Bay, VT) and paleo-karst (at Arnold Bay, Chippen Point and Sawyer Bay, VT localities).

Fractures are common throughout the Black River and their cements also record evolving burial conditions. The cement stratigraphy of the fractures indicates that their formation occurred throughout the diagenetic history of the Black River Group, from early syndepositional events associated with karst and beachrock formation, through to deep burial under reducing conditions. Figures 3 and 4 illustrate some of the observed patterns. In the first thin section (Figure 3, following page) two cement events are visible in the fracture. The first consists of nonferroan scalenohedral crystals extending outward from the fracture wall. These are interpreted to have been precipitated in the meteoric phreatic zone. The later, large ferroan equant blocky crystals in the center of the fracture represent a late burial cement precipitated under reducing conditions.

In Figure 4 (following page), a photograph taken under cathodoluminescence, the zoning of rhombohedral crystals infilling a fracture can be seen. The very symmetrical zoned patterns starts (from the interior outward) with a nonluminescent nonferroan core, a dull rim, a bright orange rim, another dull rim, to another bright rim, and fading to nonluminescent outer rims. The nonferroan to ferroan zonation is indicative of increasing reducing conditions during cementation. The cement stratigraphy of the fractures indicates that their formation occurred throughout the diagenetic history of the Black River Group, from early syndepositional events associated with karst and beachrock formation, through to deep burial.

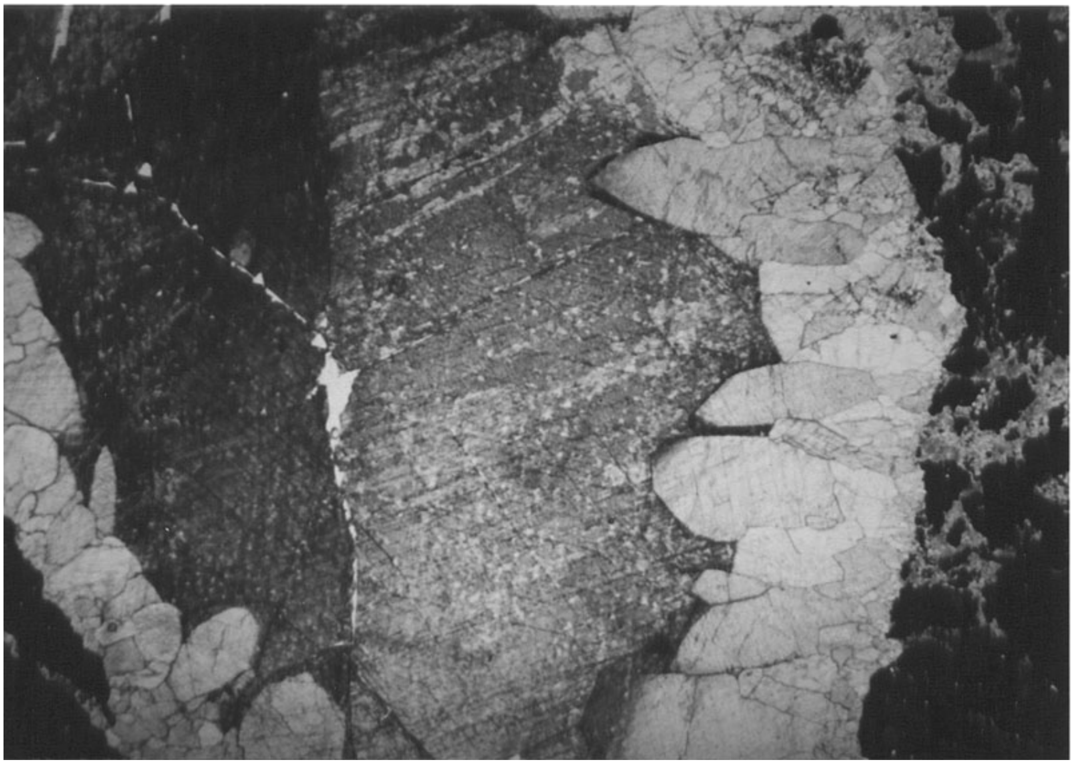

Figure 3. Two generations of fracture cement; field of view is 1.8cm

Figure 4. Cathodoluminescent photograph of fracture cements (field of view 0.5mm)

TRENTON GROUP

The contact between the Glens Falls Limestone (Trenton Group) and the Black River Group is covered at most localities in the Champlain Valley, including here at the Crown Point Historical Site. At Arnold Bay, to the northeast of the Crown Point site, the contact is exposed and is interpreted to be a disconformity; the dark gray colored, massively bedded Black River is in sharp contact with the nodular-bedded, laterally discontinuous beds of the Trenton. MacLean (1987) measured the Trenton Group around the Champlain Valley and found that here at Crown Point only 9 meters are exposed (but more may be covered by lakeshore muds) whereas a few miles to the north, at Button Bay it is more than 15 meters thick. The Button Bay exposure is significant because it is the only place where both the lower upper and contacts, with the Black River and Cumberland Head Argillite, respectively, are seen. Bechtel (1993) summarizes the variable nature of the Black River/Trenton contact around the Champlain Valley, New York and Ontario, and he notes that the regional variation in the nature of this contact, as well as the thickness variation along strike, would be expected in a foreland basin actively undergoing syndepositional block faulting.

MacLean (1986) identified seven lithofacies in the Glens Falls Formation, recognized by variations in lithology, bedding style, sedimentary structures and biota. The shallowest bathymetry is represented by a lithofacies consisting of grainstones composed of peloids and oncoids exhibiting pinch and swell bedding, graded bedding and cross laminations, which suggest wave reworking within fair-weather wave base (Figure 5, following page; arrow points to hummocky cross laminae in the grainstone). At the other bathymetric extreme are bioturbated (*Teichinchnus* and *Chondrites*) mudstones interbedded with shale and distal tempestites/turbidites composed of wackstone/packstone to mudstone couplets (Figure 6, following page). MacLean also recognized a bryozoan-rich wackstone/packstone lithofacies (his Facies F) that is characterized by abundant *Stictopora* and *Eridotrypa* that form dense thickets or lens-shaped patches on the muddy sea floor. In addition to these ramose bryozoa, this lithofaces contains trilobite, gastropod and brachiopod remains; algae are notable absent. These bryozoan thickets are interpreted to have accumulated on the shelf near fair-weather wave base. Although the absence of algae suggests bathymetry below the photic zone, the abundance of interbedded shale suggests that turbidity may have reduced sunlight infiltration.

Using occurrences of *Cryptolithus* and *Prasopora* for correlation, MacLean documented that the Glens Falls stratigraphy in the northern Champlain Valley differed from that in the south; the transition from shallow water (inner ramp) to deep water (outer ramp) is abrupt in the Champlain Islands whereas in the south (including Crown Point) bathymetric change is hard to discern and more cyclic.

The Glens Falls stratigraphy visible at Crown Point is very representative of the bulk of this unit in the southern Champlain Valley, dominated by grainstone to wackestone/mudstone couplets in laterally discontinuous beds (MacLean's Facies C). Grainstone beds are commonly laminated and exhibit fine-tail grading (Figure 7). Thin shale seams separate beds. *Chondrites* and *Helminthropsis* are the dominant ichnofauna. Rare bedding planes exhibit a bedding plane fauna of bryozoa (*Prasopora*, *Stictopora*) and trilobite fragments. Moving up section at Crown Point the hummocky bedding style of the base of the section is replaced by more laterally continuous grainstone beds with sharp, planar bases that grade upwards into bioturbated wackestones/mudstones (MacLean's Facies E). These horizons are interpreted to be bioclastic turbidites.

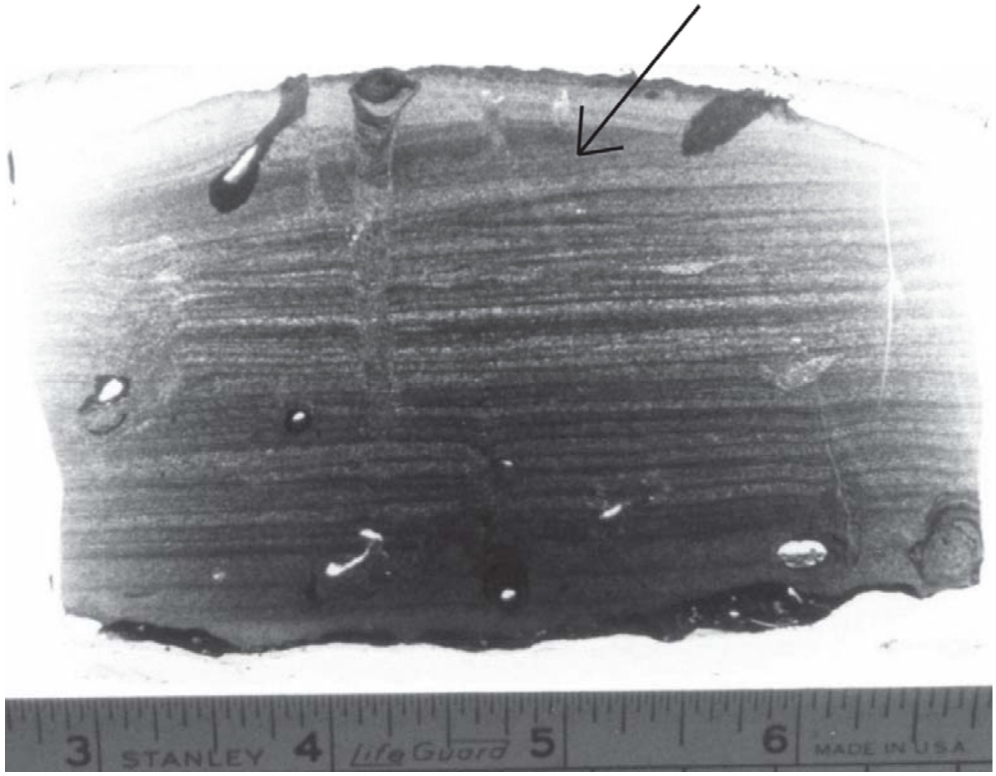

Figure 5. Large thin section (tape for scale) of cross lamination in a Glens Falls Fm. grainstone.

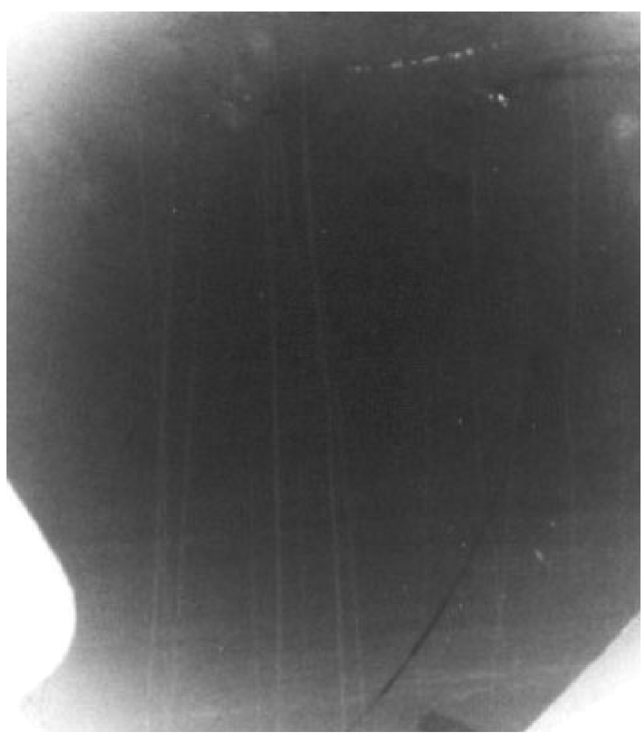
Figure 6. Large thin section (12 cm long) of the outer ramp lithofacies of the Glens Falls Fm.

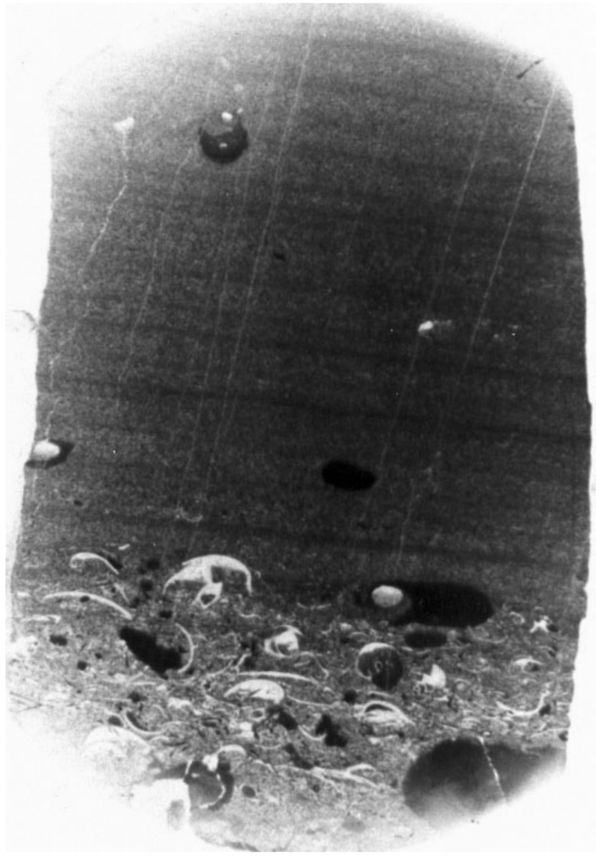

Figure 7. Large thin section (18 cm in length) of graded bed in the Glens Falls Fm.

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FIELD TRIP LOG

All the stops for this field trip are within the Crown Point Reservation Historical Site. From the west, take NY Route 22 north from Ticonderoga, continuing north through the village of Crown Point. Turn east approximately five miles north of the village, following signs to the “Bridge to Vermont.” The Crown Point site is immediately before the bridge crossing. From the east, take VT Route 22A north from Fairhaven, or south from Burlington, and follow signs to “Bridge to New York.” The Crown Point site is immediately over the bridge crossing. Stop locations are keyed to Figures 1 and 2. The Crown Point Reservation Historical Site is a no hammer, no collecting outcrop.

<u>Stop 1 – the Redoubt</u>

Approximately 6 meters of burrowed, slightly dolomitic, thin to medium bedded bioclastic packstone and grainstone is exposed at this stop. Some beds are relatively well-sorted grainstones with sharp bases which are interpreted as tempestites. The same interpretation is made for the grainstone intraclast-rich bed visible in the low ledge at the southeast corner of the ditch. Approximately 3 meters above the base of the section abundant *Girvanella* algal oncolites are common allochems. Nuclei of fossil fragments are visible within some oncolites. In other beds fossils are relative abundant, and are best seen on slightly weathered bedding surfaces. Trilobite fragments, brachiopods, bryozoans, pelmatozoan fragments, nautiloids and large *Maclurites magnus* are present (exterior shells and opercula). Dolomite occurs in along shaley seams and in burrow fill.

The relatively high faunal diversity, general bioturbation, and storm-related sedimentation all point to a low energy shallow subtidal environment at depths slightly below fair-weather wave base. The abundant algal oncolites and discrete calcareous algal fossils (e.g., *Hedstromia spp*.) suggest depths well within the photic zone. A possible modern analogue is found in the mixed mud and sand shelf to the west of the emergent tidal flats of Andros Island, Bahamas, as described by Purdy (1963).

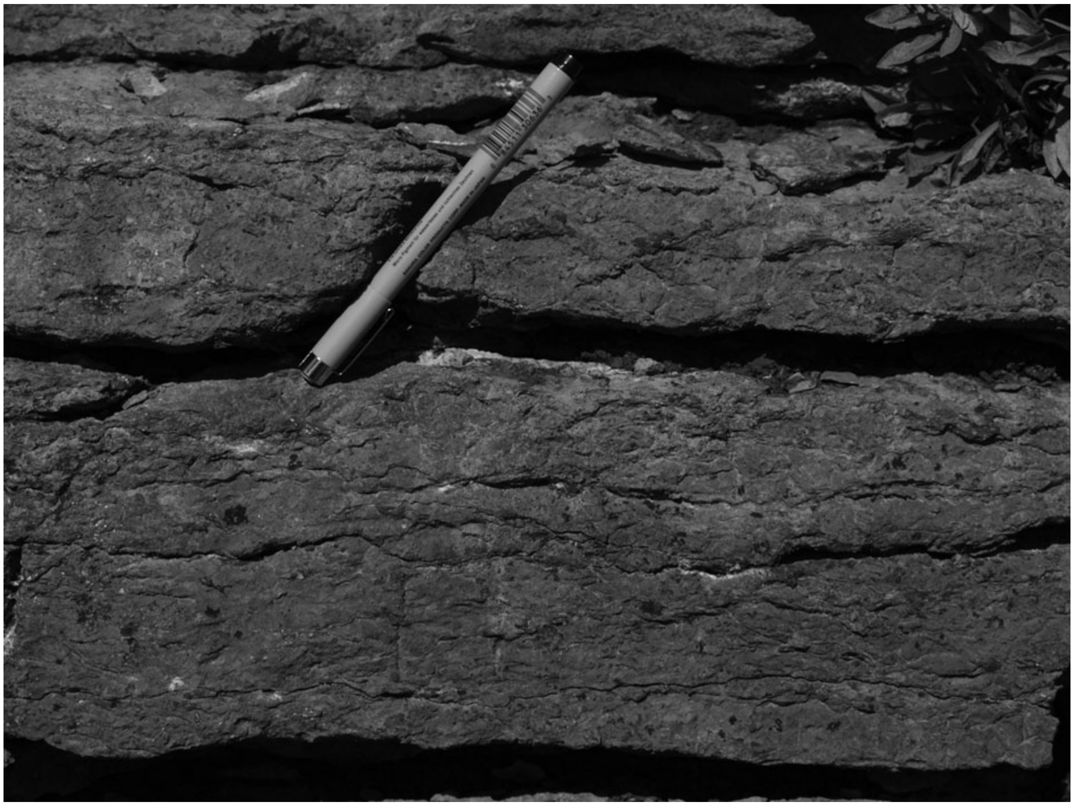

Figure 8. Nodular limestone in shaley dolostone matrix, Crown Point Fm. Pen is 13 cm long for scale

We interpret the wavy, irregular dolomite laminae as the result of dolomitization of lime mud, followed by compaction and pressure solution of calcite that produced irregular clay and dolomite-rich stylocumulate seams. Preferential dolomitization of burrows is due to contrasts in permeability of burrow-fill versus burrow-matrix sediment. The burrow-fill material retained permeability longer during diagenesis and allowed more pervasive dolomitization. In similar facies exposed on Bullwagga Bay (the west shore of the Crown Point Peninsula), modular limestone with shaley dolostone seams and stringers are present. The limestone nodules appear to have been cemented prior to significant burial compaction, whereas the shaley dolostone material was compacted around the cemented limestone. The early-cement limestones were resistant to dolomitization. This sort of fabric selective dolomitization is common in the Chazy and Black River Groups throughout the Champlain Valley.

Stop 2 – Ridge of outcrop running NE from entrance gate
WARNING: poison ivy is common along this ledge

Cross-stratified coarse bioclastic grainstones are well-exposed near the main gate along the entrance road and adjacent ridge. Nearly three meters of section form a prominent belt parallel to strike, extending from the entrance road to the main highway. Foreset cross-strata show bipolar dip directions. Angular quartz and felspar grains are concentrated in some laminae. The carbonate particles are dominantly sub-rounded abraded pelmatozoan plates with gastropod and brachiopod fragments. Large *Maclurites* fragments and grainstone intraclasts are present on the upper bedding surfaces of these ledges.

We propose that the environment of deposition of this facies was a shallow subtidal wave and current reworked bioclastic sand shoal. Active transport of abraded grains may have been accomplished by tidal currents (as suggested by the bi-directional cross-strata) or by storm-generated currents that produced complex, anastomosing patterns of cross-strata and intervening reactivation surfaces. The lack of burrowing and well-preserved whole-shell body fossils may be due to the inhospitable shifting sand substrate. A modern analogue for this environment would be the unstable sand shoals of the Bahamas Platform (Ball, 1967), as the scale and style of cross stratification are similar. Oxley and Kay (1959) report that similar strata in the northern Lake Champlain Valley are oolitic.

Stop 3 – low ledges adjacent to the entrance road (Picnic Pavilion ridge), approximately 50 meters north of Stop 2

Brown-weathering, slightly shaley dolostone exposed here contains lenses and stringers of fossiliferous lime packstone and wackestone. The fauna is similar to that at Stop 1, with trilobites, small brachiopods and *Maclurites* common. The environment of deposition is assumed to be similar to that of Stop 1, however lacking evidence of storm activity. Note that some of the fossils are almost entirely encased in dolomite, which is assumed to be of replacement origin here.

Stop 4 – SE moat of Fort Crown Point

Approximately 3 meters of thickly laminated limestone and dolostone of the Valcour Formation are exposed in the southeast moat of the British Fort. The dominant facies here is alternating 0.5-2.0 cm thick laminae of limestone and dolostone, we term "ribbon rock." The limestone ribbons are very fine-grained peloid grainstones or "calcisiltites" and appear blue-grey color on slightly weathered surfaces, and as indentations on more deeply weathered surfaces. The dolostone ribbons weather tan-brown in color, and consist of an interlocking mosaic of 20-300 micron dolomite crystals of replacement origin. Quartz silt grains are present in the dolostone ribbons, versus medium to fine-grained quartz sand in the limestone ribbons, suggesting the limestone horizons were slightly coarser-grained than the dolostone when deposited.

An erosional surface with 10 to 20 cm of relief is exposed near the base of the south wall of the moat (Figure 9). Similar erosional surfaces occur within this facies in other Chazy exposures in the Champlain Valley. We interpret these to represent micro-karstic solution surfaces on a tidal rock platform that developed during subaerial exposure of cemented limestone. Typically, the rock below the surface is mostly calcite limestone, suggesting that cementation and diagenetic stabilization of the limestone occurred prior to development of the erosional surfaces. Overlying rock horizons contain more dolomite. *Maclurites* shell hash can be found in pockets on the erosional surface. Dolomitized burrows cut across the limestone ribbons in some parts of the outcrop. Trough cross-strata filling low scour surfaces are also visible.

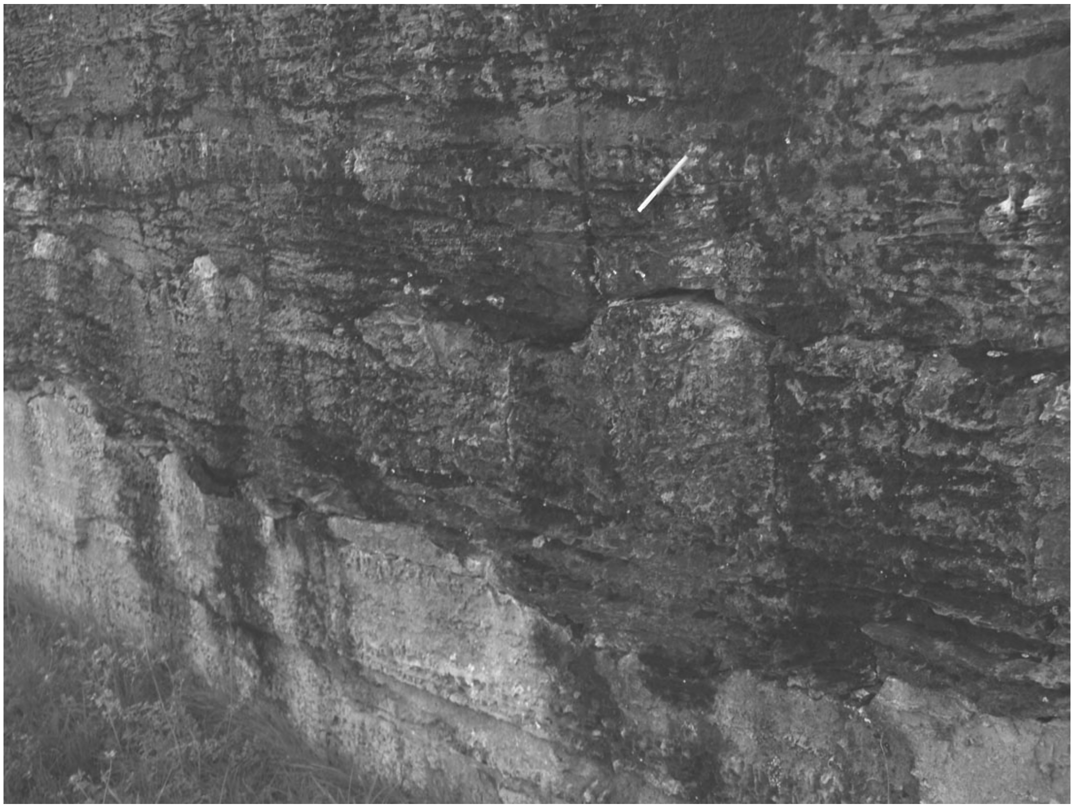

Figure 9. South wall of moat, with erosional surface near the base overlain by differentially weathering “ribbon rock.” Pen is 13 cm long for scale.

On the less weathered prominence on the SE corner of the moat, shallow scours containing brachiopod and gastropod debris can be seen. Intraclasts or pseudoclasts of limestone in dolostone are also present. Some “clasts” appear to be cored by dolomitized burrows.

We interpret the stratigraphy seen at this stop as representing a tidal flat to shallow subtidal facies. The alternating limestone/dolostone ribbon rock represents rhythms of slightly coarser-grained (limestone) and finer-grained (dolostone) sediment deposited on the lower reaches of a tidal flat, similar to the rhythmic bedding described by Reineck and Singh (1980) from the mud/sand tidal flats of the North Sea. These coarse-fine alternations might also reflect storm-related, ebb-surge deposition (Nelson, 1980). Early cementation of the slightly coarser-grained limestone ribbons made this lithology less susceptible to dolomitization, which affected the finer-grained muddy ribbons that became dolomitized. Variations in the intensity of burrowing is interpreted to reflect subtle differences in duration of subaerial exposure of the tidal flat and/or the extent of reworking by tidal currents. Limited *in situ* faunal diversity is to be expected in the tidal flat setting, where organisms are stressed by salinity fluctuations. The absence of mudcracks or evidence of evaporite minerals may indicate that only the lower portion of a humid climate tidal flat system is preserved here.

Stop 5- Parade Grounds near barracks:

As we enter the parade ground from the southwest corner of the moat, note the array of carbonate rocks used in construction of the barracks walls. Chazy, Black River and (rarely) Trenton lithologies can be identified. Restoration of the barracks began in 1916 and in 1976 the New York State Division for Historic Preservation undertook protection and stabilization of the ruins.

The low rock pavement just north of the barracks is within the upper part of the “ribbon rock” unit seen at Stop 4. Immediately up-section, cross-stratified grainstone beds are exposed. Coarse-grained quartz and feldspar sand is easily seen on weathered surfaces. Trough cross-strata and herringbone co-sets of planar-tabular cross-strata are visible on the vertical surfaces. Large, angular clasts of slightly dolomitic grainstone and *Maclurites* are present on bedding surfaces. We interpret this facies as representing a current-dominated sand shoal environment, similar to that seen at Stop 2.

Stop 6- West Parade Grounds

Bechtel and Mehrtens (1993) suggested that the sandstone unit in the westernmost parade ground is the basal sandstone of the Black River Group, an interpretation which differs from that of Speyer and Selleck (1988), who suggested that this unit was part of the underlying Chazy Group. In thin section this sandstone is a quartz arenite, very poorly sorted, and containing fewer lithic fragments and phosphatic fragments than stratigraphically lower Chazy sandstones. Visible at the very easternmost portion of this ridge is a buff-colored dolostone bed containing pockets of quartz sand (burrow infills?). These basal sandstone and dolostone lithologies are very similar to those described by Walker (1972) in his description of the Pamelia Formation at its type locality in north-central New York. Alternatively, placement of the sandstone unit within the Chazy Group is consistent with the common presence of coarse-grained quartz and feldspar sand within the Chazy at the Crown Point Preserve, whereas the siliciclastic material in the Black River Group at Crown Point is mainly silt and clay. Whatever the stratigraphic placement of the sandstones and dolostones here, it marks an interval when sands were transported from a nearby source area to this possible peri-tidal setting. This interval was followed by marine reworking of the sand and deposition of fine-grained limestone of the basal Black River Group.

Stop 7 – Northeastern moat
WARNING: poison ivy is common along the base of this exposure

There is approximately one meter of covered interval between Stop 6 in the parade grounds and Stop 7 in the northeastern moat. Along this wall are exposed several meters of the lower Black River Group (Lowville Formation, House Creek Member) which in Vermont is termed the Orwell Limestone. At the base of this exposure a series of stylolitized gastropod-bearing (*Liospira*) wackestone beds are overlain by thicker beds of *Phytopsis*-burrowed aphanitic mudstones. This sequence can be interpreted as a shallowing-up cycle (SUC) consisting of subtidal overlain by peritidal lagoonal muds. Examine the sharp contact of the aphanitic mudstone with the overlying wackestones, a contact which in thin section appears to be a firmground (Bechtel, 1993). These SUCs also comprise the base of the Black River Group at other localities in the Champlain Valley. There are three motifs of repetitious bedding that occur in the Black River Group and the cycles seen here at the base are the the thickest, occurring at a macroscopic scale, interpreted to represent 4^{th} order (10,000 to 100,000 years) or smaller cycles. Examination of *Phytopsis* burrows in thin section (Figure 10) reveals that many are filled with graded (fining-up) geopedal silt, evidence of cementation in the meteoric vadose zone.

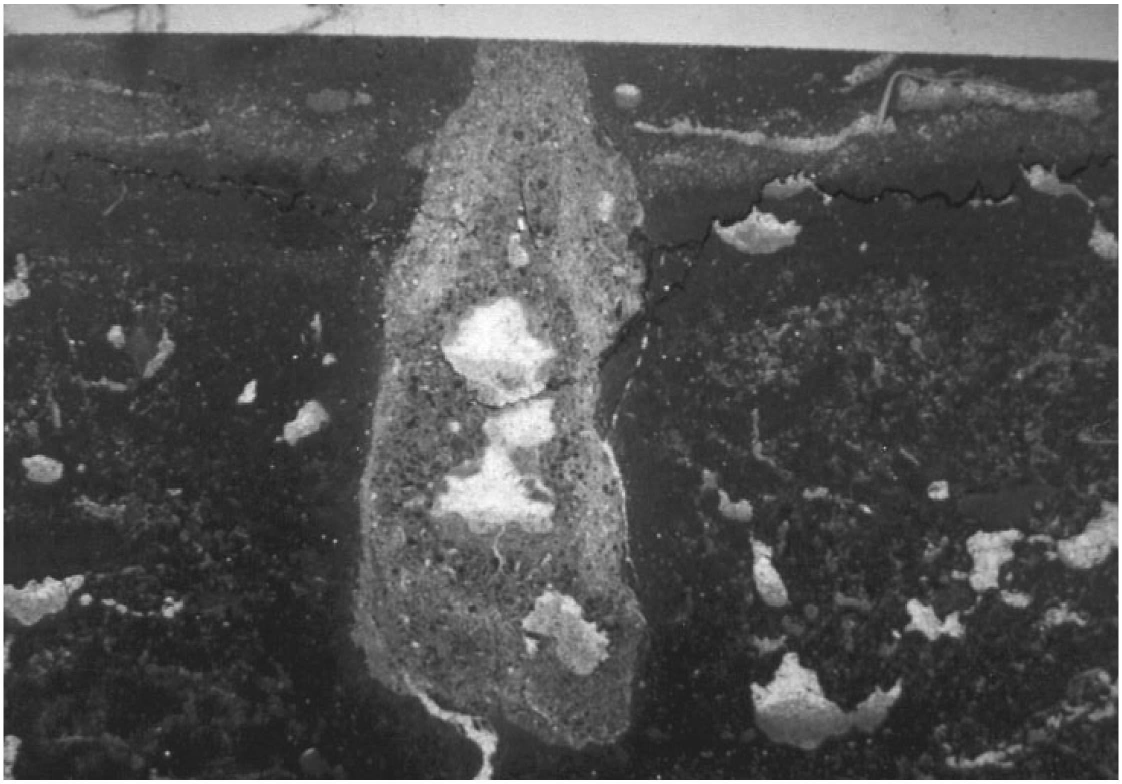

Figure 10. Thin section photomicrograph of burrow fill. Field of view 2.4cm

Continuing up section, several thick packstone beds are exposed. More detailed examination of these beds reveals that they consist of alternating one to six centimeter thick intraclast and oncolite-rich packstone horizons interpreted as tempestites, interbedded with fossiliferous wackestone/packstone horizons. The tempestites consist of graded and crudely imbricated intraclasts and skeletal fragments. Note the nature of the upper and lower contacts of these horizons. The second common bedding motif in the Black River Group are these tempestite horizons interbedded with *in situ* fossiliferous muds.

The uppermost third of the outcrop appears to be a massive bed of limestone, however closer examination also reveals small scale cycles of alternating wackestone/packstone and grainstone, the third motif of bedding in the Black River strata. These cycles are characterized by a base of thinly laminated or cross laminated grainstone horizons 0.5 to 1 cm thick, overlain by fossiliferous wackestones and packstones. In thin section the bases of the grainstones can be identified as firmgrounds, recognizable by the truncations of allochems and cements present in the underlying mud (Figure 11).

Figure 11. Thin section photomicrograph of a firmground; field of view is 1.5cm

The very top of this exposure (best seen at the next stop) exhibits a burrow mottled fabric with selected dolomitization of many burrows. *Tetradium* occurs in life position in these horizons.

Stop 8 – east-west ridge across the service road

A black chert layer near the top of Stop 7 provides the correlation to Stop 8, the outcrop across the service road. The limestone beds on this ridge commonly consist of alternating wackestone/packstone and planar to cross laminated grainstone beds, as seen at Stop 7, however bedding plane exposures permit identification of many fossils in these, the most faunally diverse beds in the Black River Group. Specimens of gastropods (*Liospira*, *Lophospira, Hormotoma*), *Lambeophyllum*, *Tetradium*, stromatoporoids, the bivalve *Cyrtodonta,* the brachiopod *Stromphomena sp*. and cephalopods are recognizable. This ridge exposure is most notable for its bedding plane exposures of *Tetradium* and *Lambeophyllum.* It is interpreted as representing a wave baffle margin lithofacies similar to that described by Walker (1972) at the Black River type section.

The third cycle motif we recognize, laminated grainstone overlain by bioturbated wackestone, so common in this portion of the stratigraphy, are interpreted to represent smaller scale 4th order cycles, which could be the result of facies mosaicing and/or small scale base level changes.

There are at lest two distinct types of chert occurrences in the Black River Group. One, the infilling of horizontal burrows, is fabric selective. Chert also occurs less frequently as broad bedding plane parallel sheets. The uppermost chert horizon on this ridge, traceable down to the shoreline, is of this latter variety. Clearly, there was a significant source of silica available for chert formation, perhaps a combination of silica derived from sponges (*Tetradium*?) and bentonite alteration (the Ordovician sequence in the Champlain Valley is notorious for the paucity of bentonites compared to equivalent strata in central New York and southern Quebec, although Ryan, *et al*., 2007 have recently been successful in identifying altered bentonites in the Beekmantown Group in the Champlain Valley). In thin section the chert cross cuts all previous cements, including late fracture-filling calcite, and it is therefore the youngest form of diagenesis present in these rocks.

Stop 9 – Quarry to the north

WARNING: be extremely careful around the quarry as the thick algal scum in the quarry water obscures where the grass ends and the quarry wall drops off.

The older weathered south walls of the quarry show, by color differentiation, two cycles. There are 3 to 4 cm thick beds of planar laminated skeletal and peloidal hash overlain by burrowed wackestones overlain by intraclast-rich horizons. These cycles represent our third motif of Black River bedding. Interbedded with these cycles are tempestite couplets of mudstone/wackestone and fossil hash layers in which brachiopod-rich layers are abundant. The abundant quarried blocks lying about provide the opportunity to look for cycles, and from these, topping directions.

Stop 10 – shoreline of eastern Bulwaga Bay

The uppermost horizons in the quarry can be traced down to the shoreline to the north where the uppermost Black River strata can be seen (Chaumont Formation). Shoreline bedding planes exhibit horizontal traces of *Chondrites* and opercula of *Maclurites.* There is a thin covered interval of approximately one meter to the basal beds of the Glens Falls Limestone of the Trenton Group. The Glens Falls is characterized by thin beds of nodular to wavy bedded wackestones, mudstones and rare grainstones. Bedding planes along the shoreline contain mostly *Chondrites* and *Helmenthopsis* burrows, however as one moves up section, recognizable fragments of *Cryptolithus, Isotelus,* orthid brachiopods, *Stictopora*, and *Prasopora simulatrix* can be found; the latter is important because it permits the correlation of the lower Glens Falls here in the Champlain Valley to the lower Denley Limestone at the Trenton type section in central New York (Mehrtens and Barnett, 1979).

MacLean (1986) interpreted the lithofacies of the basal Glens Falls visible here to represent sedimentation in a shallow subtidal environment periodically influenced by storm activity. In thin section the nodular and wavy bedded wackestones appear thoroughly bioturbated, a process which would influence and enhance subsequent differential compaction. Grainstone beds exhibit more planar bases with basal skeletal fragment lags or finely crushed debris of brachiopod, trilobite and crinoidal material and capped by carbonate mud. MacLean interpreted these as tempestite deposits. Moving further up section (in horizons not seen at Crown Point) the Glens Falls records progressively deepening conditions from the subtidal, storm influenced conditions seen here to bioclastic turbidites separated by authochthonous shale horizons. For those familiar with the lower Trenton Group localities in central New York State (ex., Rathbun Brook, City Brook, Inghams Mills), the paucity of fossiliferous bedding planes here at Crown Point is noteworthy. The overall fine grain size and ichnofauna suggest that bathymetry increased significantly and rapidly from the Black River into the Glens Falls, a transition that might reflect not only rising sea level but base level changes as well. The sedimentologic and faunal transitions from the Glens Falls to the overlying Cumberland Head Argillite and the Stony Point Shale are much more gradational than that of the Black River/Glens Falls contact.

GEOLOGY AND GEOCHRONOLOGY OF THE EASTERN ADIRONDACKS

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INTRODUCTION

The eastern Adirondacks contain a wide variety of rock types, and this trip aims to visit representatives of the major lithologies. Stops include units recently dated by U-Pb geochronology including multi- and single-grain TIMS and SHRIMP II analyses.

The Adirondack Mountains represent a southwestern extension of the Grenville Province via the Thousand Islands – Frontenac Arch region of the St. Lawrence River (Fig. 1). The region is topographically divided into the Adirondack Highlands and Lowlands separated by the Carthage-Colton Mylonite Zone (CCMZ, Fig. 2). The former is underlain largely by orthogneisses metamorphosed to granulite facies and the latter by upper-amphibolite grade metasediments, notably marbles. Both sectors have experienced multiple deformations resulting in refolded major isoclines. The tectonic history consists of a tripartite division, as summarized below and in figure 3. Broadly, the various lithologies fall into the following groups: 1300-1350 Ma tonalites and granodiorites; 1160-1150 Ma anorthosites, mangerites, charnockites, and granites (AMCG suite); 11000-1090 Hawkeye granite; and 1050-1040 Ma Lyon Mt Granite (LMG). The three major orogenic events associated with these are the Elzevirian (ca1350-1220), the Shawinigan (1215-1160 Ma) and the Ottawan Orogenies (1090-1045 Ma). The Elzevirian involves protracted outboard arc magmatism and accretion and was followed by the collisional Shawinigan Orogeny between Laurentia and the Adirondack Highlands – Green Mountain Terrane. The Ottawan was a Himalayan-type collision of Laurentia with Amazonia (?). Most of the metamorphic and structural effects present in the Adirondacks are the result of the Ottawan Orogeny, but Shawinigan and Elzevirian features can be recognized locally. Both the AMCG suite and the LMG are thought to be late- to post-tectonic manifestations of delamination of overthickened orogens.

Structurally, the eastern Adirondacks are dominated by the same large, recumbent fold-nappe structures (F_2) as found in the southern Adirondacks (Figs. 2, 4, 5), and with fold axes oriented dominantly ~E-W parallel to lineation. As in the southern Adirondacks, the fold-nappes are thought to have sheared-out lower limbs, but this has yet to be demonstrated on a map scale. At least two distinguishable upright fold events are superimposed on the nappes: F_3 with shallow plunging ~E-W axes and F_4 with shallow-plunging NNE axes. All of three fold sets affect Hawkeye and older units, and thus must be of Ottawan age. This also the case with the strongly penetrative rock fabric, including strong ribbon lineations, that are present in these rocks and are associated with the large fold-nappes. Intense fabric and nappe structure are largely absent from the Lyon Mt Granite and this is interpreted to reflect its intrusion in late, post-nappe stages of the Ottawan. In the northern portion of the eastern Adirondacks the NNE, F4, folds become quite tight and have a strong lineation associated with them. This may be the result of rock sequences being squeezed between large, domical prongs of anorthosite. Finally, we note that there exists abundant local evidence of small isoclinal F_l folds that pre-date F_2. These, and their associated fabrics, are thought to be pre-Ottawan in origin, and in a few cases this can be shown to be the case.

A dominant feature of the eastern Adirondacks is the great Marcy anorthosite massif that underlies almost all the High Peaks (Fig. 2). Over two-dozen zircon age determinations demonstrate that the anorthosite was emplaced at 1155 $\pm$ 8 and that associated granitoids and coronitic metagabbros are coeval with it (Fig. 6). These relationships make it clear that thermal energy from the anorthosite had nothing to do with the regional granulite facies (ca 1050 Ma) that characterize the Adirondack Highlands and post-dates the AMCG suite by 60-100 million years. It is likely that most of the region experienced peak temperatures on the order of ~750-800^{o}C and pressures of ~8 Kbar. Based upon extensive isotope work by John Valley and his students, it appears most likely that this metamorphism proceeded under fluid-absent conditions (Valley *et al*., 1983). Note, however, that this does not exclude the presence of late, post-peak fluids associated with the emplacement of Lyon Mt Granite.

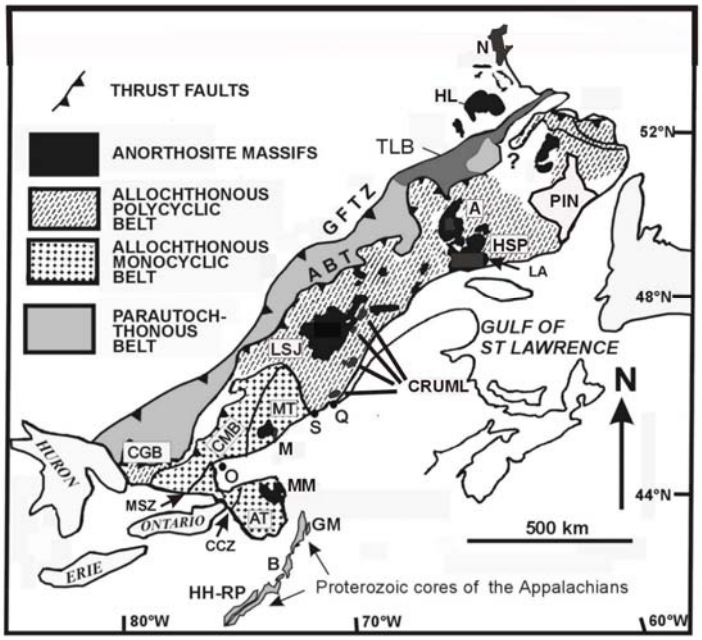

Fig. 1. Generalized location map of the Adirondacks as a southwestern extension of the Canadian Grenville Province whose three major tectonic divisions (Rivers, 1997) are shown. ABT-Allochthonous Boundary Thrust; GFTZ- Grenville Front Tectonic Zone; TLB (dark gray) - Trans-Labrador Batholith, GM – Green Mountains, H – Housatonic Mountains, HH-RP – Hudson Highlands and Reading Prong. The major anorthosite massifs (with ages) of the region are: MM- Marcy (ca 1150 Ma), M) Morin (ca 1160 Ma), LSJ) Lac St. Jean (ca 1150 Ma), HSP) Havre St-Pierre, A) Atikonak (ca 1130 Ma), ME), HL) Harp Lake (ca 1450 Ma), N) Nain-Kiglapait (ca 1300 Ma),; P) Pentecote (ca 1350 Ma).

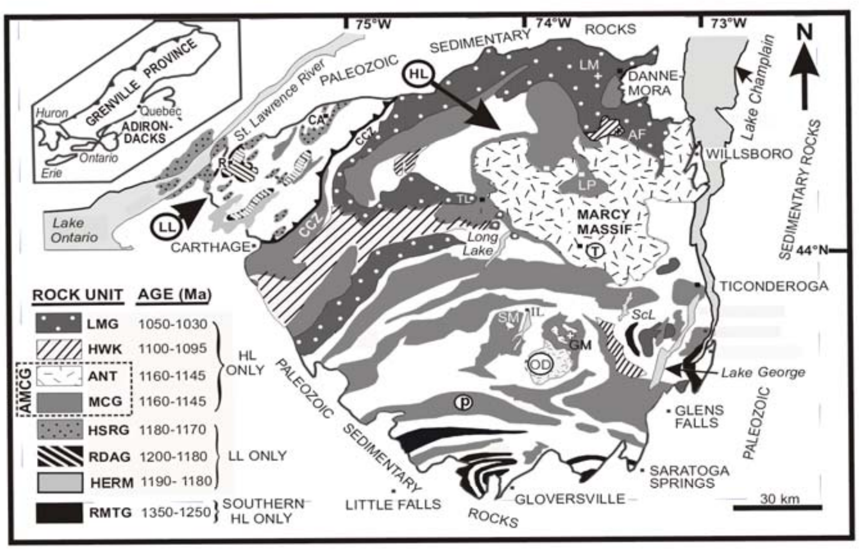

Fig. 2. Generalized geological/geochronological map of the Adirondacks showing locations of dated AMCG samples. Units designated by patterns and initials consist of igneous rocks dated by U-Pb zircon geochronology with ages indicated. Units present only in the Highlands (HL) are: RMTG – Royal Mountain tonalite and granodiorite (southern HL only), HWK -Hawkeye granite, LMG – Lyon Mountain Granite and ANT - anorthosite. Units present in the Lowlands (LL) only are: HSRG – Hyde School and Rockport granites (Hyde School also contains tonalite), RDAG – Rossie diorite and Antwerp granodiorite. Granitoid members of the AMCG suite (MCG) are present in both the Highlands and Lowlands. Unpatterned areas consist of metasediments, glacial cover, or undivided units.

Recent investigations in the Adirondacks have demonstrated that emplacement of Lyon Mt. Granite accompanied extensional collapse of the Ottawan orogen and exhumation of the high-grade granulite core. The most complete study of this sort is that of Selleck et al. (2005) that utilized zircon geochronology to document that down-to-the-west displacement along the northwest dipping Carthage-Colton shear zone (CCZ, Fig. 2) was coeval with intrusion of Lyon Mt. Granite (Fig. 2) into the fault complex at ca 1045 Ma. It is thought that at peak Ottawan contraction the overthickened lithosphere experienced delamination by foundering, convective thermal erosion, or both. Following delamination of the dense lithospheric keel, the orogen rebounded and hot new asthenosphere moved to the crust-mantle interface and caused melting of deep crust to yield Lyon Mt. Granite, As the granite ascended into the crust it both lubricated and enhanced low-angle normal faults formed in response to increased topographic elevations and leading to orogen collapse. In such situations, it is not uncommon to find that collapse is quasi-symmetrical around the orogen core thus giving rise to a mega-gneiss dome or double-sided core complex such as the Shuswap complex. Given this, we have undertaken research aimed at establishing whether, or not, low-angle down-to-the-east normal faulting took place in the Adirondacks at ca 1045 Ma. Geological evidence is consistent with this proposition, and kinematic indicators related to the issue are described in Stop 9 of the Road Log where the authors present evidence supporting topside down-to-the-east displacement. Some of the more regional geological evidence is discussed below.

The Canada Lake isocline is a prominent feature of the southwestern Adirondacks (Fig. 7). The lithologies comprising it reappear to the east of Great Sacandaga Reservoir in the southeastern Adirondacks, but given the pattern of strikes and dips, there are only two ways of achieving the repetition: either by thrust faulting or normal faulting. Although there is remaining uncertainty, kinematic indicators tend to support the normal fault alternative. The fault trace consists of a distinctive belt of pegmatites and quartz veins that are locally mylonitic and stained by hydrothermal fluids. We propose that this belt represents the trace of a low angle (~45º) normal fault whose projection is shown on figure 7. Coarse synkinematic pegmatites are common in, and near, the belt, and secondary muscovite occurs locally. To the east of the fault trace metapelites and ca 1300 Ma tonalites are common, whereas east of it they are not encountered. These issues are discussed in greater detail at stop 9.

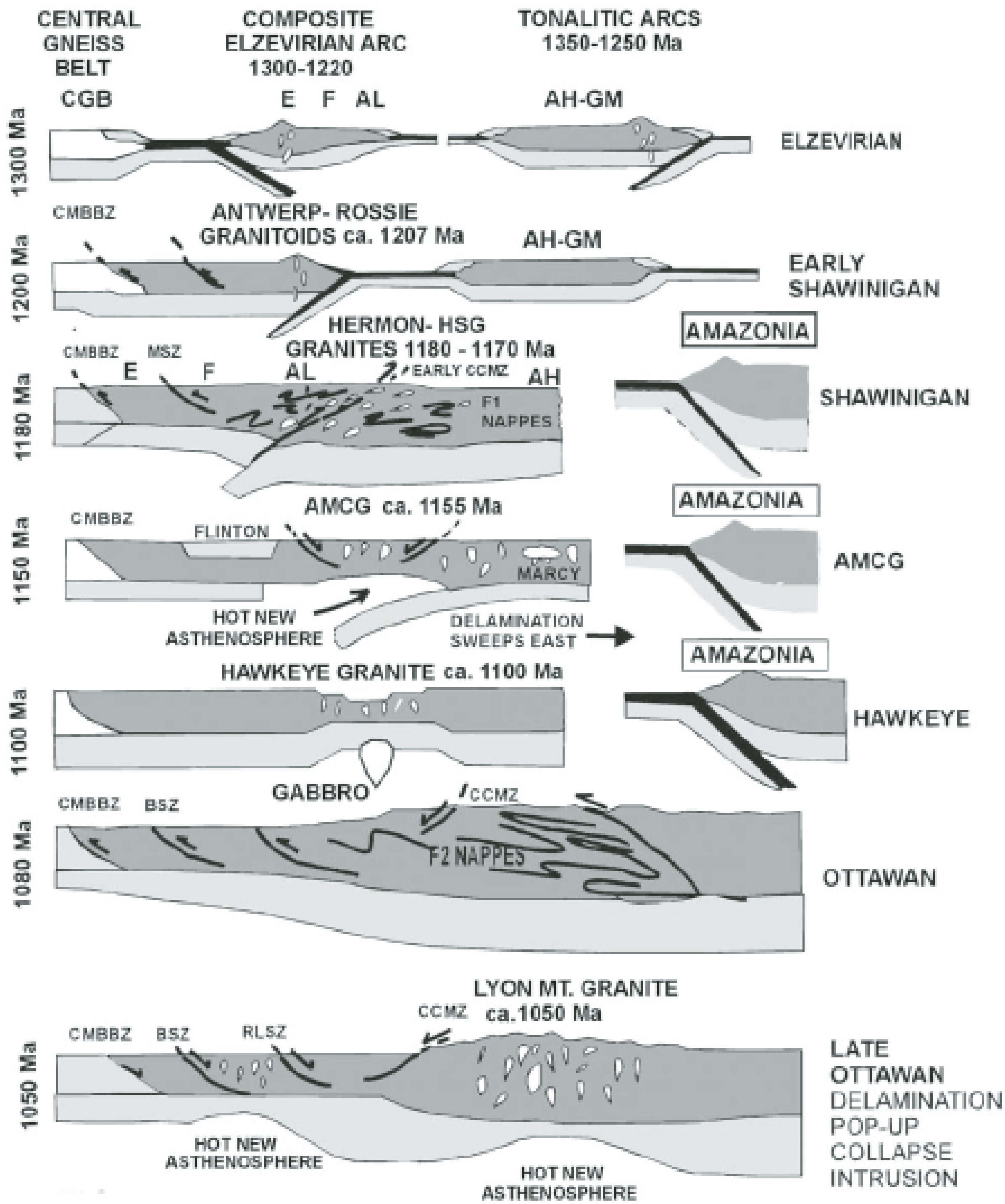

Figure 3. Generalized tectonic evolution of the Adirondack region.

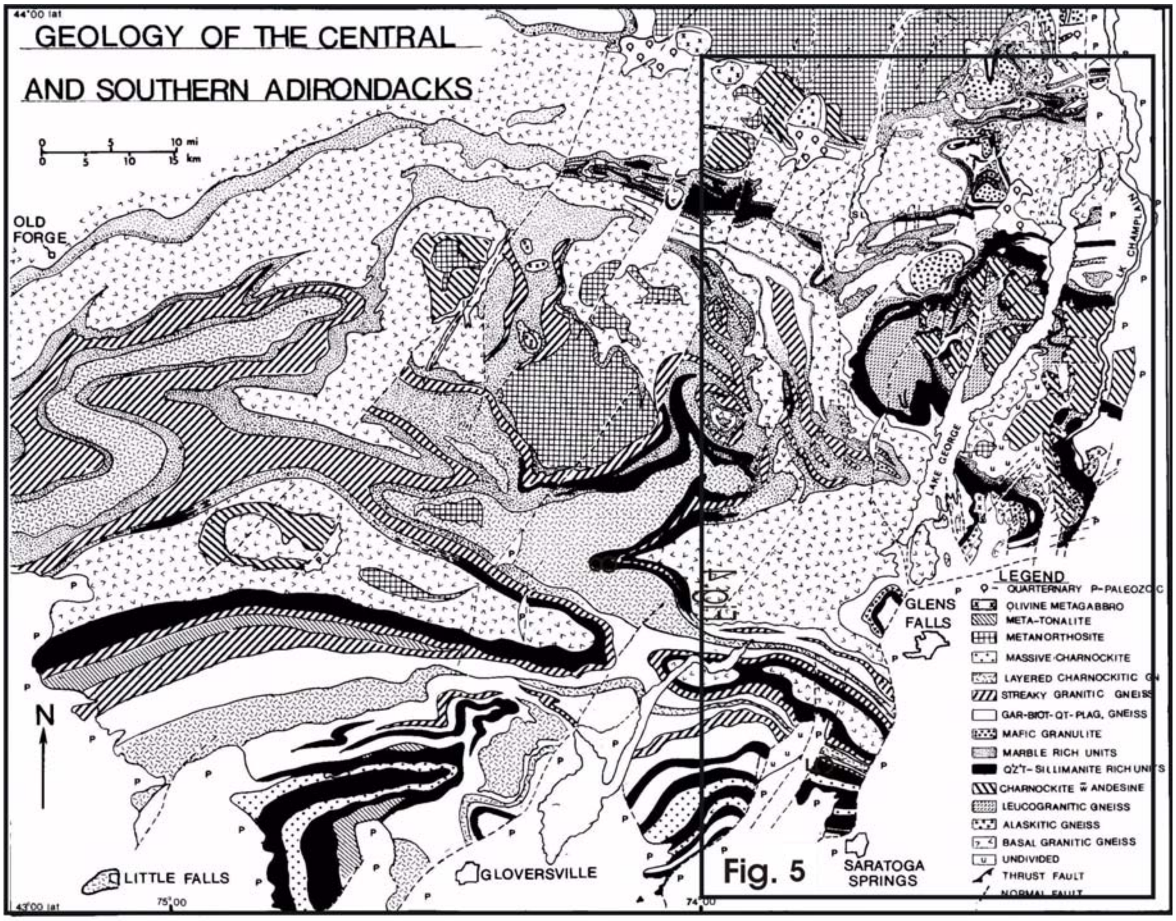

Figure 4. Generalized geology of the southern Adirondacks. Region of Fig. 5 shown.

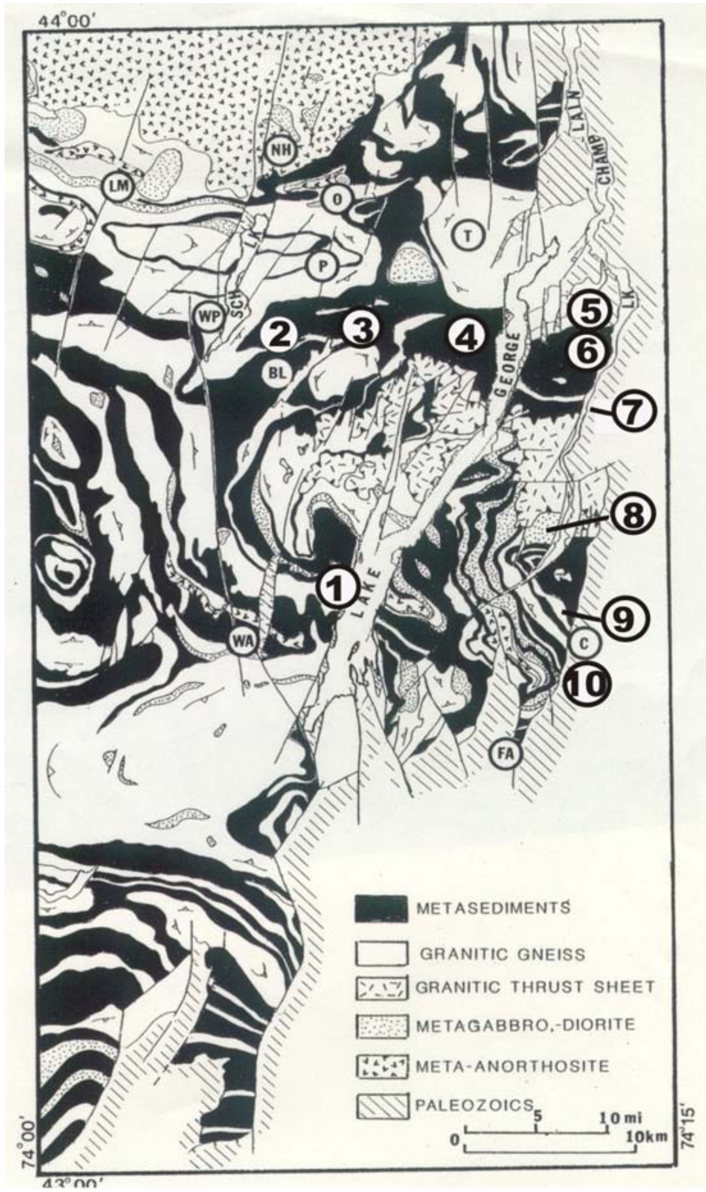

Figure 5. Generalized geologic map of the eastern Adirondacks with stop locations shown. BL – Brandt Lake; C – Comstock; FA – Fort Ann; LGV – Lake George Village; LM – Loch Mueller; NH – North Hudson; O – Owl's Head Mt.; P – Pharoah Mt.; T – Ticonderoga; Wa – Warrensburg; WP – Whitney Point.

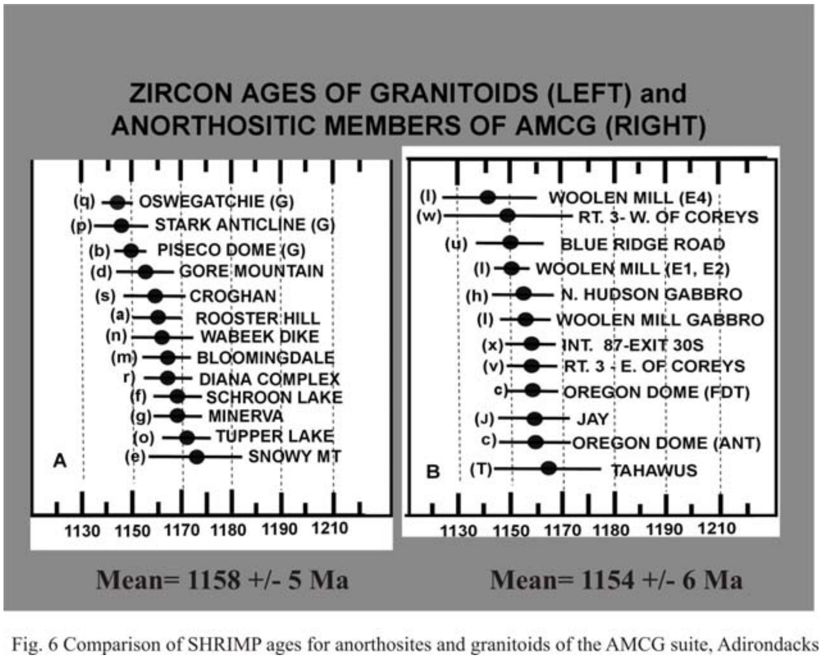

Figure 6. Comparison of SHRIMP ages for anorthosites and granitoids of the AMCG suite, Adirondacks.

ROAD LOG

Mileage

0.0 From parking lot of the Fort William Henry Conference Center turn left (south) on Rt. 9.
0.5 Pass Prospect Mt entrance on right.
1.0 Junction with 9N. Turn right at light.
1.2 Turn right (north) on Northway (Rt 87). Note exposures of pink granite along Northway.
4.0 Sulfidic staining on NNE fault in granite.
5.5 Exposure of hornblende granite on left.
5.9 Pegmatite on right.
6.0 Pass metagabbro
6.7 Hornblende granite on both sides.
7.0 Hornblende granite
7.2 Hornblende granite with metagabbro
7.4 Metagabbro, pegmatite, and metasediments on right.
7.5 Take Exit 23 off of Northway
8.0 Turn right at stop sign
8.4 Turn left (north) onto Schroon River Road.
8.8 Junction with Wall St on right. PARK

STOP 1. The Poor Man's Gore Mt. (30 min)

These exceptionally good roadcuts provide instructive observations regarding the origin of the famous mega-garnets of Gore Mt. and dated at ~1050 Ma. Although these garnets are smaller than the classic examples at Barton Mines, they occur in the same sort of coarse grained, almandine amphibolite and have the same sort of relationships as the former. Moreover, they are similar to 6-7 other occurrences scattered around the Adirondacks. Importantly, each of these, including Barton Mines, is characterized by its location along a fault zone characterized by pegmatites, quartz veins, and various other hydrothermal manifestations. We shall return to this key observation.

The protolith of the garnetiferous amphibolite is demonstrably a coronitic olivine gabbro of the type associated with the coeva1 ca. 1155 Ma massif anorthosites of the Adirondacks (Fig. 6). These anhydrous, silica undersaturated protoliths are commonly well preserved in pluton interiors. The timing of garnet corona formation has been dated at ~1050 Ma, when the Adirondack Highlands were at upper amphibolite to granulite facies conditions (Basu, pers. comm., Connelly, 2005). Note that many garnets appear to be localized in trains that suggest that they formed along fractures.

Rims of coarse, black hornblendes that resemble reaction rims surround the large garnets. However, these are best accounted for as reaction “behinds”, i.e., as the original gabbroic composition reacted to form garnet at a given nucleation site, there was an excess of hornblende-forming components that accumulated into the rims. In gabbroic anorthosites the more feldspathic bulk composition results in lag deposits of white, plagioclase rims.

The northern contact of the amphibolite is defined by a very coarse, white pegmatite consisting of striated oligoclase, K-spar, quartz and hornblende is one many oligoclase-bearing pegmatites of the eastern Adirondacks that have been dated at ca, 1042 Ma and are related to ~1050 Ma Lyon Mt. Granite. A narrow offshoot of the pegmatite crosscuts within the main body of the amphibolite. A highly deformed, ~1155 Ma hornblende granite lies north of the pegmatite and is crosscut by quartz veins. Both the pegmatite and quartz veins manifest a locally abundant hydrothermal phase present at ~1050 Ma. Just prior to this time the olivine metagabbros were at ~ 700C and 10-20 km AND DRY. Fine-grained coronas may have been in the process of forming between olivine and plagioclase. At this point, a high angle fault developed and provided access for hydrothermal, pegmatitic differentiates of Lyon Mt Granite. As these fluid-rich materials encountered the olivine gabbro, hydrous solutions moved along fracture-related pathways and resulted in greatly enhanced mobility of chemical species at ~700C. The result was the almandine amphibolite that we see today. The reason that such grain size is anomalous is that hydrous fluids are rarely present at these temperatures and depths. The model also explains the association of over a dozen other similar deposits, including Gore Mt, along pegmatite-bearing faults.

The final chapter in this story concerns the presence of light colored clots of calcic plagioclase and orthopyroxene in the rock. These are associated with garnets and commonly develop as rims between hornblende and garnet or within pressure shadows associated with garnet. This assemblage is due to the reaction:

Garnet + Hornblende $\rightarrow$ Ca-Plagioclase + Orthopyroxene.

The entry of orthopyroxene marks passage into granulite facies conditions and cannot merely be due to “loss of water” since breaking down the reactant assemblage requires that its stability curve be exceeded to the high temperature side. Note that the plagioclase-orthopyroxene assemblage also occurs in what appear to be pressure shadows on the garnets and have an igneous appearance. Anatexis could have been promoted by dehydration reactions upon entering the granulite facies.

Return to vehicles and turn around.
9.6 Enter Rt I87N
31.3 Exit Northway at Pottersville.
31.6 Stop sign. Turn right (north) at junction with Rt 9.
31.7 Continue north on Rt 9.
32.3 Turn right (east) on Glendale Road by Word of Life Bible Camp.
32.8 Cross Schroon Lake outlet.
34.0 Quartz-sillimanite nodules in Lyon Mt Granite.
34.2 Roadcut of Lyon Mt Granite on left (north) side of Road. Park on right next to Smith Pond.

STOP 2. Ca 1050 Ma Lyon Mountain Leucogranite (30 Minutes).

The medium grained, pink leucogranite is characteristic of the main quartz-microperthite unit of Lyon Mt Granite that occurs in broad tracts along the perimeter of the Adirondack Highlands. The classic low-Ti Kiruna-type magnetite deposits of the Adirondacks are closely associated with the granite, either occurring within it or in rocks (commonly metacarbonates) immediately adjacent to it. Magnetite mineralization is accompanied by sodic alteration manifested by checkerboard albite and by quartz-albite (Ab_{98}) replacements of the original quartz-microperthite rocks (McLelland et al., 2002). A total of 12 occurrences of Lyon Mt Granite from across the Adirondacks have been dated by U/Pb techniques including single and multigrain TIMS and SHRIMP II (Fig. 2). All of these fall into the interval 1050 $\pm$ 10 Ma (McLelland et al., 2001). The present roadcut was dated by single grain TIMS methods and yields an age of 1049 $\pm$ 2 Ma (Fig. 3). Staining of the roadcut reveals that it contains several sheets and layers of quartz-albite rock one of which is associated with a irregular veinlet of magnetite. Locally, quartz-sillimanite nodules are present in these rocks and are attributed to leaching by high temperature hydrothermal fluids (McLelland et al., 2002). Pegmatites and quartz veins are common. Oxygen isotope studies indicate that sodic alteration and magnetite deposition took place at relatively high temperatures of ca 650°C. The granite exhibits some deformation and grain-shape fabric, but it is very much less intense than that present in older rocks (cf. the ca 1155 Ma Hornblende granite at Stop 1).

Evidence from the Carthage-Colton Zone along the Highlands-Lowlands boundary demonstrates that the detachment normal faulting along that zone overlapped emplacement of the Lyon Mt. Granite in the zone. As discussed in Selleck et al (2005) the granite enhanced the faulting by providing lubrication and elevated fluid pressures.

34.9 Turn left (north) on Short Street
35.4 Turn left (north) on Adirondack Road.
37.3 Turn right (northeast) on Pease Hill Road.
38.2 Stay left at intersection
40.9 Turn left (north) at intersection with Palisades Road.
41.6 Quartz-sillimanite nodules in Lyon Mt Granite
43.0 Junction with Rt 8. Turn left (northeast) at stop sign.
43.7 Metapelites assigned by Walton and deWaard to their “Older Paragneiss”.
44.3 Marble, calcsilicate of Walton’s “Paradox Lake Formation in near continuous outcrop for 2.6 miles.
46.9 Long roadcut of migmatitic metapelite assigned by Walton to the “Treadway Mt Formation”. Park.

STOP 3. Deformed Migmatitic Metapelites (> 1300 Ma). (30 Minutes).

The southern and eastern Adirondack Highlands, as well as the Northwest Lowlands, contain significant thicknesses of dark garnet-biotite-quartz-oligoclase + sillimanite metapelite accompanied by white garnet-bearing quartz-feldspar leucosome. McLelland and Husain (1986) interpreted the leucosomes as *in situ* anatectic products of the original rock and the dark fraction as restite. The presence of armored spinel, and rarely corundum, in the restite is consistent with this interpretation. In regions of low strain the leucosomes occur as crosscutting, anastamosing, irregular sheets, dikes, and veins of clearly coarse-grained pegmatitic material. As strain increases, the pegmatitic material gets pulled into psuedoconformity, disrupted, and grain size reduced so as to yield porcellaneous layers that are commonly parallel but locally retain their original crosscutting configurations. In short, the apparent layering in these rocks is almost wholly tectonic and has nothing to do with original stratigraphic superposition. They are, in fact, mylonitized migmatites, of the variety referred to as “straight gneiss”. These metapelites are crosscut by the ca 1300 Ma tonalitic suite, and therefore pre-date or are penecontemporaneous with these rocks. Our interpretation is that their current reintegrated compositions are best accounted for by a greywacke-shale precursor, and that these were arc-related, flysch-type sediments approximately coeval with the Elzevirian tonalites. This would also help to explain the relative proximity between these lithologies.

The age of anatexis within the metapelites has been ascertained by SHRIMP II studies on zircons from leucosomes and restites in both the Highlands and Lowlands (Heumann et al, 2004, Bickford et al, 2008) and reveals cores of age ca 1340-1240 Ma and narrow metamorphic overgrowths of 1010-1040 Ma. Sandwiched between these are mantles with oscillatory zoning and ages of 1170-1180 Ma. We interpret these mantles as zircon grown during anatexis thus dating this event as Shawinigan. Within the southwestern Highlands, the absence of further significant anatexis during the granulite facies Ottawan Orogeny (1090-1030 Ma) is thought to be a consequence of earlier dehydration. Within the Lowlands, the absence of any Ottawan zircon is interpreted as a reflection that this terrane did not experience granulite facies conditions at ca 1090-1030 Ma. Significantly, east of the present locality, zircons exhibit both Shawinigan and Ottawan (ca 1050-1040 Ma) overgrowths, and the latter can be quite robust. We interpret this observation to reflect the introduction of fluids into the eastern Adirondacks late in the Ottawan Orogeny. Within the outcrop at this stop, the thick vein of coarse pegmatite shows excellent Ottawan overgrowths exhibiting excellent oscillatory zoning.

A second, highly deformed leucosome was also dated and yielded the same results. An important difference is that oxygen isotope values for zircons from the large, coarse vein suggest that it was not formed in situ whereas those in the highly deformed vein are consistent with an in situ origin, as is the case in all other metapelitic migmatites studied.

Both mafic and granitic sheets can be traced across the roadcut face and reveal the presence of fault offsets. Note that the name “Treadway Mt Formation” assigned to this unit by Walton suggests that it has stratigraphic characteristics and continuity. This assumption is no longer considered to be correct and is, at best, is a lithotectonic assignation.

47.2 Granite and gabbro.
47.9 Large isoclinal fold in quartzites and metapelitic rocks.
48.1 Undeformed gabbro.
48.4 Marble. Park in parking area.

STOP 4. Swede Mt Sequence (60 Minutes, Including Lunch).

The metasediments exposed along the Rt 8 at Swede Mt provide an exceptional opportunity to closely examine both the rocks and structure. At the southwestern end of the sequence (mile 47.9) a well-exposed isoclinal fold can be seen in the large roadcut on the south side of Rt 8. Folded units include marble, quartzite, sillimanite-garnet-quartz-feldspar (khondalite) gneiss, and rusty sulfidic Dixon Schist. The latter is thought to represent a sheared, altered, and graphitic variety of the khondalite. The fold axis trends ~E-W at a low angle of plunge about the horizontal, and its axial plane, which dips from ~50 degrees SW to horizontal, has been folded about upright E-W axial planes. The fold is typical of refolded isoclines in the Adirondacks. Similar rocks at Dresden Station, on the east side of Lake George, are intruded by a metagabbro dated at 1106 $\pm$ 7 Ma. At this locality the metagabbro truncates foliation and even individual garnet grains in the khondalite. These relations indicate that the khondalite was deposited and first metamorphosed in Elzevirian times. Early workers considered the khondalite to be a stratigraphic unit (Hague Gneiss or Spring Hill Pond Formation), but regional mapping suggests that individual units are continuous over distances of miles only. Both Dixon schist and the khondalite were mined for flake graphite during the early part of the century. The region around Swede Mt is known as the Dixon National Forest and the former mining hamlet of Graphite is located at mile 49.8.As Rt 8 is followed eastward, an undeformed dolerite dike is encountered. The dike strikes parallel to the highway and has caused brecciation and alteration of the country rock that consists principally of marble at this locality. At the top of the hill, and across from Swede Pond, a long roadcut exposes typical Adirondack marbles. At the eastern end of these exposures, a thick bed of quartzite is wrapped around the nose of an isoclinal fold. Note the linear features along its axis. Xenoliths of the quartzite occur in the metagabbro across the road.

49.7 Elephant rock. Consists of leucocratic sillimanite-garnet-quartz-feldspar gneiss (khondalite).
49.8 Khondalite, Dixon Schist and graphite. Trail on right side of road leads to old graphite mines.
54.2 Stop sign. Turn left (north) at junction with Rt 9N.
61.2 Charnockite of the Ticonderoga Dome.
63.7 Rotary. Bear left (north) on Rt 9N.

64.4 Stop light. Junction with Rt 74. Turn right (east).
67.1 Ticonderoga Railroad Station and Fort View Inn.

72.1 Park on right side of road

STOP 5. Classic Great Unconformity (10 Min)

This roadcut provides a classic exposure of flat-lying Potsdam sandstone (~550 Ma) in profound angular unconformity with steeply dipping Proterozoic gneisses (ca. 1300-1050 Ma). A conglomeratic layer is developed at the base of the Potsdam, and zone of alteration is present along the unconformity.

72.7 Turn left into large parking area

STOP 6. Isoclinal Folds And Intrusion Breccia (30 Min)

Roadcuts on the west side of the highway expose excellent examples of Adirondack isoclinal folds developed in granitic gneiss of uncertain, but possibly ~1150 Ma (AMCG) age. The fold axes are oriented almost E-W and plunge gently around the horizontal, which is common in the Adirondacks. However, the axial planes are vertical which is very unusual for Adirondack isoclinal folds, which are commonly recumbent. The upright axial planes appear to be the result of a late, upright anticlinal fold that has rotated recumbent isoclines to their present upright positions. In places the folded units appear to be crosscut by other granitic material. Towards the north end of the roadcut a deformed pegmatite fills a small reverse fault.

South from the folded outcrop a large roadcut exposes a series of intrusion breccias in which granites have disrupted amphibolites and other rocks. A number of crosscutting relationships can be seen. Many of the granitic rocks display minimal deformational fabric and are thought be Lyon Mt. Granite (see Table 1 for whole rock geochemical data). In any case, the intrusion breccia is one of the very few recognized outside of the anorthosite massifs. Its proximity to the upright isoclinal folds is considered to be significant but is not yet clearly understood.

Table 1. Whole rock XRF analyses of the Lyon Mt. Granite at Stop 6

	White LMG	Pink LMG
SiO_2	74.49	75.29
TiO_2	0.06	0.07
Al_2O_3	13.42	13.25
Fe_2O_3	0.70	1.11
MgO	0.01	0.03
CaO	0.76	0.45
Na_2O	2.22	3.28
K_2O	7.45	5.77
P_2O_5	0.10	0.079
Sum	99.30	99.44

78.1	Parking area with potable spring water
79.1	Park along right side of road

STOP 7. Dresden Station Roadcut (30 Min)

Large roadcuts on either side of the highway contain some of the best exposures seen anywhere in the Adirondacks. Lithologies include strongly foliated ferrodiorite at the south end of the western roadcut and an olivine metagabbro at the south end of the eastern roadcut. Also present are marbles, calcsilicates, garnet-sillimanite metasediments, and graphitic schist. Dips are everywhere steep to the south, and a strong subhorizontal E-W lineation is widely developed. Well-exposed isoclinal folds are present at several localities in the roadcuts. These features are described below.

a) The ferrodiorite belongs to a much larger mass of this rock that can be seen a few hundred feet south in the next roadcut. It is typical of late liquid fractionates of anorthosite, which it grades into to the west and uphill. The contact with quartzite at the south end of the roadcut is highly foliated and may represent a detachment surface. Elsewhere this structural horizon is associated with intense mylonitization.

b) A few tens of feet to the north an isoclinal fold is clearly visible and is cored by marbles.

c) The south end of the eastern roadcut passes through a dark metagabbro with a U-Pb zircon age of 1106 $\pm$ 7 Ma indicative of AMCG suite affinities (McLelland and Chiarenzelli, 1989, Aleinikoff, 2006 pers. comm.). The rock contains olivine and two pyroxenes with clinopyroxene exceeding orthopyroxene. The olivines are rimmed by orthopyroxene-clinopyroxene-garnet coronas and small grains of spinel clouded plagioclase. McLelland and Whitney (1980) have extensively described coronas of this type, and relevant reactions are given in their publications. The coronas are thought to have formed during granulite facies conditions characteristic of the Ottawan Orogeny. The northern contact of the metagabbro exhibits a good chill margin and adjacent garnet-sillimanite metapelite has undergone microscopically visible anatexis along the contact. Within the metagabbro good ophitic texture is well preserved and randomly oriented plagioclase laths are undeformed despite the fact that these rocks experienced the Ottawan Orogeny (1090-1030 Ma). Preservation of original, delicate igneous textures in the metagabbros is thought to be the result of their strength, low ductility, and anhydrous nature. The pristine nature of the chilled contact margin may reflect the fact that the local country rock had already been dehydrated. This possibility is manifested by relationships plainly displayed on the top surface of the roadcut where the metagabbro truncates strong linear (~N-S) and planar fabrics in a garnet-sillimanite xenolith (?). It follows that the fabric is older than ca 1106 Ma and most likely represents Shawinigan (ca 1210-1160 Ma) penetrative deformation. This is one of the few examples of unequivocal pre-Ottawan fabrics in the Adirondack Highlands. Within the metagabbro, and exposed near the truncated contact, are planar surfaces that exhibit strong E-W lineation defined by elongate mineral grains. This lineation parallels Ottawan fabrics and is interpreted as due that that event. Across the highway, the metagabbro thins and is exposed at the top of the roadcut as a narrow (~0.5 m) body hydrothermally altered mylonite surrounded by metasediments. This exposure may represent a fold nose or a neck resulting from attenuation. Whatever structure is involved is somewhat immaterial compared to the fact that at this locality the metagabbro exhibits mylonitic foliation and lineation parallel to regional and local Ottawan trends thus substantiating that the Ottawan Orogeny was strongly felt in the eastern Adirondack Highlands.

d) As one proceeds north at road level from the metagabbro contact, a series of metasediments are encountered. The first of these is the garnet-sillimanite (+ feldspar, quartz, and graphite) with a fabric orientation that now parallels E-W regional Ottawan structures. Presumably the fabric was originally formed in the Shawinigan but was rotated into parallelism by intense Ottawan tectonism. Note the manner in which the marbles have torn through other lithologies that now sit within them as tectonic “fish”. Beyond the marble unit, a schistose unit shows a sulfidic stain. Close inspection shows that these rocks represent highly fractured equivalents of the garnet-sillimanite unit with which it is commonly associated. It is also the horizon that was mined for graphite in the late 1800’s and into the 20th century. The unit is known as the Dixon Schist; hence Dixon-Ticonderoga pencils. Presumably the fracturing provided pathways for fluids that deposited sulfides and perhaps graphite. Given the ductile nature of the Ottawan Orogeny, this fracturing must have been a late event in the geologic history of the area.

e) At the north end of the roadcut there are excellent exposures of the garnet-sillimanite unit with its distinctive, pea-size garnets. Although these garnets exhibit the pale-lavender color commonly associated with spessartine, there is very little of this molecule present and the garnets are dominantly almandine. Many garnets have spiral arms and almost all are zoned with clear rims and ilmenite-rich cores. This difference may be the result of Shawinigan vs. Ottawan growth.

85.8 Park along right of highway near intersection with Blue Goose Road

STOP 8. Ca 1350 Ma Tonalite (15 Min.)

Within the Adirondacks, the ca 1350-1250 Ma suite of tonalitic to granodioritic rocks is restricted to the southern and easternmost regions (Fig.5). Identical lithologies and ages are found in the Green Mts., Vermont and in the western Central Metasedimentary Belt. Together this group is known as the Adirondack-Dysart-Green Mt suite. Its strongly calcalkaline signatures attest to its origin in within a group of composite, outboard arcs. The rocks themselves tend to be compositionally homogeneous and consist of approximately 25% quartz, 80% intermediate plagioclase (~AN_{35}), and 10% mafics. They are characterized by their gray color and the presence of disrupted amphibolite layers thought to be disrupted synemplacement gabbroic dikes.

89.1 Junction Rts. 4 & 22 in Whitehall
96.2 Turn left into parking area

STOP 9. Kinematic Indicators & Straight Gneiss (30 Mins)

This long outcrop provides some of the best exposures of mylonitic rocks and kinematic indicators in the Adirondacks. At its north end it exposes mylonitic metapelitic migmatites that are folded around E-W recumbent isoclinal axes that are visible only in minor folds. Upright, relatively open folds that are approximately coaxial with the E-W isoclines refold them. Down-dip ribbon lineations are readily apparent on foliation surfaces and trend ~E-W. Zircons from leucosomes in these rocks have been dated at this locality and show Elzevirian cores, Shawinigan mantles, and Ottawan overgrowths, and are consistent with the fluxing of Ottawan anatexis by Ottawan fluids.

Towards the south end of the roadcut the mylonitic migmatites are intruded by pink megacrystic granite thought to be of AMCG (ca 1155 Ma) age. The granite is variably deformed into a L-S tectonite gneiss defined by elongate feldspar and quartz grains. Foliation in the gneiss generally dips gently SE (Figure 8), although in some areas, the foliation wraps around weakly deformed mafic boudins. Lineations trend SE, typically down-dip of the foliation. The fabric in this granite is similar to those in the surrounding metapelite units.

This granitic gneiss unit is important for understanding the late stage deformation of this region because it is younger than the surrounding metapelites and presumably did not experienced as complex a deformational history. Measurements of strained porphyroclasts in the granitic gneiss at this location generally indicate near plane strain (2D) deformation of the granite, although some samples fall into the constrictional field (Figure 9). We note that in some samples it was difficult to accurately measure strain because of high strain (S_1:S_3 ratios as high as 50:1), large grain sizes, and complex deformation. However, the best-constrained samples indicate plane strain. As such, we interpret this deformed zone as one that has experienced bulk plane strain after the intrusion of the megacrystic granite.

Strained feldspar porphyroclasts in this granitic gneiss also allow an excellent opportunity to assess the kinematics of this zone. An analysis of >100 tailed feldspar porphyroclasts indicates that ~45% indicate a top-SE sense of shear (normal sense offset in current orientation) based on the asymmetry of sigma-type porphyroclasts (Figure 10), while 25% indicate a top-NW sense of shear (Kowalkoski et al., 2008). The remaining 30% of porphyroclasts with clear tails were symmetrical. When ranked for the confidence of the kinematic indicator, the top-SE sense of shear is even more dominant. These data could be interpreted in several ways. Given the predominance of top-SE kinematic indicators, this zone could be interpreted as a region of normal ductile shear that post-dates the emplacement of the granite (post 1155 Ma?). This extension may have overprinted early thrust displacement, which would explain the presence of top NW sense indicators, although it may be unlikely that early kinematic indicators would be preserved. Alternatively, the presence of both top–SE and NW as well as porphyroclasts with symmetrical tails could also be interpreted as a zone of pure shear that produced thinning and extension of the lower to middle crust.

Based on these preliminary studies, we hypothesize that the last stage of ductile deformation within this zone of the eastern Adirondacks involved extensional deformation, possibly along a SE-dipping normal detachment fault. It is possible that this extension was coeval with the NW-dipping ca 1045 Ma Carthage-Colton Zone that dropped the Lowlands down to the northwest and into juxtaposition with the Highlands. Ongoing geochronologic studies will provide new information of the timing of deformation in this zone. We also suggest that this eastern zone of extension may be an along-strike continuation of the Tawachiche Shear Zone in Quebec (Fig. 11) that dropped the Montauban-La Bostonaiss arc down to the east and into juxtaposition with the Morin Terrane (Nadeau et al, 1992, Corrigan and van Breemen, 1999, Bickford et al, 2008). The picture that emerges is one of late-stage orogen collapse affecting the Adirondack-Morin Terranes at the end of the Ottawan Orogeny. The collapse, which was accompanied by emplacement of Lyon Mt Granite, gave rise to a symmetrical core complex or gneiss dome on the scale and style of the Shuswap Complex.

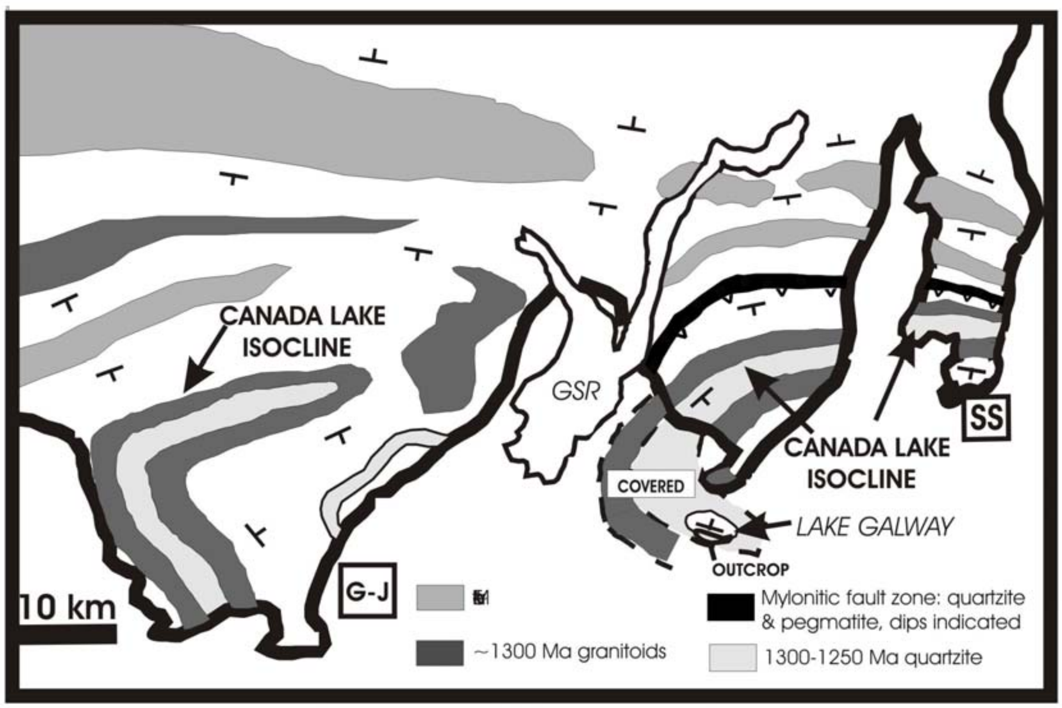

Fig. 7. Schematic structure of the southern Adirondacks showing Canada Lake isocline and eastern repetition

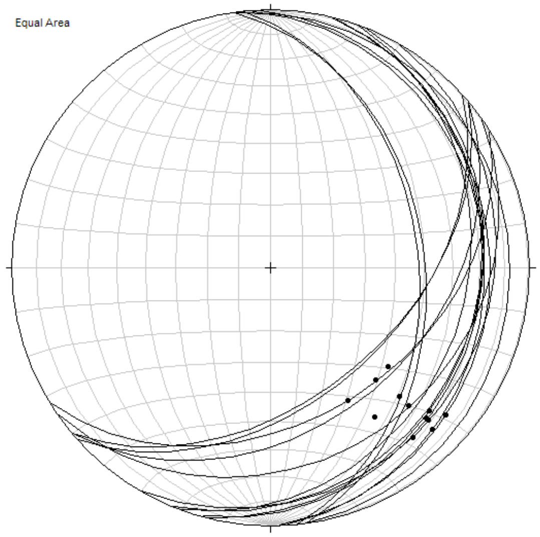

Figure 8. Stereographic projection of foliations and lineations at stop 9. Foliations typically dip gently southeast or east. Lineations plunge down dip.

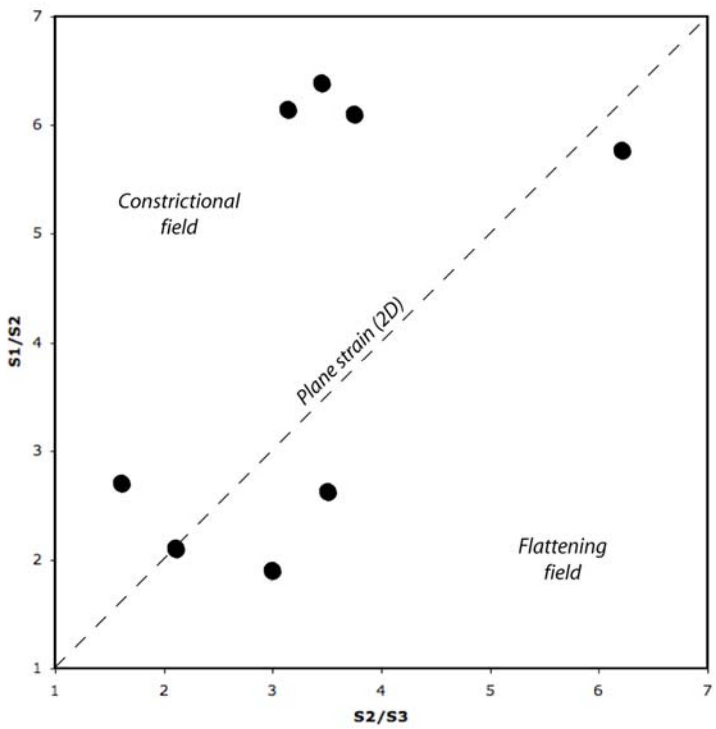

Figure 9. Flinn diagram of samples from the granitic gneiss unit at stop 9. Strain in the granitic gneiss is largely apparent plane strain (2D) although some samples fall in the constrictional field.

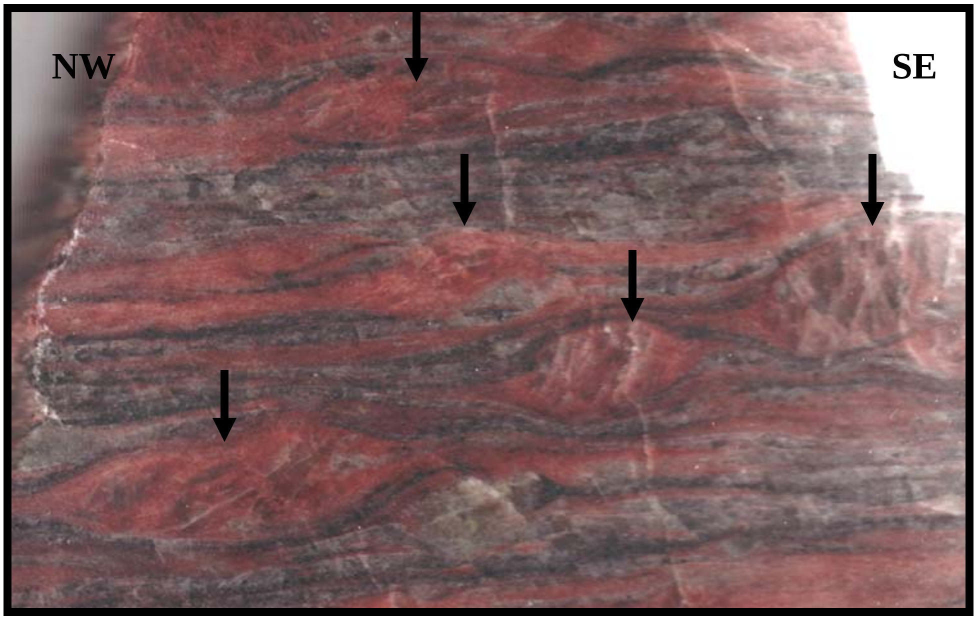

Figure 10. Photograph of slab of granitic gneiss from sample 9 cut perpendicular to foliation and parallel to the lineation. The slab shows a number of K-feldspar porphyroclasts with asymmetric tails that predominantly indicate a top-SE sense of shear (arrows). Image is ~10 cm across.

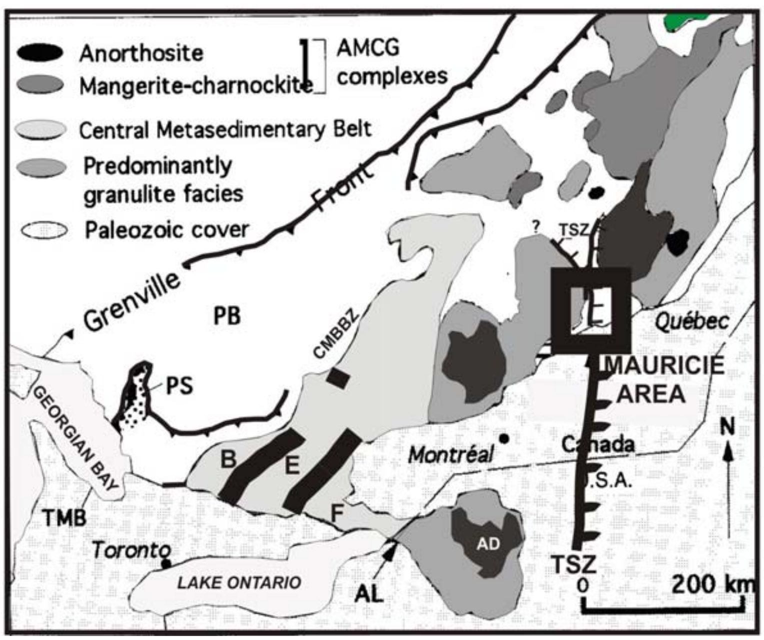

Figure 11. Schematic map of the regional setting of the Adirondacks Highlands (AD) and Lowlands (AL). PS – Parry Sound; B – Bankcroft Terrane; E – Elzevir Terrane; F – Frontenac Terrane; CMBBZ – Central Metasedimentary Boundary Zone; TSZ – Tawachiche Shear Zone and its along-strike projection to the south. Black box encloses the Mauricie are in which lies the town of Shawinigan.

96.6 Stop light at junction of Rts. 4 & 22 at Comstock
97.1 Park along right shoulder of road

STOP 10. Mylonitic Straight Gneiss (20 Min)

This roadcut contains excellent straight gneisses of several lithologies as well as some unusual rock types. Near its SE end are superb mylonitic straight gneisses of sillimanitic migmatitic metapelites. Pegmatites are in various stages of disruption. To the north these become interlayered with garnetiferous gabbroic anorthosite that closely resembles the metapelite leoucosomes except for the presence of hornblende and coarser and more abundant garnets. Gabbroic layers are also present as well as a peculiar aluminous ultramafic that contains diaspore. A whole rock analysis yields SiO_2 = 33%, Al_2O_3 = 17.5 %, Fe_2O_3 = 85 MgO = 24.32%, and CaO = 3%. Calcsilicates are found on both sides of the highway, but on the west side contain pods and layers of orange grossularite-diopside-wollastonite skarn. Nearby these is a discontinuous layer consisting of ~80% red garnet. These types of lithologies are indicative the presence of fluids at high temperature.

99.1 Intersection with Flat Rock Road. Unconformity with Potsdam sandstone expos
100.1 Junction Rt. 149 and Rt. 4 in Ft. Ann. Turn right (west) at traffic light; proceed 20 miles to Lake George Village.

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GEOLOGY AND MINING HISTORY OF THE BARTON GARNET MINE, GORE MT. AND THE NL ILMENITE MINE, TAHAWUS, NY AND A REVIEW OF THE MACINTYRE IRON PLANTATION OF 1857

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INTRODUCTION

This field trip examines the geology and history of mining at two types of ore deposits, one a metal and one an industrial mineral. Both are strongly identified with the Adirondack Mountains of New York and both have histories that extend for more than a century. While mining activities began at both in the nineteenth century, only the Barton garnet mining venture has remained in continuous operation. Attempts, successful and unsuccessful, to exploit the ore at Tahawus occurred sporadically through the nineteenth and twentieth century. During this time the minerals that were the target and gangue at Tahawus essentially reversed roles.

BARTON GARNET MINE, GORE MT.

The Barton Mines Corporation open pit mine is located at an elevation of about 800 m (2600 ft) on the north side of Gore Mountain. For 105 years, this was the site of the world's oldest continuously operating garnet mine and the country's second oldest continuously operating mine under one management. The community at the mine site is the highest self-sufficient community in New York State. It is 16 km (10 mi) from North Creek and 8 km (5 mi) from NY State Route 28 over a Company-built road that rises 91 m (300 ft) per mile. This road, like others in the vicinity, is surfaced with coarse mine tailings. About eleven families can live on the property. The community has its own water, power, and fire protection systems. On the property are the original mine buildings and Highwinds, built by Mr. C.R. Barton in 1933 as a family residence.

Garnet is used in abrasive blasting, coated abrasives, metal, ceramic, and glass polishing, water filtration, waterjet cutting and has been used to remove the red hulls from peanuts (Fig. 1). Paint manufacturers add garnet to create non-skid surfaces and television manufacturers use it to prepare the glass on the interior of color picture tubes prior to the application of the phosphors. However, this latter use is decreasing with the advent of LCD and other flat screen technologies. Garnet, including Barton's garnet, has advantages relative to other abrasive materials. Garnet lacks toxic elements and is chemically inert. Garnet provides greater cutting speed, less dust, and lower volume requirements than competing abrasives. Overall, however, garnet comprises only a small portion (~2%) of the abrasives market (Fig. 2).

The markets currently served by Barton Mines Co. LLC are, in decreasing order, waterjet cutting, television glass lapping, finishes, and abrasive coatings. Abrasive airblasting and water filtration media are minor. Barton Mines supplies 10 plants that produce abrasive paper and cloth in New York, Virginia, Massachusetts, Michigan, Minnesota, Mississippi, Ohio, and Pennsylvania. Barton provides virtually all of the garnet exported by the United States (13,300 tons in 2005) (Evans and Moyle, 2008). Domestic garnet consumption was 50,300 tons in 2005 (Olson, 2006). In 2004, the price for crude concentrates ranged from $50-150/t, refined garnet was $60-400/t, and garnet used for waterjet cutting was $300-600/t.

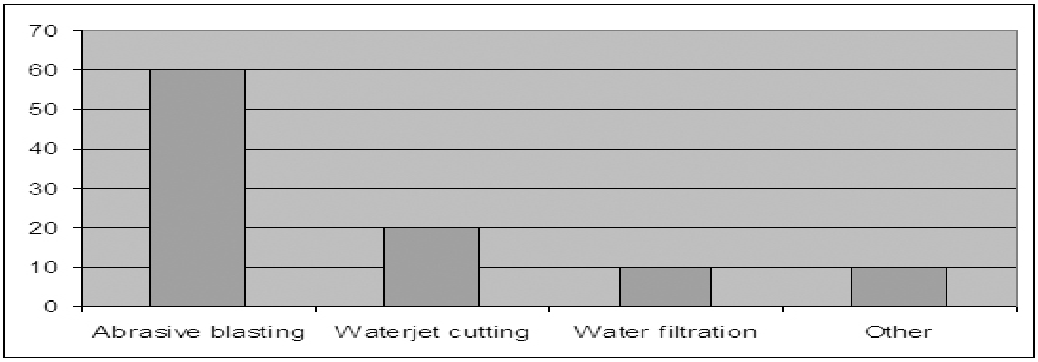


Figure 1. Chart showing principal uses of garnet, in percent, in the world (Olson, 2004)

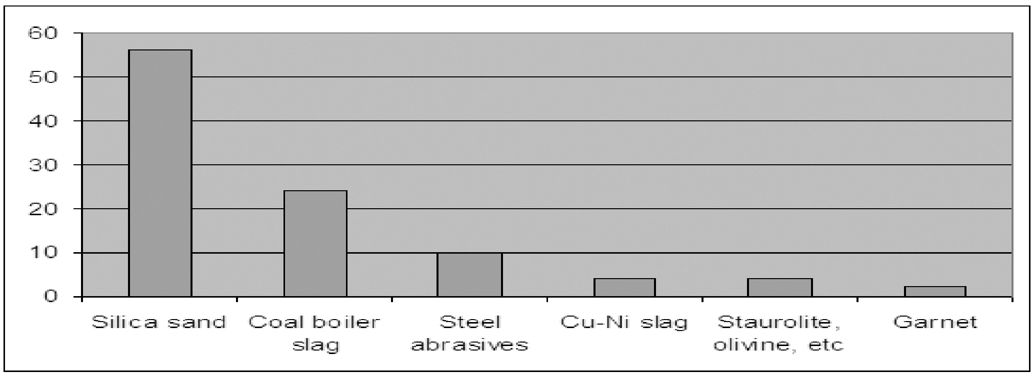


Figure 2. Chart showing U.S. abrasive demand, in percent.

Garnet has been designated as the official New York State gemstone. Barton produces no gem material but collectors are able to find rough material of gem quality. Stones cut from Gore Mountain rough material generally fall into a one to five carat range. A small number of stones displaying asterism have been found. Garnets from this locality are a dark red color with a slight brownish tint. Special cutting schemes have been devised for this material in order to allow sufficient light into the stone.

History

The early history of the Barton garnet mine has been compiled by Moran (1956) and is paraphrased below. Mr. Henry Hudson Barton came to Boston from England in 1846 and worked as an apprentice to a Boston jeweler. While working there in the 1850's, Barton learned of a large supply of garnet located in the Adirondack Mountains. Subsequently, he moved to Philadelphia and married the daughter of a sandpaper manufacturer. Combining his knowledge of gem minerals and abrasives, he concluded that garnet would produce better quality sandpaper than that currently available. He was able to locate the source of the Adirondack garnet stones displayed at the Boston jewelry store years before. Barton procured samples of this garnet, which he pulverized and graded. He then produced his first garnet-coated abrasive by hand. The sandpaper was tested in several woodworking shops near Philadelphia. It proved to be a superior product and Barton soon sold all he could produce.

H.H. Barton began mining at Gore Mountain in 1878 and in 1887, bought the entire mountain from the State of New York. Early mining operations were entirely manual. The garnet was hand cobbed *i.e.* separated from the waste rock by small picking hammers and chisels. Due to the obstacles in moving the ore, the garnet was mined during the summer and stored on the mountain until winter. It was then taken by sleds down to the railroad siding at North Creek whence it was shipped to the Barton Sandpaper plant in Philadelphia for processing. The "modern" plant at Gore Mountain was constructed in 1924. Crushing, milling, and coarse grading was done at the mine site. In 1983, the Gore Mountain operation was closed down and mining was relocated to the Ruby Mountain site, approximately 6 km (4 mi) northeast, were it continues at present.

Mining and milling

The mine at Gore Mountain is approximately 2 km. in length in an ENE-WSW direction. The ore body varies from 15 m (50 ft) to 122 m (400 ft) and is roughly vertical. Mining was conducted in benches of 9 m (30 ft) using standard drilling and blasting techniques. Oversized material was reduced with a two and one-half ton drop ball. The ore was processed through jaw and gyratory crushers to liberate the garnet and then concentrated in the mill on Gore Mountain. Garnet concentrate was further processed in a separate mill in North River at the base of the mountain. Separation of garnet was and is accomplished by a combination of concentrating methods including heavy media, magnetic, flotation, screening, tabling and air and water separation. Processes are interconnected and continuous or semi-continuous until a concentrate of 98% minimum garnet for all grades is achieved (Hight, 1983). Finished product ranges from 0.6 cm to 0.25 micron in size.

C*haracteristics of Gore Mountain garnet*

The garnet mined at Gore Mountain is a very high-quality abrasive. The garnets display a well-developed tectonic parting that, in hand specimen, looks like a very good cleavage. This parting is present at the micron scale. Consequently, the garnets fracture with chisel-like edges yielding superior cutting qualities. The garnet crystals are commonly 30 cm in diameter and rarely up to 1 m with an average diameter of 9 cm (Hight, 1983) The composition of the garnet is roughly 43% pyrope, 40% almandine, 14% grossular, 2% andradite, and 1% spessartine (Levin, 1950; Harben and Bates, 1990). Chemical zoning, where present, is very weak and variable (Luther, 1976). The garnet has been so well analyzed isotopically that it is frequently used as an ^{18}O / ^{16}O standard (Valley et al., 1995). Typical chemical analyses of the garnet are presented in Table 1. Hardness of the garnet is 7.5 and the average density is 3.95 gm/cm^3.

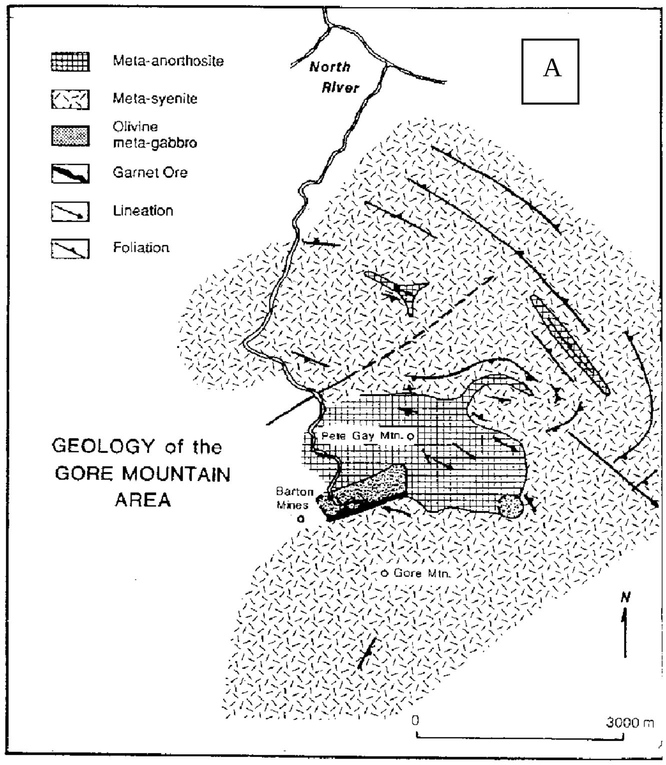
Table 1. Electron Microprobe analyses of Gore Mt. garnet (almandine-pyrope) normalized to 8 cations and 12 anions. * Calculated by charge balance (Kelly and Petersen, 1993).

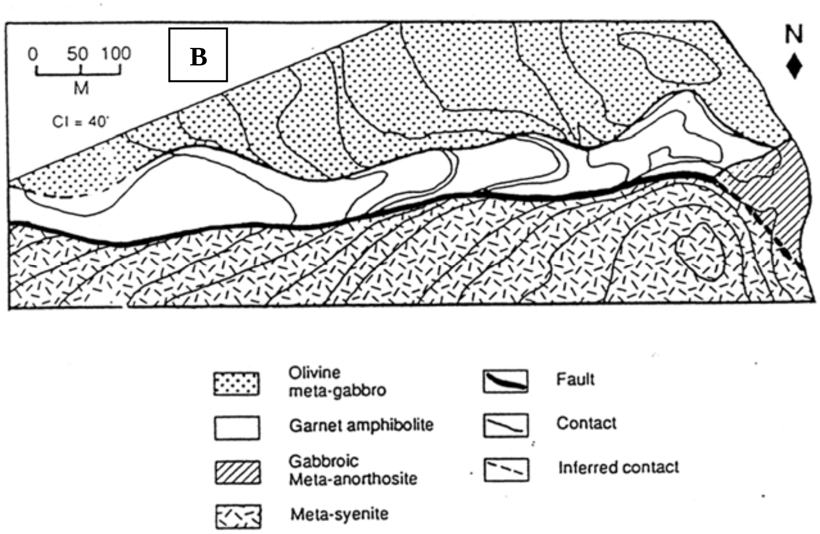
Oxide Weight Percent	#29	#41
SiO_2	39.43	39.58
Al_2O_3	21.40	21.20
TiO_2	0.05	0.10
FeO*	22.80	24.45
Fe_2O_3*	1.44	0.72
MgO	10.65	9.60
MnO	0.48	0.74
CaO	3.85	3.97
Na_2O	0.00	0.00
K_2O	0.00	0.00
Total	100.09	100.36

Geology

The garnet mine is entirely hosted by a hornblende-rich garnet amphibolite unit along the southern margin of an olivine meta-gabbro body (Fig. 3). The garnet amphibolite grades into garnet-bearing gabbroic meta-anorthosite to the east. To the south the garnet amphibolite is in contact with a meta-syenite; a fault occurs parallel to this contact in places.

Figure 3. Geologic maps of Barton garnet mine
(A.modified from Bartholome, 1956, B. Goldblum & Hill, 1992)





The olivine meta-gabbro bordering the ore zone is a granulite facies lithology with a relict subophitic texture. Preserved igneous features, faint igneous layering, and a xenolith of anorthosite have been reported in the meta-gabbro (Luther, 1976). Prior to metamorphism, the rock was composed of plagioclase, olivine, clinopyroxene and ilmenite. During metamorphism, coronas of orthopyroxene, clinopyroxene and garnet formed between the olivine and the plagioclase and coronas of biotite, hornblende and ilmenite formed between plagioclase and ilmenite (Whitney & McLelland, 1973, 1983). The contact between the olivine meta-gabbro and the garnet amphibolite ore zone is gradational through a narrow (1 to 3 m wide) transition zone. Garnet size increases dramatically across the transition zone from less than 1 mm in the olivine meta-gabbro, to 3 mm in the transition zone, to 50 to 350 mm in the amphibolite (Goldblum and Hill, 1992). This increase in garnet size coincides with a ten-fold increase in the size of hornblende and biotite, the disappearance of olivine, a decrease in modal clinopyroxene as it is replaced by hornblende, and a change from green spinel-included plagioclase to white inclusion-free plagioclase (Goldblum and Hill, 1992). Mineralogy in the garnet amphibolite ore zone is mainly hornblende, plagioclase and garnet with minor biotite, orthopyroxene, and various trace minerals. In both the olivine meta-gabbro and the garnet amphibolite, garnet content averages 13 modal percent, with a range of 5 to 20 modal percent (Luther, 1976; Hight, 1983; Goldblum, 1988).

The garnet amphibolite unit is thought to be derived by granulite facies metamorphism of the southern margin of the olivine meta-gabbro. At the west end of the mine, a garnet hornblendite with little or no feldspar is locally present. This rock may represent original ultramafic layers in the gabbro (Whitney et al., 1989). In the more mafic portions of the ore body, the large garnet crystals are rimmed by hornblende up to several inches thick. Elsewhere, in less mafic ore, the rims contain plagioclase and orthopyroxene. Chemical analyses of the olivine meta-gabbro and garnet amphibolite show that the garnet ore was derived by isochemical metamorphism, except for an increase in the H_2O and fO_2, of the olivine meta-gabbro (Table 2; Luther, 1976). A strong, consistent lineation and weak planar fabric coincide with the zone of large garnet crystals and are an important feature of the garnet ore zone (Goldblum and Hill, 1992). The lineation is defined by parallel alignment of prismatic hornblende crystals, elongate segregations of felsic and mafic minerals, plagioclase pressure shadows, and rare elongate garnet. The foliation is defined by a slight flattening of the felsic and mafic aggregates.

Origin of garnet

Although the garnet crystals in the ore zone at Gore Mountain are atypical in size, the modal amount of garnet is not unusually high for Adirondack garnet amphibolites. Garnet amphibolite that is texturally and mineralogically similar occurs elsewhere in the Adirondacks, usually on the margins of gabbroic rock bodies. The ore at the currently operating Barton Corporation mine at Ruby Mountain, for example, is of the same tenor but the garnets rarely are larger than 2.5 to 5 cm.

Petrologic studies (Buddington, 1939, 1952; Bartholome, 1956, 1960; Luther, 1976; Sharga, 1986; Goldblum, 1988; Goldblum and Hill, 1992) have agreed that the growth of the large garnets is related to a localized influx of water along the margin of the granulite facies olivine meta-gabbro body. The Gore Mountain garnets are chemically homogeneous suggesting that (a) the garnets grew under conditions in which all chemical components were continuously available and the (b) temperature and pressure conditions were uniform during the period of garnet formation. A zone of high fH_2O along the southern margin of the original gabbro body may have enhanced diffusion and favored growth of very large garnets and thick hornblende rims at the expense of plagioclase and pyroxene. Luther (1976) speculates that physical and chemical conditions were favorable for the <u>growth</u> of garnet but poor for the <u>nucleation</u> of garnet so that the garnet crystals that did nucleate grew to large size. The presence of volatiles, particularly H_2O, promotes the growth of large crystals by aiding transport of components. The formation of the garnets has been dated at *ca.*1050 Ma (Mezger, *et al.*, 1992).

Recognition that the garnet ore body and deformation fabric coincide with the southern margin of the olivine meta-gabbro body led Goldblum and Hill (1992) to hypothesize that the high fluid flow required for growth of large garnet crystals was the result of ductility contrast at a lithologic contact during high-temperature shear zone deformation. The olivine meta-gabbro is a granulite facies rock with a poorly developed foliation and little evidence of ductile deformation. In the transition zone between the olivine meta-gabbro and the garnet amphibolite, increased ductile deformation resulted in grain-size reduction of plagioclase and pyroxene. Microstructures in plagioclase in the transition zone indicate plastic deformation, and the concurrent modal increase in hornblende indicates an influx of fluid. Fabric development and hydration are most apparent in the garnet amphibolite of the ore zone. According to Goldblum and Hill (1992), the olivine meta-gabbro remained competent and initially deformed by brittle processes along its southern margin while the adjacent feldspar-rich meta-syenite and gabbroic meta-anorthosite deformed plastically during deformation at amphibolite facies conditions. Initial grain-size reduction by cataclasis along the margin of the meta-gabbro allowed hydration and metamorphism to produce the garnet amphibolite. As the hydrated ore body replaced the olivine meta-gabbro, ductile deformation mechanisms replaced cataclasis. During metamorphism, the garnet amphibolite was likely a high-strain zone of reaction-enhance ductility. Eventually, metamorphic reactions apparently outpaced the rate of deformation and grain coarsening impeded ductile deformation processes (Goldblum and Hill, 1992).

Inclusions in Gore Mountain garnet

Gore Mountain garnet hosts a number of inclusion types, the most common of which is acicular rutile. Many rutile needles are crystallographically controlled and the asterated specimens (Figure 4a) described earlier may be due to the orientation of these inclusions parallel to {111} of the garnet. Other common solid inclusions include pyrite, plagioclase, pyroxene, hornblende, ilmenite, apatite, and biotite (Valley, et al, 1995).

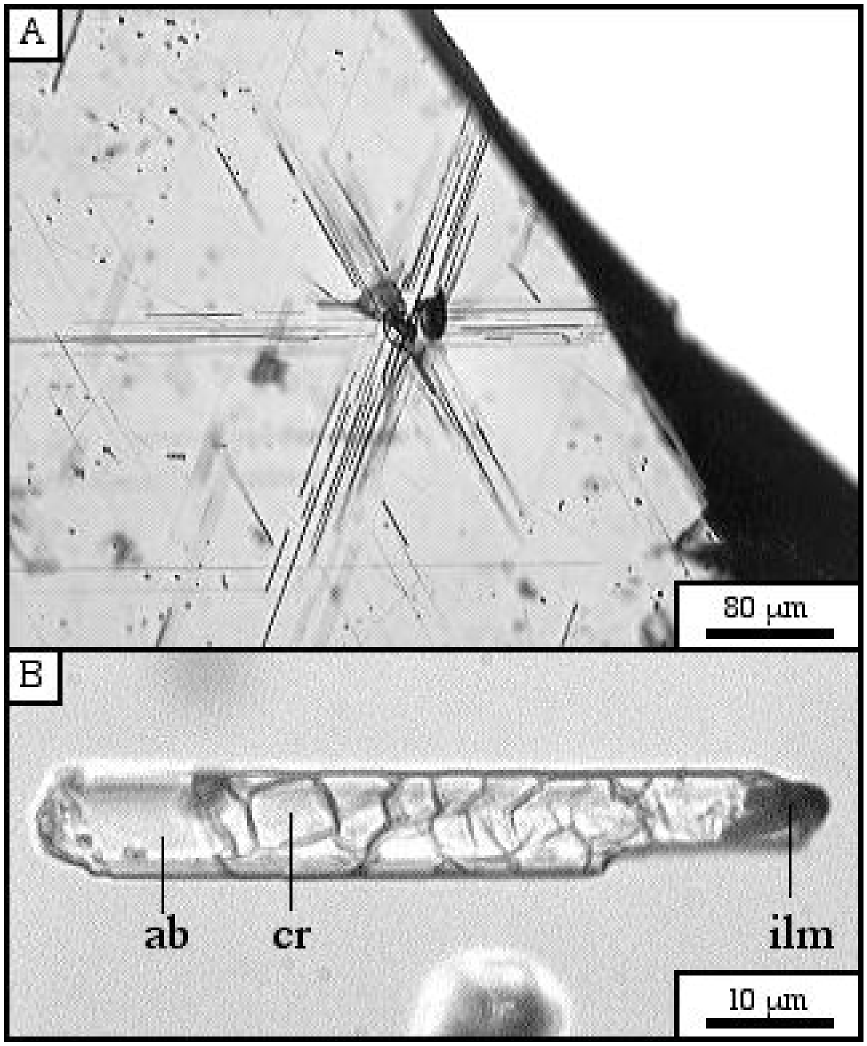


Figure 4. (a)Crystallographically-controlled rutile needles in garnet
(b) Cristobalite, ilmenite, albite inclusion

Darling et al. (1997) report a most unusual type of multiple solid inclusion containing the low pressure, very high temperature SiO_2 polymorph, cristobalite (Figure 4b). This phase is accompanied by albite and a small quantity of ilmenite. The cristobalite is easily recognized by its inherent fractures, formed by the 5% volume decrease upon transforming from the beta-phase at temperatures between 260 and 270^{o}C. Darling et al. (1997) propose the cristobalite + albite + ilmenite inclusions began as small water-rich melt inclusions which then experienced diffusive loss of water. This led to an internal pressure decrease (under nearly isochoric, isothermal conditions) to the point where cristobalite, instead of quartz, crystallized in the melt inclusions. It should be noted that identical cristobalite-bearing multiple-solid inclusions also occur at the former Hooper and North River Mines (Charles et al., 1998) so their formation is not unique to Barton Mine garnet. The most remarkable outcome is that the cristobalite never reconstructively transformed to quartz even during protracted cooling from starting conditions of approximately 800^{o}C. Darling et al. (1997) infer the absence of water was the primary reason for the preservation of cristobalite.

Fluid inclusions are rare in Gore Mountain garnet despite the importance of water in the formation of garnet amphibolite as well as large crystal sizes. Ironically, the most common fluid inclusion in Gore Mountain garnet is CO_2–rich and is texturally secondary. These inclusions, like those in many other Adirondack rocks, most likely formed along the retrograde path following peak metamorphic conditions. Precisely how CO_2-rich inclusions can form in garnet (or more commonly quartz) is unknown as neither mineral is soluble in liquid CO_2. Two possible explanations include: 1) low temperature mineral growth from the aqueous portion of an immiscible H_2O-CO_2 fluid while trapping CO_2, or 2) diffusive loss of H_2O from an original mixed H_2O-CO_2 inclusion.

NL ILMENITE MINE, TAHAWUS, NY

History

The history of mining operations in the vicinity Tahawus extends from the early nineteenth century to the end of the twentieth. The following account is drawn from Stephenson (1945) and National Lead Company literature (Anon. 1963). Gradual expansion of industrial development in the United States in the nineteenth century drove exploration for mineral resources into increasingly "back-country" locations. Such was the case when the iron ore in the region around Sanford and Henderson Lakes was discovered by Europeans in 1826 (a Native American led them to the ore). They were purportedly led to the deposit over what is now the Indian Pass Trail. The ore mineral sought was magnetite. The original discoverers were not initially aware of the presence of ilmenite in the ore. Since 1826, great sums of money have been spent in attempts to bring the resource into production. The first and ultimately unsuccessful try was made by the original discoverers, Archibald MacIntyre, Duncan and Malcolm MacMartin, and David Henderson. Their first blast furnace was built in 1838 and a "new" furnace was constructed in 1854. However, the venture completely failed owing to a number of circumstances including the accidental killing of Henderson in 1845, a flood in 1856, an economic recession in 1857 and, most importantly, the presence of titanium in the ore and transportation difficulties.

Beginning in 1894, another attempt was made to exploit the ore as a source of iron. The MacIntyre Iron Company planned a railroad (never constructed) and did extensive diamond drilling. A crushing plant and concentrator were constructed in 1912-13 and ore was hauled over very bad roads to blast furnaces in Port Henry. However, there was insufficient interest in the deposit and it languished.

In 1941, the deposit was acquired by National Lead Company (later NL Industries, Inc.). Production began in 1942. The target mineral in this case was ilmenite. Iron (magnetite) was a co-product. In 1908, a French metallurgist working for the MacIntyre Co. discovered that titanium dioxide made an excellent white paint pigment. With the advent of Would War II, TiO_2 became a critical strategic resource. It was used in paint, paper, rubber and ceramics but it was an essential raw material in chemical smoke screens. To bring the mine into production required the construction of 8.5 miles of road, a 42 mile power line from Ticonderoga, 29 miles of railroad from North Creek, housing for 180 families and 160 single men, schools, fire department, medical services, and recreational facilities. Over the years, 4,564 men and women worked at the mine.

The mine continued in production until 1982 when the closure of the NL ilmenite processing plant in New Jersey led to the closure of the mine. Magnetite concentrates continued to be shipped from stockpile until 1989. Attempts were made to use the magnetite as blast furnace feed stock but these were not particularly successful. The bulk of the magnetite was sold as ballast, for ferroconcrete, and most importantly, to coal producers for use in a slurry to separate coal from waste rock. The mine site was reclaimed in 2005-2006. Of the 11,000 acres owned by NL (now Kronos Worldwide, Inc.) most was purchased by the Open Space Institute with plans for eventual transfer to the State of New York. The lands will become part of the surrounding Adirondack Park. A total of more than 141 million tons of rock were removed during the course of mining to process sixty million tons of ore. From this ore, more than twelve million tons of ilmenite and seventeen million tons of magnetite were extracted. Thirty-one million tons of ilmenite and twenty-one million tons of magnetite remain.

Mining and milling

The ore in the most recent operation was blasted in forty-foot high benches, hauled to the mill in Euclid trucks where it was reduced from run-of-the-mine size (pieces up to four feet) to minus 7/16 inches by jaw and cone crushers. This product was further reduced to twenty-eight mesh by six foot by twelve foot rod mills. Magnetic separators then removed some of the magnetite which was sent to a sintering plant or stockpile. The ilmenite and remaining gangue minerals were sent to reciprocating tables which retained the heavy ilmenite behind riffles and discharged the lighter (silicate) particles as waste. This ilmenite concentrate was dried and subjected to a final magnetic separation after which it was sent to storage bins or loaded directly onto railroad cars for shipment to a titanium pigment plant. The very fine grained fraction of the ground ore was separated by froth flotation. In this process, reagents in water are added to the feed stock in a processing tank which is agitated by air. The reagents coat the ilmenite only and cause the ilmenite particles to adhere to air bubbles and rise to the surface while the uncoated waste particles fall to the bottom of the tank and are removed. This fine grained ilmenite is collected in the froth on the top of the tank, then dried, loaded or stored.

Characteristics of the Tahawus ore

The ore deposits of the Tahawus region consist of roughly equal amounts of titaniferous magnetite and hemo-ilmenite with ilmenite being slightly more abundant. Lamellae of ilmenite in magnetite originated via subsolidous oxidation-exsolution (Buddington and Lindsley, 1964). Pleonaste spinel is a common exsolution phase in magnetite. Iron sulfides occur as accessory phases. Gangue minerals include plagioclase (10-20%), garnet (gros-alm-and) 3-8%, ortho-and clinopyroxene (4-7%), hornblende (1-3%), biotite (≅ 1%) Heyburn, 1960). Other minor phases include apatite, prehnite, barite, orthoclase, scapolite, "leucoxene", epidote and quartz (Gross, 1968). Both titanomagnetite and hemo-ilmenite form abundant, small, rod-like inclusions in the plagioclase, making the feldspar appear black in hand specimen. The average composition of the titanomagnetite is $Mt_{81}Usp_{19}$ and of the hemo-ilmenite is $Ilm_{94}Hem_{6}$ (Kelly, 1979).

The Sanford Lake District ores occur in two major modes: (1) as lean or disseminated ore in gabbro (gabbroic ore) and (2) as rich ore generally located in anorthosite but locally in gabbro (anorthositic ore). The lean ore within gabbro displays a gradational contact with the host rock with which it is commonly conformably layered (Ashwal, 1978) while in anorthosite, the massive, rich ore has a sharp crosscutting relationship. The gabbro, which bears the lean ore, also has a crosscutting relation to the anorthosite. With the exception of nelsonitic rocks near Cheney Pond (discussed below), the P_20_5 concentration is very low. Whole rock chemical analyses are given below.

	Tahawus gabbro	Tahawus massive ore	Woolen Mill gabbro
SiO_2	39.04	4.59	45.59
TiO_2	6.78	18.58	3.49
Al_2O_3	13.09	5.48	14.23
Fe_2O_3	19.09	nd	18.42
FeO	nd	66.37	nd
MnO	0.24	0.28	0.27
MgO	5.31	3.39	3.07
CaO	9.77	0.31	9.25
Na_2O	2.02	0.22	2.63
K_2O	0.66	0.09	0.79
P_2O_5	0.23	0.01	1.26
V_2O_5	nd	0.45	nd
H_2O	0.03	0.01	0.70
Total	96.26	100.06	100.14

nd = not determined

Geology and origin of the ore

Open pit mining and diamond drilling have demonstrated that the ore is concentrated in westward dipping lenses measuring 600-700 meters long and 150-300 meters wide. This configuration could be the result of crystal settling, intrusion, or accumulation of immiscible oxide-rich liquid. Late differentiates of the anorthosite liquid tend toward enrichment in iron-titanium oxides giving rise to rocks of increasingly ferrogabbroic composition along with mafic cumulates. The gabbro at Tahawus is similar to the Woolen Mill gabbro and other gabbros, which are typical of late anorthosite differentiates (Bohlen. *et al*. 1992). This suggests that the ores at Tahawus are the result of progressive differentiation from a gabbroic anorthosite parent and that, late in the differentiation process, the magma became so enriched in Fe and Ti that they either precipitate oxide cumulates (Ashwal, 1978) or a (presumably immiscible) Fe-Ti liquid forms (Stephenson, 1945, Kelly, 1979). In the former case, the comformable oxide-rich layers represent cumulates and the cross cutting ore represents mobilized cumulates. In the later case, both the conformable and cross-cutting ore is due to oxide liquid. A third possibility is that magnetite and ilmenite begin to precipitate along with the silicate minerals that form the gabbro but continued fractionation might still result in an oxide liquid at a very late stage.

It has long been argued that liquid immiscibility could not have occurred in the Tahawus ore because of the lack of apatite in the rock (except at Cheney Pond, see below). However, Lindsley (1991) noted that phosphate and Fe-Ti oxides mutually enhance their solubility in silicate liquids. Apatite may not necessarily travel with the immiscible oxide liquid. The key role of apatite, then, may have been to enhance the concentration of an Fe-Ti-rich phase of the residual liquid until immiscibility occurs. The absence of apatite in the resulting ore may not be proof of the lack of immiscibility. Furthermore, it has been shown experimentally that the melting temperatures of Fe oxides are significantly reduced in the presence of carbon (Weidner, 1982). Consequently, the formation of liquid oxide melt at geologically reasonable temperatures is possible.

Cheney Pond nelsonite

Kolker (1980, 1982) describes an occurrence of nelsonite hosted by gabbro just southwest of the Cheney Pond test pit, 2 km west of the Tahawus mill site. This is one of two occurrences of nelsonite in the Adirondacks; the other is located 110 km to the southwest in the village of Port Leyden (Darling and Florence, 1995). Nelsonite near Cheney Pond contains approximately 44-52% magnetite, 4-18% ilmenite, and 24-31% apatite and is the same orebody as that described by Stephenson (1945) except the "gangue silicates" are properly identified as apatite (Kolker, 1982). The host gabbro contains plagioclase, orthopyroxene, clinopyroxene, garnet and abundant magnetite and ilmenite. It is petrographically similar to oxide-apatite gabbro norites (OAGN) described by Owens and Dymek (1992). Nelsonites are believed to form by magmatic immiscibility (Phillpotts, 1967; 1981) or by cumulate processes (Owens and Dymek, 2001), but in both cases the source rocks are either anorthosites or OAGN's. Both rocks occur in the Cheney Pond area but the chemical relationships between the anorthosites, OAGN's, and nelsonites, have not been determined.

Chronology of Tahawus ores

The Tahawus magnetite-ilmenite ores were dated by Silver (1969) and gave an essentially concordant age of 1000 +/- 13 Ma. Chiarenzelli and McLelland (pers. comm., 2002) dated a concentration of magnetite-ilmenite on Forest Road that passes along the north shore of Upper Saranac Lake and got a nearly concordant age of 998 +/- 10 Ma. Recent SHRIMP dating by Bickford and McLelland (pers. comm., 2002) have revealed the presence of abundant *ca*. 1000 Ma zircon overgrowths in *ca.* 1150 Ma anorthositic-gabbroic rocks. Overgrowths in granitoid rocks are generally not so young and cluster around 1050 Ma. McLelland suggests that the explanation for the 1000 Ma ages in the mafic rocks rests upon the fact that they - and the ores - are rich in ilmenite in which Zr is a proxy for Ti.

When ilmenite crystallizes from magma, it accepts small quantities of Zr as a substitute for Ti. Titaniferous pyroxene behaves similarly. Upon cooling from magmatic temperatures, the Zr will tend to reside metastably in the ilmenite unless a pulse of activation energy occurs. Post-1150 Ma there were two such pulses: Hawkeye Granite heating at *ca.* 1100 Ma followed by the Ottawan Orogeny at 1090-1030 Ma. Both of these were high temperature events and tectonism may have been insufficient to promote widespread exsolution at these high temperatures. Notwithstanding, the common occurrence of 1050 rims and discrete metamorphic grains attests to liberation of Zr either by exsolution or more likely by the reaction plag+opx + (mgt-ilm) = gt + cpx + SiO_2 (McLelland, Appendix A, IGCP-304 Guidebook, 1992, Bohlen (ed.)). The involvement of ilmenite as a reactant releases Zr that combines with the SiO_2 to form zircon. In ilmenite-rich rocks, especially the ores, the plagioclase and orthopyroxene are scarce so this reaction is a minor factor and Zr remains fixed. After 1050, the area cools down and Zr becomes increasingly metastable in the ilmenite. At ~ 1000 Ma a late orogenic but low relatively temperature pulse hits the area and Zr exsolution takes place, probably in the presence of SiO_2-bearing hydrothermal solutions. It is at this time that the 1000 Ma overgrowths form. This late pulse is recorded on the Grenville Front Tectonic Zone, in the Green Mts. (Stamford Granite), and in the eastern

Grenville Belt where it is referred to as the Rigolet Pulse (Rivers, 1997). McLelland (pers. comm., 2002) proposes that the ilmenite-rich rocks are dating this event due to the factors described above.

ADIRONDACK IIRON AND STEEL COMPANY

The "new furnace" at Tahawus

The word "plantation" connotes to most people a large antebellum agricultural operation located in the southeastern United States where cotton or sugar cane might have been the principal product. However, the term can equally be applied to certain isolated, more or less self-contained, communities built around iron production in the Pennsylvania, New York and elsewhere. One such example is the community of Tahawus and the operations of the Adirondack Iron and Steel Company from about 1830 until 1857. The requirements of blast furnace technology of the era dictated that ore, water power and vast amounts of fuel be located in close proximity. This generally dictated that the facilities be located in rural, if not remote, areas. The Adirondack Iron and Steel Company works were located on 106,000 acres of forested land, forty-five miles west of Lake Champlain near the source of the Hudson River but over 100 miles north of the limit of navigability. The Town of Adirondac (sic) was built entirely by the Company. It consisted of a church and school, twenty-five houses, a massive boarding house, ice house, carpenter shop, blacksmith shop, saw mill, grist mill and power house (Seely, 1978, 1981). Two farms provided some of the staples needed by the residents such as hay, barley, potatoes, sheep and cattle but due to the climate, flour, salt pork and everything else to supply the community had to be imported from Albany. The ore deposits were at the surface, actually outcropping in the Hudson River and surrounding the community. The ore was mined by open-pit methods. The blast furnace(s) were supported by charcoal kilns, a puddling furnace, magnetic and wet ore separators, and a not-very-successful cast-crucible steel plant (Seely, 1981).

Two blast furnaces were constructed, both charcoal fueled, fed by wooden charging bridges. The first was in operation in 1844 and the second was constructed in 1854. The second furnace, the "new furnace", was sixty feet (including chimney) high and had a hotter blast than the first, a requirement necessitated by the titanium content of the ore. This furnace featured ore stamps on the bridge to reduce the size of the feed stock and a hot blast stove to preheat the blast air stream (Fig. 5). Water power was used at both installations (Fig.6). A dam was key to increase storage capacity. The new furnace had a dam 180 feet long across the Hudson River that stood twenty-five feet high. Because of the large loss of power in line-shaft transmission, the furnace has to be close to the water wheel.

All of these features, the community, the materials, the power supply and the layout of the operation are characteristic of a charcoal iron works of the 19th century. The Adirondack Iron and Steel Company was a true iron plantation. The owners spent about $500,000 on the project and lost it all. The titaniferous quality of the ore and the ever-envisioned, never-built railroad connection to the property contributed greatly to their downfall.

The second of the two blast furnaces built by the Adirondack Iron and Steel Company stands in outstanding condition. It is the most intact iron furnace in the northeastern United States and arguably, in the world (Youngken, 1989). While the iron industry in general has a long history of consuming its past (old technology is fed to the furnace as new technology is introduced) the circumstances of the failure and closing of this operation led to much better preservation. In a short span of time three events transpired which were fatal to the company. First, one of the principal and very active owners, David Henderson, was accidentally killed by an abdominal gunshot wound. Secondly, a flood destroyed the dam at the "lower works" allowing an artificial lake to drain. This lake had been used to move supplies and pig iron by water for several miles downstream from the furnace site. Finally, a "financial panic", i.e. an economic recession, in 1857 caused the mine and furnace to cease operations. The furnace was idled. The community dissipated. The facility stood, essentially undisturbed and un-scrapped, for nearly 150 years.

Even though the Adirondack Iron and Steel Company was located in a very remote part of the mountains, the owners clearly kept abreast of iron technology innovations that occurred in Britain and elsewhere in the United States. Some of these innovations, such as a hot blast stove, were added to the Company's equipment within just few years introduction into the US from England. The new furnace shows many such technological improvements (Fig. 7).

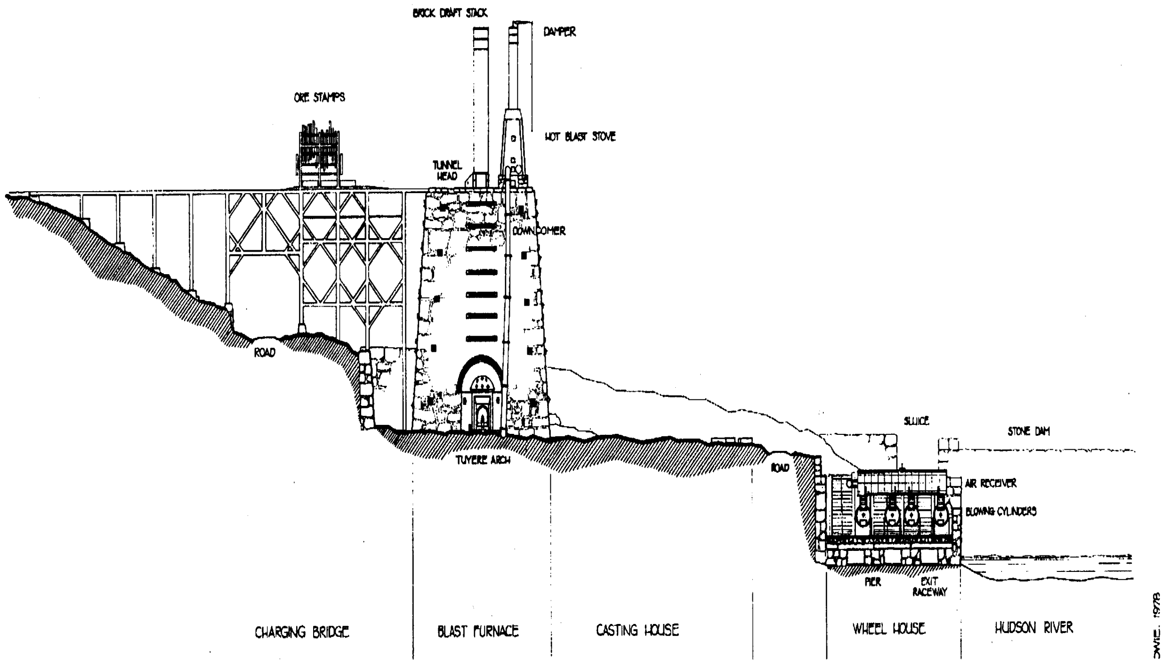


Figure 5. Partially restored sectional view of the "new" furnace and wheel house. (Seely, 1978)

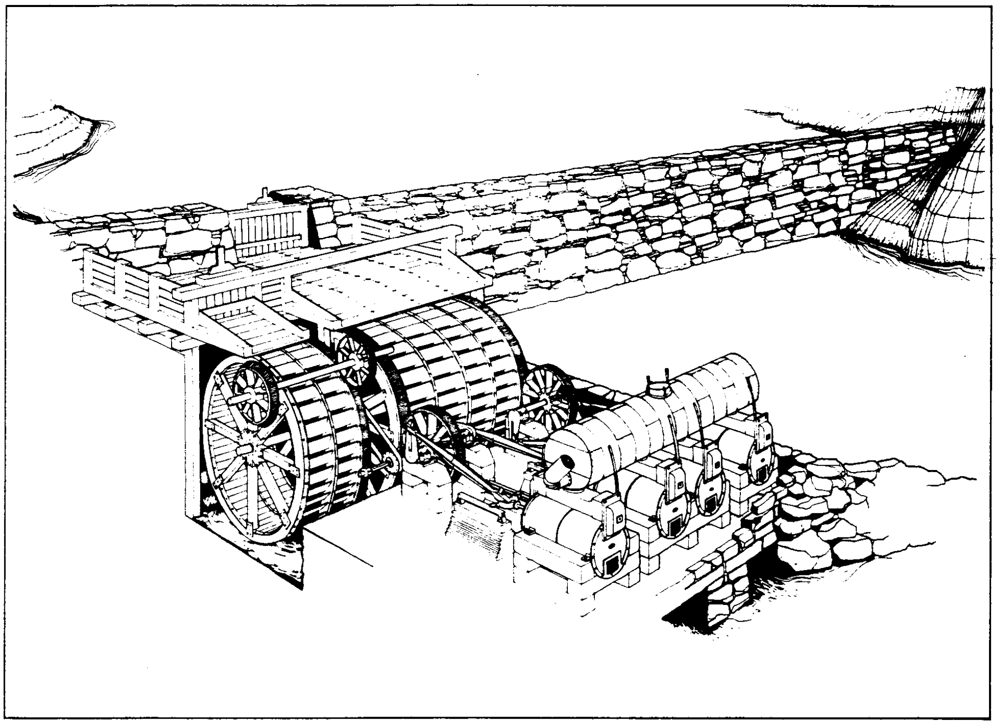


Figure 6. Reconstruction of the dam and blowing engine of the "new" furnace.(Seely, 1978)

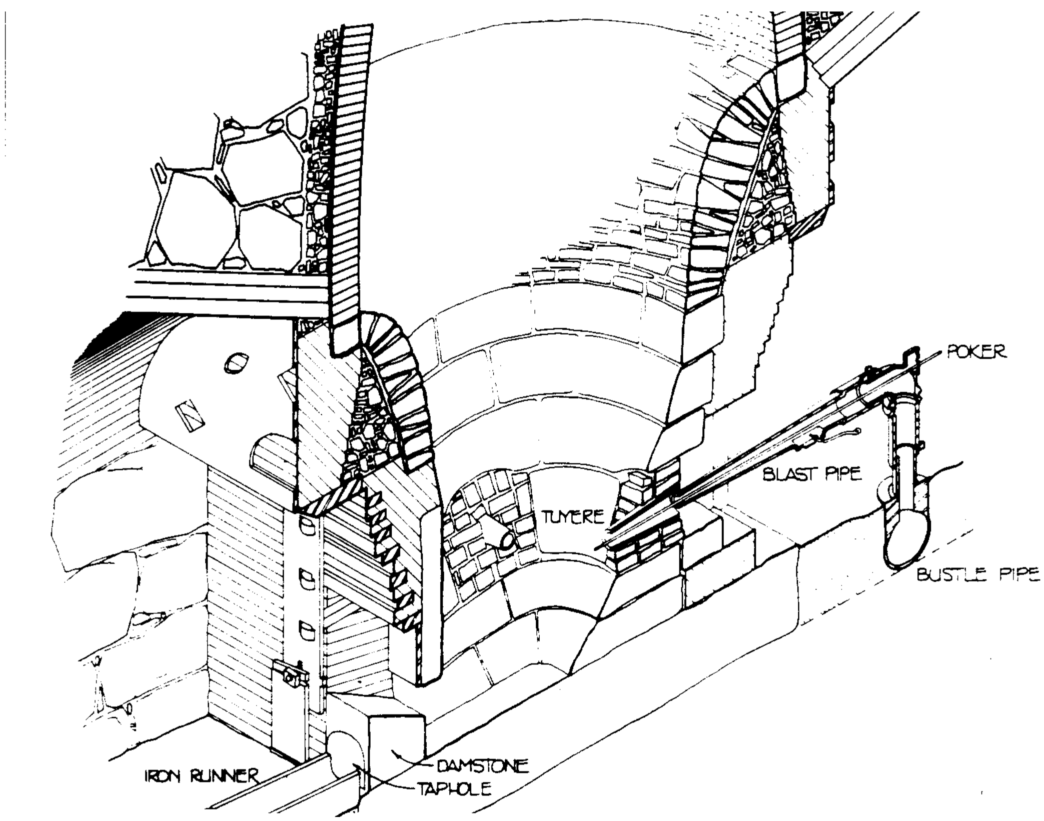


Figure 7. Isometric drawing of "new" furnace, Adirondack Iron & Steel Co.(Seely, 1978)

Many of these first appeared in iron furnaces using mineral fuels in Pennsylvania. Although the owners of the Adirondack Iron and Steel Company recognized that the importation of coal was impossible, they did adapt their charcoal-burning furnaces with up-to-date improvement. First of all, the height of the stack new furnace was forty-five feet with a brick chimney topping that. The old furnace probably was originally constructed at twelve to fifteen feet, and eventually increased to thirty to thirty-five feet. The bosh diameter of the new furnace is twelve feet (seven feet in old furnace). The new furnace had a top-mounted hot-blast stove to increase the temperature of the blast air from ambient to 400-600°F. The old 1844 furnace had a hot blast stove sitting beside it on the ground, and then located on top in 1845. This idea was only introduced in US from Britain in 1840, demonstrating the company's commitment to new technology. Not surprisingly, the introduction of one new technological development drives others. The hotter furnace caused damage to the cast iron tuyeres so these had to be water cooled. The downcomer pipe, carrying hot air from the stove, had to be upgraded to inch-thick cast iron from galvanized sheet metal to accommodate the hot blast. The stack was reinforced with wrought iron bars to improve stability. These bars doubtless enhanced the preservation of the stack. In order to increase the pressure of the blast air, double acting cylinders were installed with adjustable cranks that could vary the stroke from thirty-six to sixty-six inches.

A visit to the site of the new furnace offers a glimpse into an industry that flourished a century and a half ago. The site of the new furnace was cleared and documented in the late 1970s under the auspices of the Historic American Engineering Record Program. NL Industries was closely involved in this study. Although the forest has reclaimed the site, much of the equipment is still well exposed. Since the transfer of the property from Kronos to the Open Space Institute, archeologists from the NYS Department of Conservation and the NYS Museum have cleared the site and begun documentation and restoration.

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ROAD LOG

The Gore Mountain mine of the Barton Mines Corporation is generally open seasonally for visitors. However, it is suggested that visitors call in advance to be sure the mine is open. Arrangements for group tours at any time should be made through Barton Mines Corp., North Creek, NY 12853, (518) 251-2296. The mine at Tahawus and the gate on the road to the development pit at Cheney Pond, currently owned by Kronos Worldwide, Inc., Houston TX, are not open to visitors without prior permission. For information, call (518) 582-2601.

This trip log begins at the Warrensburg exit (23) of the Northway, Route 87. From Lake George Village, travel north on Rt. 87 to reach Exit 23.

	Mileage	Cumulative mileage
Travel north on Rt. 87 to Warrensburg, Exit 23	0	0
Go to end of exit ramp, turn left	0.3	0.3
Go to traffic light at NYS Rt. 9 north, turn right	0.1	0.4
Travel through Warrensburg on Rt. 9		
Traffic light, go straight	0.2	0.6
Traffic light at Rt. 418, go straight	0.6	1.2
Traffic light at fork, fork to right on Rt. 9	0.5	1.7
Travel Rt. 9 north to intersection with NYS Rt. 28,		
Turn left on Rt. 28 west	2.9	4.6
Blinker at Rt. 8 intersection, go straight	11	15.6
Intersection with Rt. 28N, go straight	6.6	22.2
Turn left on Barton Mine Road	4.6	26.8
Travel to end of Barton Mine Road	5	31.8

Note: the intersection of Barton Mine Road and Rt. 28 is marked by a small cluster of buildings. Among these is Jasco's mineral shop. On the east side of Rt. 28 facing south there is a sign opposite Barton Mine Road indicating the Barton Mine (Gore Mt.) mineral shop.

STOP 1. BARTON MINES, GORE MT. (1.5-2 hours) Note: there is a charge of $1.00 – 2.00 per pound of material collected, payable at the mineral shop. There is also a charge to visit the mine as a tourist. The cost is usually reduced for school groups. This NYSGA trip will visit the No. 4 pit of the Gore Mt. mine. Visitors are normally escorted into the No. 1 pit, the earliest and lowest, which was developed in the garnet hornblendite. The No. 4 pit offers access to the olivine gabbro protolith of the ore in addition to the garnet ore per se. Proceeding down the access road on the north side of the pit takes the visitor across the contact zone between the protolith and the ore. In a matter of a few meters, the rock changes from altered olivine gabbro to garnet amphibolite. Garnets appear in the rock and increase in size and abundance over a very short distance. The texture and minerals of the olivine gabbro disappear. Description of the ore is presented in the text above. The opposite (south) side of the mine is in a "syenitic" gneiss, separated from the ore by a vertical fault. Minor granitic pegmatite occurs at the contact with the fault.

Travel back to Rt. 28	5	36.8
Turn right at stop sign, go to Rt 28N at North Creek	6.6	43.4
Turn left on Rt 28N, go straight at 4-way stop	0.1	43.4
Turn right at Blue Ridge Road	21.5	64.9
Note: Sign for NL Ind., MacIntyre Development and sign for High Peaks Wilderness Area		
Turn left at Tahawus Road	1.1	66
Turn left at Upper Works Road	6.5	72.5
Proceed to dirt road to Cheney Pond	0.6	73.1
Boulders on right, gate on left		

STOP 2. KRONOS Worldwide, INC., CHENEY POND DEPOSIT (1.5-2 hours)

At this stop, examine the large boulders that block the dirt roads on the east and west side of the main road. Both "gabbroic" (disseminated) and "anorthositic" (massive) ore types are displayed here. The massive ores contain xenoliths of anorthosite and xenocrysts of plagioclase derived from the anorthosite, attesting to the intrusive nature of this ore. Garnet reaction rims occur on both the xenoliths and xenocrysts. Deformed (sheared) and relatively undeformed representatives of the host rocks are present. Minerals are: blue-gray plagioclase (~An_{45}) megacrysts, white, fine grained plagioclase matrix (~An_{28}), magnetite, ilmenite, garnet, amphibole and clinopyroxene. Sulfide mineralization is minor and late stage.

Walk (or drive) up the old haul road about 1km to the prospect pit developed in the Cheney Pond deposit. This is entirely disseminated ore. However, a great deal of "massive" ore of low grade has been hauled to this site to build the road so confusion regarding the exact source of the ore is possible. Sampling is recommended alone the haul road.

South of the top of the pit at Cheney Pond an overgrown logging road leads to the left into the woods. Follow this road about 75 m to a prospect pit, perhaps 7 meters across, on the right. This prospect was developed in nelsonite, although the identity of the apatite was not recognized when the pit was excavated.

Return to cars. Proceed north on Upper Works Road	2.0	75.1

Furnace stack is on the right, by the edge of the road, wheel house is close to the river.

STOP 3. ADIRONDACK IRON AND STEEL CO. "NEW" FURNACE (1 hour)

Please take only photographs at this stop. Do not take artifacts as souvenirs. Visible at the present time are the furnace stack, tuyere arches and tuyeres, casting arch, slag piles, and the entire contents of the wheel house with cross-heads, blowing tubs, and manifold more or less intact and in working position. Of the dam, wooden water wheel and casting house, little remains. Description given in text above.

ORIGIN OF THE LEWIS WOLLASTONITE DEPOSIT

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INTRODUCTION

The Willsboro-Lewis mining district is a world-class wollastonite ($CaSiO_3$) producer in the eastern Adirondack Mountains, where wollastonite skarn is associated with Mesoproterozoic anorthosite. This district has estimated reserves of about 6 million short tons (St) from the Lewis open pit, Fox Knoll (the underground mine), and Oak Hill, half of which are represented by the currently operating Lewis mine. NYCO's Willsboro, NY plant produced 58,000 St of finished product in 2007. Combined with another 60,000 St produced at its Mexican operations, NYCO supplies approximately ¼ of the world's wollastonite. Wollastonite and its association with clinopyroxene and garnet was first described from the mining district by Vanuxem (1822), and his chemical analysis may be the first of a mineral from the Adirondacks (Buddington 1977). Fieldwork by A.F. Buddington in 1936 and 1937 led to the recognition that wollastonite skarns formed between anorthosite plutons and carbonate country rocks (Buddington and Whitcomb 1941). Oxygen isotope studies (Valley and O'Neil 1982; Valley et al. 1990; Clechenko and Valley 2003) reveal the importance of heated meteoric water during anorthosite emplacement and formation of the wollastonite-clinopyroxene-garnet ores.

Mining in the Willsboro-Lewis mining district is ongoing at the Lewis mine, which has been in operation since 1983. Mining in the area was intermittent until the early 1953, when Cabot Corporation began wollastonite mining operations at the Willsboro mine for use as a ceramic base and filler (Olmsted et al. 1992; Whitney and Olmsted 1995). Underground mining at Willsboro began in 1960 and ended in 1982. Currently at Oak Hill, NYCO is in contract with Graymont Materials of Lewis to crush the caprock for aggregate applications, thus lowering the overall strip ratio. Wollastonite mining will commence in approximately six years, with an overlap in supply to the plant in Willsboro from both the Lewis and Oak Hill mines. At this time the Deerhead deposit has been deemed uneconomical for mining. Wollastonite from Lewis and NYCO's operation in Sonora, Mexico is primarily used for applications such as plastics & elastomers, paints & coatings, construction materials, friction and metallurgical products, and finally for conductive/anti-static purposes.

GEOLOGIC OVERVIEW

Wollastonite ores in the Willsboro-Lewis area are associated with the contact zone of the Westport dome of the Marcy anorthosite massif (Fig. 1). The Marcy anorthosite and associated granitic rocks intruded at ca. 1155 Ma (McLelland et al., 2004). This AMCG (Anorthosite-Mangerite-Charnockite-Granite) suite dominates the Adirondack Highlands, and is contemporaneous with the Morin and Lac St. Jean anorthosite suites in Quebec (Doig 1991; Higgins and van Breemen 1996). In the Adirondack Highlands, anorthosite-suite rocks were emplaced into crust made up of 1300–1250 Ma arc-related tonalities and metasediments metamorphosed during the ca. 1210–1160 Ma Shawinigan orogeny (Heumann et al. 2006). Anorthosite-suite plutons and country rocks are overprinted by ca. 1080-1050 Ma granulite-facies mineral assemblages of the Ottawan orogeny (McLelland et al., 2001).

The Willsboro, Lewis, Oak Hill, and Deerhead wollastonite deposits are located in a ca. 25 km long, 1.5 km thick belt of metasedimentary and metaigneous rocks that border the Westport anorthosite dome on the north and west (Fig 1b; Buddington and Whitcomb 1941; Whitney and Olmsted 1993). Anorthositic rocks of the Westport dome are primarily made up of gray intermediate plagioclase megacrysts surrounded by finer-grained (often recrystallized) white or pale gray plagioclase, pyroxenes, Fe-Ti oxides, hornblende, garnet, and sulfides. Deformation of anorthositic rocks is shown in apparent igneous flow-alignment of plagioclase megacrysts and foliation (with or without gneissic banding) that could be related to emplacement and subsequent metamorphism.

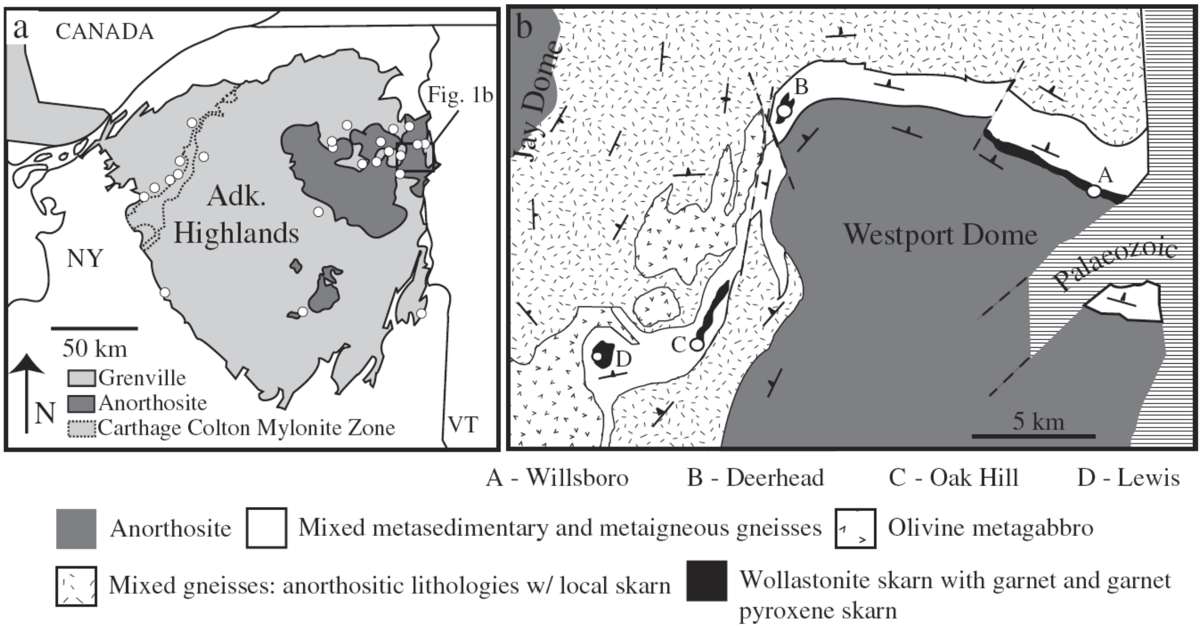


Figure 1. (a) Regional map showing Grenville-aged exposure of the Adirondack Highlands. Open circles are wollastonite occurrences from Valley et al. (1990). (b) Geology of the Willsboro-Lewis mining district, after Whitney and Olmsted (1998). Maps from Clechenko and Valley (2003).

Metamorphic clinopyroxene, clinopyroxene, and garnet, often showing coronitic textures with igneous minerals, formed during the Ottawan orogeny at ca. 850-800°C to 650°C and 6.5 to 8 kbar (Bohlen et al., 1985; Spear and Markussen, 1997). Compositions of relict igneous pyroxenes suggest emplacement temperatures of ≥1000–1200°C (Bohlen and Essene 1978). A depth of emplacement for the anorthosite massif of <10 Km is constrained by the infiltration of meteoric water into the contact aureole during skarn formation (Valley and O'Neil 1982), consistent with the aluminum contents of igneous clinopyroxenes (Spear and Markussen, 1997).

GEOLOGY AND GEOCHEMISTRY OF WOLLASTONITE ORE DEPOSITS

The ore zone in the Willsboro-Lewis district is a mixed zone made up of mafic (anorthositic to gabbroic) gneiss, granitic gneiss, amphibolite, quartzite, pelite, calc-silicate, marble, and skarn (Whitney and Olmsted 1995; Olmsted et al. 1992). Skarn lithologies include garnetite, garnet-pyroxene rocks, and wollastonite-garnet-clinopyroxene ores. Individual lenses of foliated metaigneous and metasedimentary rock in the ore zone range from meters to over a hundred meters in thickness. Some metaigneous bodies are observed to intrude metasediments and thicker gabbroic layers display relict igneous textures (Whitney and Olmsted 1995). Both 1070-1050 Ma Ottawan deformation and deformation predating ca. 1155 Ma skarn formation is evident in the ore zone. Large euhedral skarn garnets formed during anorthosite emplacement are locally preserved in zones of low strain, but most garnet was recrystallized during Ottawan deformation (Clechenko and Valley 2003). Skarn-converted dikes that cut pre-existing foliation are also preserved (Whitney and Olmsted 1998), and demonstrate that some of the observed fabrics are older than ca. 1155 Ma.

Wollastonite ore bodies are discontinuous, range up to 30 meters thick, and are composed of wollastonite + garnet + clinopyroxene. The general lack of excess reactants (calcite or quartz) in these rocks has led most workers to attribute wollastonite formation to voluminous circulation of fluids in the contact zone between the Westport anorthosite dome and carbonate country rocks (e.g. Valley and O'Neil 1982). Trace element and mineralogical composition are consistent with that majority of ore being formed from metasomatized carbonate rocks, with some rocks forming from altered anorthosite-suite rocks (i.e. endoskarn). Skarns with igneous protoliths can contain up to several percent apatite and titanite with accessory scapolite, plagioclase, clinozoisite, idocrase, and zircon (Whitney and Olmsted 1998).

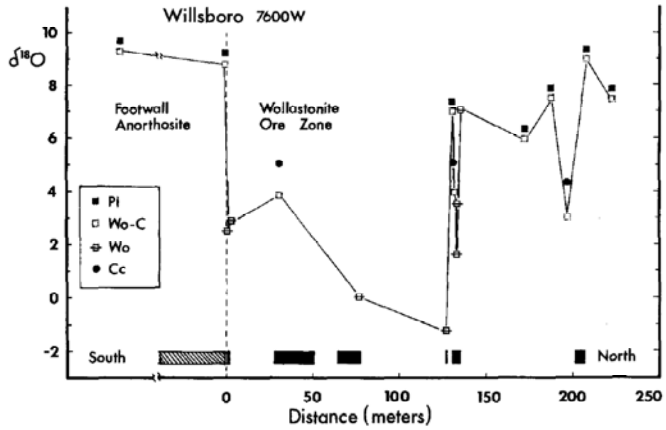
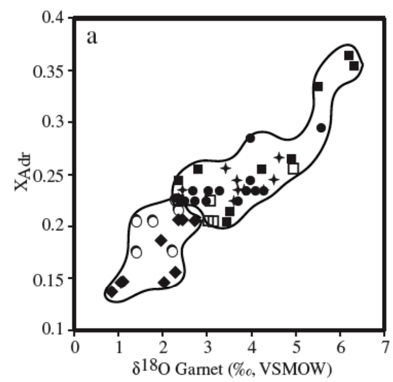


Figure 2. Oxygen isotope ratios of minerals from a traverse of Willsboro ore zone in drillcore. Low $\delta^{18}O$ values in skarn ores are caused by interaction with heated meteoric water during anorthosite emplacement. Sharp gradients in $\delta^{18}O$ between different lithologies are preserved through Ottawan granulite facies metamorphism, showing limited post-intrusion modification by fluids. Figure from Valley et al. (1990).

Oxygen isotope ratios of minerals in wollastonite ores demonstrate the importance of surface waters in skarn formation (Fig. 2). Ore rocks from the Willsboro, Lewis, Oak Hill, and Deerhead deposits have low $\delta^{18}O$ values (as low as -1.3‰ SMOW), which is indicative of interaction with heated meteoric water (Valley and O'Neil 1982; Clechenko 2001; Clechenko and Valley 2003). Major and trace elements and oxygen isotope chemistry in ore-zone garnets shows early formation of contact metamorphic garnet (HREE enriched, high $\delta^{18}O$, low Fe, higher Mn, Ti, Mg, Zr, negative Eu anomaly) and later formation of hydrothermal garnet (LREE enriched, low $\delta^{18}O$, high Fe, low Ti, Mn, Mg, and Zr, positive Eu anomaly) in meteoric fluids with some igneous component (Whitney and Olmsted 1998; Clechenko 2001). Interestingly, at the Willsboro mine, relict skarn garnets in garnetite preserve zoning that reveals details of a different hydrothermal circulation system or event (Fig. 3). Zoning in garnetite ranges from grossular-rich garnet with low $\delta^{18}O$ to andradite-rich garnet with higher $\delta^{18}O$ values. This is interpret to be the result of low $\delta^{18}O$, Mg-rich meteoric fluids and high $\delta^{18}O$, Fe-, Mn-, and Ti-rich magmatic fluids both being involved in garnetite genesis (Clechenko and Valley 2003).

Figure 3. Oxygen isotope ratios and mole fraction andradite in garnets from Willsboro mine garnetite. Low $\delta^{18}O$ and low iron contents are attributed to meteoric fluids which mixed in the skarn zone with high $\delta^{18}O$, higher iron igneous fluids. Different symbols are different individual garnets. Figure from Clechenko and Valley (2003).



The presence of meteoric water during anorthosite intrusion places important constraints on the depth of emplacement of the Marcy anorthosite. Low $\delta^{18}O$ values caused by hydrothermal alteration by meteoric fluids are not limited to the Westport Dome, for example anorthosites in the southern Marcy massif have $\delta^{18}O$ values as low as 3.0‰ (Morrison and Valley, 1988). Large volumes of surface water during contact metamorphism is strong evidence of shallow (<10 km) anorthosite emplacement, and is an important line of evidence for contact metamorphism not being the heat source for regional granulite-facies metamorphism (Valley and O'Neil 1982). In contrast, skarns associated with cotemporaneous Morin anorthosite suite do not show isotopic evidence for meteoric water (Peck 1996; Peck et al. 2005), nor do wollastonite-bearing rocks elsewhere in the Adirondacks (Valley et al. 1990), including the ca. 1155 Ma Valentine wollastonite deposit near Harrisville in the NW Adirondacks (Gerdes and Valley 1994) or the ca. 1155 Ma Canton Saint-Onge wollastonite deposit from the Lac St. Jean anorthosite (Higgins et al. 2001). This may indicate deeper emplacement in the crust or differences in hydrothermal flow conditions for these intrusive suites.

FIELD TRIP TO THE LEWIS OPEN PIT

The geology of the Lewis deposit is described by Whitney and Olmstead (1993), and Figure 4 shows the current open pit workings. The footwall to the ore zone is gabbroic anorthosite gneiss. The ore zone, which is on average 8-10 meters thick, is gneissic with alternating wollastonite and garnet+pyroxene layers. Garnetite is locally abundant near the contacts with plagioclase-bearing rocks, such as anorthosite. The ore zone and adjacent units strike E-W and dip 10-15° S or SW (Whitney and Olmsted 1993). Historically, the Lewis deposit was one continuous ore section that was up to 21 meters thick in some areas of the mine. As mining proceeded southward, the ore fingers out with layers of "interburden" amid the ore. The interburden zones can range from 0.3 meters to 9.0 meters thick which create a challenge to mining operations.

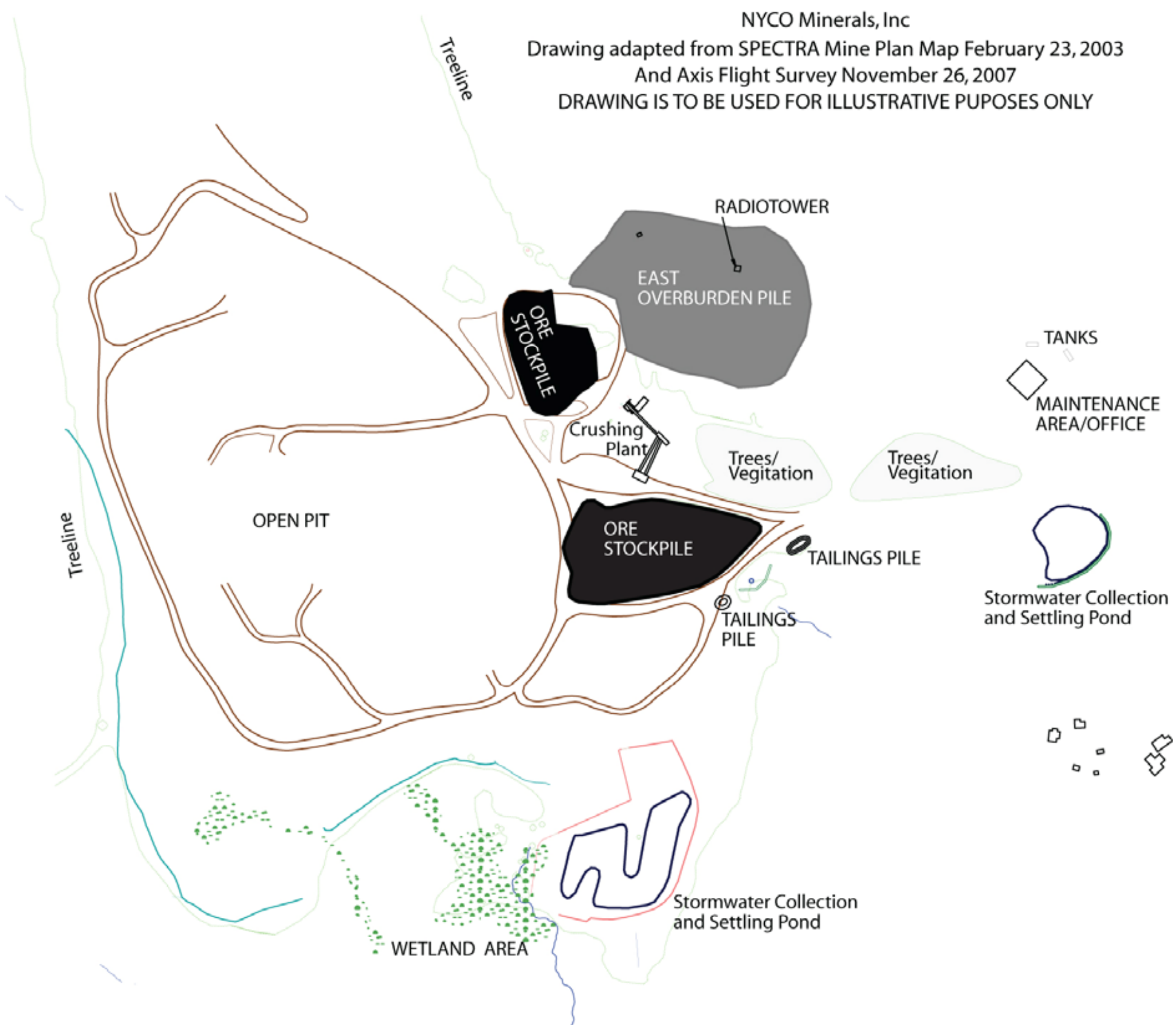


Figure 4. Simplified drawing of the Lewis open pit and area.

ACKNOWLEDGEMENTS

This summary of the geology of wollastonite deposits around the Westport Dome draws heavily on the work of Phil Whitney, Jim Olmsted, John Valley, and Cory Clechenko. The authors would like to thank John Valley and Cory Clechenko for stimulating and fun field trips to Adirondack wollastonite deposits and discussions about their geology. John Valley is also thanked for his comments on this manuscript and Bruce Selleck is thanked for his patient editorial guidance.

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MINING HISTORY, MINERALOGY AND ORIGIN OF THE GNEISS (GRANITE)-HOSTED Fe-P-REE AND Fe OXIDE AND GABBRO-HOSTED Ti-Fe OXIDE DEPOSITS FROM THE MINEVILLE-PORT HENRY REGION, ESSEX COUNTY, NY

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"*…genesis of these important ore bodies in Northern New York may reach the conclusion that there is a hopeless disagreement among those who have studied these deposits" (Alling, 1925)*

INTRODUCTION

New York State was a significant supplier of iron ore used in developing the resources and industries of the United States through the first part of the 20th century. The most important regions with iron deposits in the Adirondack Mountains are the northwestern area extending from Jefferson County into St. Lawrence County that includes the Antwerp-Keene hematite belt (Sterling and Caledonia mines) and the Jayville and Clifton mines; the central region with the Benson mines (St. Lawrence County) and Tahawus (Essex County); and the northeastern region with the Mineville-Port Henry-Fisher Hill group of mines (Essex County), and Ausable and Lyon Mountain mines (Clinton County). The earliest discoveries and mining operations of iron ores in the Adirondack Mountains were in the northeastern part (Linney 1943).

Iron ore was mined on the northeastern side of the Adirondack Mountains between Lake Champlain and the Adirondacks as early as 1804 (Birkinbine 1890). At the end of the 19th century three main areas were mined for iron in this region: Mineville-Port Henry-Fisher Hill, Chateaugay Mine at Lyon Mountains and Crown Point. The Mineville-Port Henry-Fisher Hill mines were considered "to represent one of the greatest magnetite deposits in the world" at that time (Anderson and Jones 1945). The first record of mining in the Mineville-Port Henry was in 1775 at a location that later came to be called Cheever Mine (Farrell 1996), but the first lease reported for this mine has been given later in 1820 (Smock 1889). The Mineville-Port Henry Iron Mining District had a cyclical pattern of boom and bust in its long mining history that spanned over 150 years. The various mines were repeatedly opened and then shutdown only to be reopened later depending on world events, economic changes and ownership's interests.

The Mineville-Port Henry mines were important and continuous iron producers, but they were also known for their apatite by-product and for the beauty and specific forms of the magnetite crystals that can still be seen today in many museums.

In this field trip we will examine the mineral composition and texture of the iron ore, the characters of the country and host rocks, and the relationships between magnetite ore and the host rock from the Craig Harbor, Cheever, Pelfshire, Mineville (the waste dumps of the Clonan and Bonanza pits), and the Barton Hill mines (Essex County).

MINING HISTORY

Although the presence of iron ore was known since 1775, the first significant operating mine in the Mineville district was the Cheever Mine, located north of Port Henry, which was opened in 1820 by Charles Fisher. Soon thereafter, the same owner started the works for the Fisher Hill Mine that was worked intermittently until 1893;

later exploration in the 1920's proved that the Fisher Hill ore body would yield at least 40,000,000 long tons of ore and in 1941 it began operations as a Defense Plant Corporation project of the Republic Steel Corporation (Anderson and Jones 1945).

The Mineville district produced two distinct varieties of iron ore (Birkinbine 1890): magnetite-apatite and magnetite-silicate ore. The magnetite-apatite ore was produced from what was called the "Old Bed" and the magnetite-silicate ore came from the "New Bed". These names were applied to the ore mined from different mining works. The name "Old Bed" was derived from the initial opening made in 1824 at the eastern end of the Sanford Pit. In 1829 the "Old Bed" was mined by the Ore Bed Company at the Sanford Pit and the 21 Mine. It was thought to represent a series of lenses "lying en echelon, or nearly parallel, of varying thickness and dip" (Birkinbine 1890). New mine works were made northwest of the "Old Bed" and produced magnetite-silicate ore. These works opened the "New Bed" along a large area. The "New Bed" was apparently a continuous vein, but was separated in different sub-layers and complicated by "horses" (faults). Hall in his "Laurentian Magnetic Iron Ore Deposits of Northern New York" considered the "Old Bed" and "New Bed" to be located at different levels in the metamorphic succession (Birkinbine 1890).

Prior to1840, the Barton Hill Mine began as a series of openings on a magnetite body that seemed to be the northern continuation of the New Bed. The extensive development of iron mining in the area took place after 1849 when the properties came into possession of the Witherbee and the Sherman interests that were later incorporated into Witherbee, Sherman & Co (Farrell 1996). The Barton Hill area, especially the "Lover's Hole" opening is well known for the well crystallized magnetite crystals exhibiting pseudo-cleavage (octahedral parting) and interesting and rare crystallographic forms.

Not only was the magnetite of interest at that time, but also the apatite ("red sand") which accompanied magnetite in the "Old Bed". Apatite was first mined in 1852 by the Moriah Phosphate Company with the intention of producing fertilizers. The mine initially exploited the outcrop, and the amount of apatite from the surface was greater than it was from the underground works (Maynard 1874). The phosphate came to the attention of the State Geologist, Professor Ebenezer Emmons who supervised the American Mineral Company formed in 1853. The company mined mainly apatite for fertilizers and iron as by-product, but because the market did not react very well to their products, the company had to lease their properties and mineral rights to Port Henry Iron Ore Company.

Between 1858 and 1900, due to the pressure of economic factors, some of the mines were sold to different groups of investors or the "old" companies changed their names. The decline of the Mineville mining district started as early as 1875, when only a few mines were operating. The main economic factor that led to the decline of the Mineville deposits was the competition from the large and newly found iron ranges in Minnesota and Michigan.

The year 1900 brought new mining development in Mineville and it became the largest operating iron district east of Mississippi River (Farrell 1996). The only competitors were the Chateaugay Mine at Lyon Mountain and the small mines at Crown Point and Ausable, all producing the same type of ore as Mineville, but not in such high quantity or quality, and the Lake Sanford deposit that was Ti-rich, a property not relished by metallurgists.

After 1920 mining economically collapsed again in the Mineville region; a few mines operated until 1932 – 1933 when they were completely shut down. After 1935 there was an attempt to revitalize the mining, with a small spurt of production in 1936 – 1937 plus the creation of the Republic Steel Corporation in 1938 which leased all the Witherbee Sherman and Company's properties. Production and mining interest in the area again started to decline; the mines were repeatedly shut down and reopened for short periods until they permanently closed in 1971.

A new era for apatite began in 1940. Initially apatite was considered only useful for the production of fertilizers, but a U. S. Geological Survey report showed that the phosphate was very rich in rare earth elements. Immediately, Molycorp, a REE producing company leased the mineral rights and started their recovery from the tailings, but further feasibility studies were unfavorable and they did not acquire the property. Interest in the REE-bearing apatite was renewed in 1983 and Williams Strategic Metals of Colorado purchased the apatite-rich tailings then re- sold them in 1986 to Rhone-Poulenc, Inc. a French state-owned company (Farrell 1996). Now, Rhodia Inc. of New Jersey has the surface and mineral ownership.

The history of iron mining at Mineville is a reflection of the most important historical events in the world and in the United States. Figure 1 displays a very irregular pattern for the iron production at Mineville through the 20th century until the mines closed in 1971. These many ups and downs can be easily correlated with periods of economic depression, wars or the discovery of more competitive deposits elsewhere. Today, Mineville remains an important mineral locality and an iron deposit that still requires more research to be completely understood and explained.

The Craig Harbor mine situated above the Lake Champlain shore, in Port Henry, was not an important mining operation. It was mentioned by Emmons (1842) as a small mining development, but was not cited by subsequent researchers. The deposit will be visited during this trip to examine the gabbro, the pyroxene hornblendite, and the ilmenite-magnetite ore.

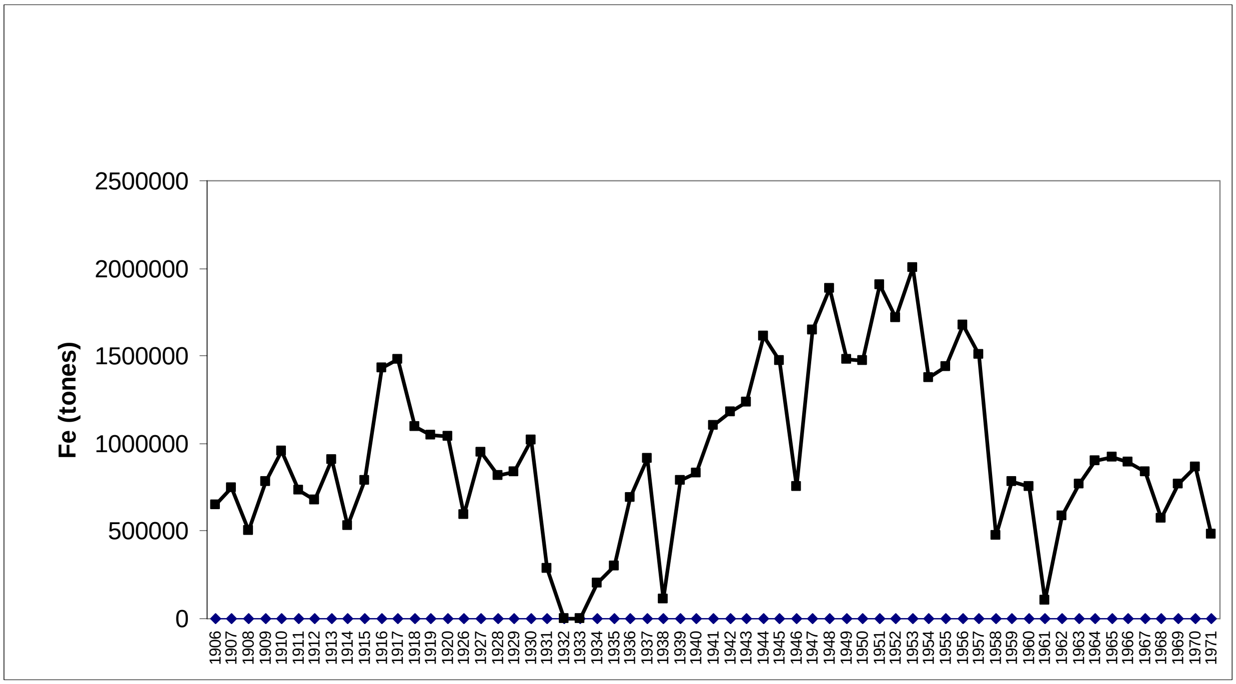


Figure 1. Iron production at Mineville in the 20th century until 1971 when the mines closed (data from Farrell 1996).

GEOLOGY OF THE IRON DEPOSITS IN THE MINEVILLE – PORT HENRY REGION

The iron deposits in the Mineville – Port Henry area, (Essex County), NY, belong to a large metallogenetic belt that developed in the Grenville rocks in New York and New Jersey.

The Mineville deposit is a low – Ti Fe (oxide)–P–REE ore, that is, in part, similar to Kiruna-type iron deposits in Sweden based on its mineralogical and chemical composition. The host rocks have both mafic and felsic compositions, display igneous features, and probably are correlative with the informally named Lyon Mountain granitic gneiss with 1070 to 1050 Ma radiometric ages (McLelland et al. 1988). A metasedimentary series containing Proterozoic marbles, calc-silicates and gneisses structurally overlies the igneous sequence. Kemp (1908) described the host rock for the Mineville iron deposit as "augite syenites and related types" with granite and diorite composition representing the "related types." Buddington (1939) considered a hybrid composition between granite and metasedimentary or metagabbroic rocks, and Alling (1925) suggested an igneous and sedimentary origin for the protolith metamorphosed in the Grenville orogenic cycle. McKeown and Klemic (1956), based on the mapping of the Republic Steel geologists, described a metamorphic sequence starting with a basal metagabbro (Kemp's mafic syenite) followed by the magnetite ore from the "Old Bed" ore and granite gneiss passing into a diorite then to gabbroic composition and magnetite ore from the Harmony Bed. Whitney and Olmsted (1988) proposed an anorogenic bimodal, dominantly leucocratic volcanic suite with minor sedimentary sequences metamorphosed to granulite-facies, origin for the Lyon Mountain gneiss. An igneous genesis was inferred by Foose and McLelland (1995) and is based on field and geochemical evidences. The gneiss (granite) displays felsic Na - and K -rich compositions that alternate with more basic, amphibole-, pyroxene-, and phlogopite-bearing rocks, both characterized by high amounts of REE. Pegmatite bodies with simple compositions (mostly quartz, feldspar ± magnetite ± allanite-(Ce) and minor scapolite, titanite, epidote and zircon) crosscut the magnetite ore. Both the host rock and the magnetite ore commonly exhibit similar fabrics and layering.

The petrographic study of polished thin sections made from samples of the ore, and from hanging wall and footwall rocks shows a range of compositions from gabbro to monzodiorite to tonalite (Fig. 2).

The iron ore is composed of magnetite (partially replaced by hematite), apatite and clinopyroxene. Allanite and titanite commonly rim the magnetite and zircon (in places millimeter size crystals) is associated with the iron

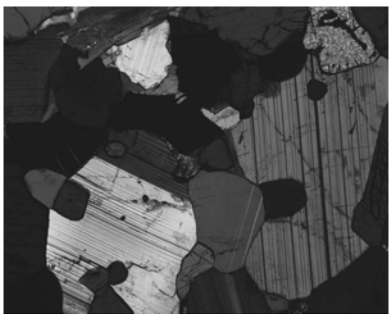
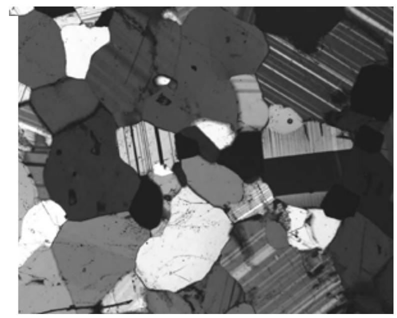


Figure 2. Left: microscopic image (transmitted light) of the gabbro in thin section. The rock contains plagioclase, clinopyroxene (dark), titanite (upper right corner). Right: tonalite (quartz as light grains and twinned oligoclase). The rocks are not deformed. Field of view: 4.5 mm.



oxide in all the rock types. Rarely, magnetite displays spinel exsolution along {111}. Ilmenite is present in the more mafic compositions (gabbro) in rounded grains, has exsolutions of titanian-hematite in variable patterns, and in places, has a narrow or wide rim of titanite (Fig. 3).

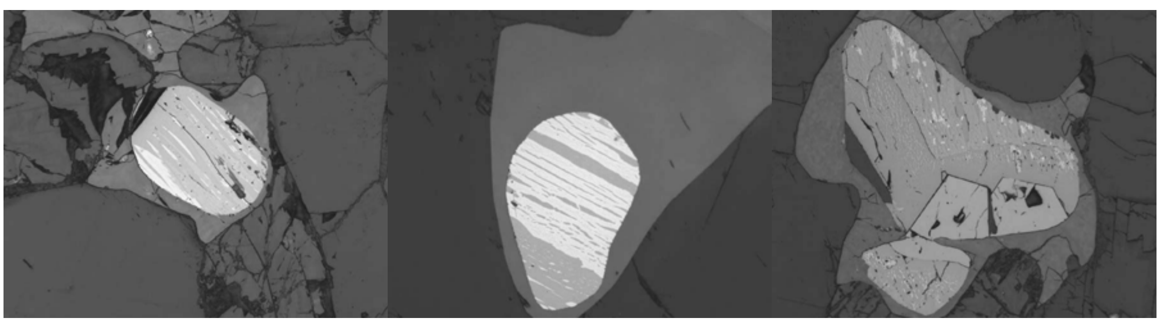


Figure 3. Rounded grains of ilmenite (reflected light) with exsolution of titanian-hematite rimmed by titanite (light gray). The other minerals (dark gray) are clinopyroxene and plagioclase. Field of view: 3 mm.

The host rock does not show any deformation at microscopic scale even though it displays folds in outcrop (Barton Hill); the oligoclase is twinned after Albite law and quartz displays only incipient undulatory extinction. K-metasomatism and formation of microcline (Fig. 4) is a late process, and proceeds initially along the contacts of magnetite and/or oligoclase; the Na-metasomatic alteration is latter than the K-metasomatism.

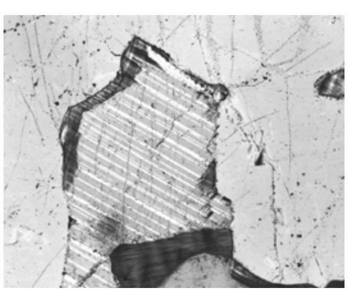


Figure 4. K-metasomatism and formation of microcline in the host rock: microcline develops on the contact between quartz (light gray) and oligoclase (with parallel twins). Field of view: 4.5 mm

Sprays of ferro-actinolite developed between magnetite and quartz. Late, low temperature veins of chlorite crosscut the magnetite ore.

Some of the general geological features of the Precambrian iron deposits mentioned by Foose and McLelland (1995) can be recognized in the Mineville iron deposit:

a. Occurs as lenses and/or layers in leucocratic rocks containing albite-oligoclase + quartz + K-feldspar composition above more mafic layers;
b. The iron ore displays similar fabrics to the host rock;
c. The crosscutting pegmatite bodies are not or are only slightly deformed;
d. Both the host rock and the ore have high amounts of REE and the gneiss is strongly enriched in Na, K, P, B, etc.

MINERALS FROM THE MINEVILLE IRON DEPOSIT

The main minerals that were identified at Mineville are magnetite, hematite (martite), fluorapatite, stillwellite-(Ce), allanite-(Ce), monazite-(Ce), edenite, actinolite, ferro-actinolite, scapolite, titanite, and zircon in the main ore and, in part, in the pegmatite bodies dolomite, smoky quartz, calcite in late veins and ilmenite and titanian hematite as tiny disseminations in the host rock. Minor tiny (µm size) phases of secondary thorite, allanite-(Ce), parisite, and monazite-(Ce) in some apatite crystals and kainosite-(Y) in edenite were recognized under the polarizing microscope and by SEM – EDAX in thin/polished sections. Bastnaesite-(Ce) (McKeown and Klemic 1956) was previously reported but not identified by the authors, and probable lanthanite-(Ce), mentioned by Blake (1858), was detected based on the electron microprobe data.

Magnetite is by far the most spectacular mineral at Mineville, though some other minerals also show interesting compositional and crystallographic features. It probably first came to mineralogists' attention at the end of the 19th century when some remarkably unique specimens from the Lover's Hole pit were found around 1887 – 1888 (Jensen 1978). The main crystallographic form was the octahedron, but some had interesting modifying forms that were well documented at that time by Professor Koenig from the University of Pennsylvania (Birkinbine 1890). Besides the octahedron (Fig. 5), he described combinations of octahedron with rhombic dodecahedron, pentagonal dodecahedron, cube or icositetrahedron. Perhaps, the most intriguing forms are the pseudocleavages (parting planes), where the parted fragments develop a pseudo-prismatic appearance (Fig. 5) with stilpnomelane coating the parting planes and crystal faces. One of the most spectacular specimens has a "…perfect octahedron with faces of over one inch, resting loosely in the socket…" (Birkinbine 1890). According to Farrell (1996) this outstanding specimen was called "Big Diamond" and is now in the collection of the A. E. Seaman Mineral Museum, at Michigan Technological University, Houghton, Michigan. The magnetite crystals that display a combination of octahedron and rhombic dodecahedrons have striations parallel to the octahedron faces (Kemp1890). Cathrein (1887) and Mügge (1889) both in Kemp (1890) previously noted and considered this type of striation as polysynthetic twinning of the spinel law. Kemp (1890) interpreted them as pressure-generated pseudocleavage planes in the mineral.

Figure 5. Magnetite crystals from the Lover's Hole, Barton Hill mines, Mineville. Left: magnetite octahedrons ((3.5 x 2.5 x 2.5 cm, NYSM 12495). Right: unusual morphology of magnetite crystals (4.7 x 1 x 1.5 cm) resulting from parting along octahedral faces (NYSM 12258). On top of the crystal on the right there is stilpnomelane.

Fluorapatite, the "red sand", was found in the ore from the "Old Bed" and is characterized by high concentrations of rare earth elements. It appears as red, brown or yellow small, 1 to 3mm, hexagonal prisms embedded in magnetite. The red and/or brown color is due to infiltrations or inclusions of hematite along the fractures or within the crystal. McKeown and Klemic (1956) described a very narrow rim of 0.05 mm of a reddish-brown aggregate of monazite, bastnaesite and hematite rimming some of the apatite crystals. We examined apatite grains in thin/polished sections under polarizing and scanning electron microscopes and found that most of the fluorapatite grains are fractured, "flushed" by late fluids, and contain tiny grains of magnetite/hematite, monazite-(Ce), allanite-(Ce) and thorite. Parts of the "Old Bed" ore are composed only of magnetite and fluorapatite in variable proportions. The textural features of the Mineville apatite are in Figures 6, 7, 8, and 9. Fluorapatite is a main component of the ore also at the Cheever Mine, but not in the same amount as in the Old Bed from Mineville.

Allanite-(Ce) is found in the pegmatite bodies that crosscut the magnetite ore and/or the host gneiss and in the pyroxene-rich rocks. It was first mentioned by Blake (1858) occurring "abundantly along the plane of contact of the bed of ore with a mass of granitic rocks…maybe an intrusion or segregation in its midst." The crystals described by Blake were very large, 20 – 25 cm long, 6 to 20 cm wide and 2.5 to 5 cm thick suggesting a pegmatitic origin. The crystals displayed "smoothed and perfect" surfaces and conchoidal fractures. He considered Mineville at that time "the best locality for obtaining cabinet specimens of allanite yet known in the United States."

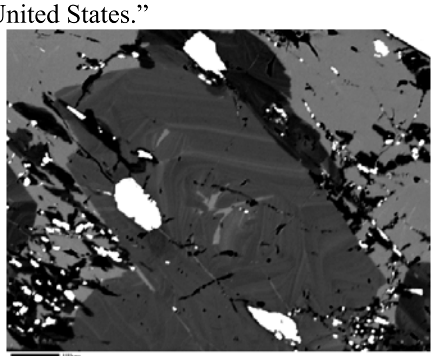


Figure 6. Zoned apatite. The bright grains are magnetite. BSE image. Scale bar: 300 microns.

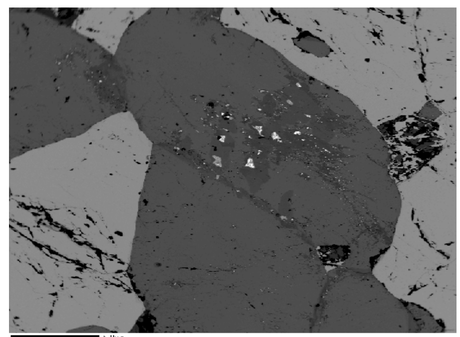


Figure 7. Factured apatite with fluid infiltrations and secondary thorite, allanite and monazite as bright spots. BSE image. Scale bar: 300 microns.

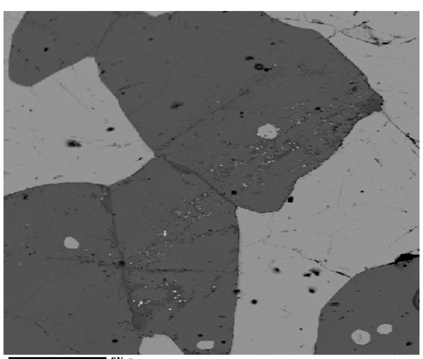


Figure 8. "Trails" of secondary thorite, allanite and monazite (all as light gray) along the fractures in apatite. BSE image. Scale bar: 300

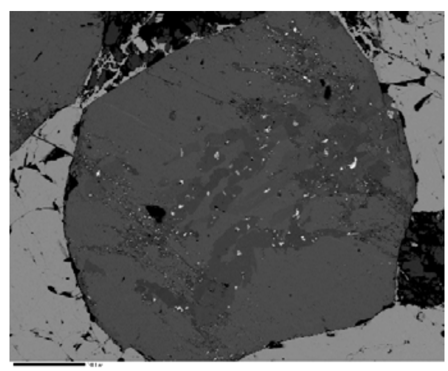


Figure 9. Fractures and patches of apatite leached by fluids and secondary thorite, allanite and monazite (bright spots). BSE image. Scale bar: 300 microns.

Allanite-(Ce) from Mineville also captured the attention of Dana (1884) and Ries (1894), who described it in short mineralogical notes. In this study allanite crystals associated with quartz and monazite-(Ce) from Mineville, were analyzed by polarized light microscopy and electron microprobe. The large, smooth-faced crystals are metamict and the chemical data are in the compositional field of allanite-(Ce). The microscopic study showed that the metamict allanite-(Ce) in the pyroxene-rich rocks is decomposed to monazite-(Ce) and has a fresh rim of Y-dominant allanite that is strongly pleochroic, in dark-brown shades.

Monazite-(Ce). There are two generations of monazite-(Ce) based on their relationships with other minerals. In some pegmatite bodies, 1-3 mm to almost 1 cm crystals of monazite-(Ce) appear associated with or as inclusions in allanite-(Ce) both embedded in quartz. The breakdown of allanite-(Ce) generated secondary monazite-(Ce) as minute grains.

Stillwellite-(Ce). Mineville is the unique occurrence of this mineral in New York. Identified for the first time by Mei et al. (1979) in a sample from the "Old Bed" collected near a fault at the 2100-feet level, and confirmed by X-ray diffraction, stillwellite-(Ce) appears as 1 to 2 mm wide tabular crystals with waxy luster and pink to reddish color in association with fluorapatite and magnetite. Under the electron microprobe, the authors detected some very tiny inclusions in stillewellite-(Ce) that contained 27.05 wt. % Ce_2O_3, 18.02 wt. % La_2O_3, 6.44 wt. % Nd_2O_3, 1.71 wt. % Pr_2O_3 and 1.50 wt. % Sm_2O_3. These small grains are probably lanthanite-(Ce).

Zircon is a common mineral from the Mineville mines. The mineral occurs in three environments characterized by different mineral associations. The first is the zircon appearing as accessory mineral in the host rock and it does not form any crystals of aesthetic interest. Then there are the beautiful, up to 1cm or more, dark colored, well developed, and terminated zircon crystals in the quartz – feldspar pegmatite segregations. These display tetragonal prisms with tetragonal pyramids, and are mostly metamict. In places, the edges between the prism and pyramid are slightly rounded. One specimen from the Barton Hill Mine has small, up to 1 mm pink and transparent zircon crystals within magnetite and/or scapolite. When zircon is included in magnetite, the whole aggregate is surrounded by titanite.

Dolomite is not a very common mineral in the Mineville ore. It forms "saddled", 3 to 5 cm crystals covered by iron hydroxides and is partially replaced by calcite.

Quartz occurs as short, terminated "smoky" prisms on top of magnetite and associated with calcite.

ORIGIN

For many years, geologists debated the origin of the iron deposits in the Grenville province of New York and New Jersey. The genesis of the Mineville iron deposit was related to the contact replacement due to highly heated igneous solutions (Kemp 1897), igneous emplacement (Kemp 1908), basic segregations due to differentiation (Kemp and Ruedemann, 1910), replacement of the country rock by igneous-derived iron-rich solutions (Alling, 1925) or possible metamorphosed sedimentary sequence (Nason 1922). Buddington (1966), Baker and Buddington (1970) and Foose and McLelland (1995) leaned toward a hydrothermal or hydrothermal-metasomatic origin.

The nature of the host rock was also in dispute for a long time. The iron deposits at Mineville are located in a metamorphic unit composed from meta-igneous and metasedimentary sequences that are crosscut by pegmatite dikes, with sharp or diffusive contacts, and with simple or complex composition. McKeown and Klemic's (1956) description (see above in the general geology section), based on their own field observations and information from the mining geologists working for the Republic Steel Company, is very important because it is the first and most substantial work emphasizing the real relationships between the magnetite ore and the gabbro as the host rock. Today, the gabbro with magnetite and the magnetite-bearing pegmatite facies can be seen at the former Barton Hill Mine. At the mines that exploited the Cheever ore body, the contact between the magnetite ore with the felsic and more mafic compositions can be easily examined. The granitic facies (Kemp's gneiss) seems to be the hanging wall and not the host for the ore in the Old Bed (McKeown and Klemic, 1956).

Our own field observations confirm that the magnetite and/or magnetite-apatite ore-bearing rock is mostly an un-deformed rock with igneous characteristics, ranging in composition from gabbro to tonalite, cross cut in some places by magnetite-bearing pegmatite bodies with simple mineral compositions. The magnetite-apatite and magnetite ores occur as layers and lenses in the rocks having mafic and/or felsic mineral compositions. In outcrop, the magnetite-apatite ore seems to occur as a dike cutting the granitic host. The mineralogical composition of the ore seems to be simple, but the field, microscopic and geochemical data show a complex process (or processes) that shaped the iron deposits at Mineville. The main ore mineral, magnetite, contains MgO (0.01 to 0.28 wt. %), Al_2O_3 (0.04 to 3.18 wt. %), and TiO_2 (0.08 to 2.43 wt. %). It is rarely associated with ilmenite in the massive ore, but ilmenite containing titanian hematite exsolution is a constant presence as disseminations in the low grade ore of the hanging wall and footwall. Fluorapatite is enriched in REE (La_2O_3 1.29 wt. %, Ce_2O_3 2.72 wt. %, Nd_2O_3 1.28 wt. %, and Y_2O_3 1.50 wt. %) and Th (ThO_2 0.21 wt. %). The morphology of the ore bodies, their mineral composition, the presence of the ilmenite grains with titanian-hematite exsolutions, the nature of the host rock and the relationship between the ore layers or lenses and the country rocks suggest a probable igneous origin. One possibility is that the ore formed through two liquids (oxide and silicate) immiscibility. This origin is in accordance with the results of the early experimental works that showed the possibility of forming two immiscible liquids, one being a mixture of magnetite and apatite and the other a rock with dioritic composition (Philpotts 1967). Further experiments showed that for a broad range of rock compositions formed under conditions of high fO_2, an immiscible FeO liquid can segregate within a magma having a felsic composition (Naslund 1976). The presence of P, Ti and Fe enhances the immiscibility in magmas (Naslund 1983). Another possibility is that the ore represents a fragmented (intruded?) synplutonic mafic dike(s) in felsic (granitic) rocks. Unfortunately, the outcrop showing these relationships is the Rutgers Mine in Ausable Forks (Clinton County) but this location is not included in our trip and won't be examined during this tour. Didier and Barbarin (1991) showed that the mafic inclusions from the granodiorite-tonalite-diorite plutons have different possible origins but many of them derived from fragmented synplutonic mafic dikes and blobs of mafic magmas in felsic rocks.

To constrain the timing of the magnetite ore formation a zircon crystal (up to almost 2 cm in size) from the Old Bed magnetite ore was analyzed for U-Th-Pb isotopes by LA-MC-ICP-MS. It yielded a 1040 ± 9 Ma radiometric age compatible with that of the Lyon Mountain granite (McLelland et al. 1988). Small, pink zircon crystals (2 to 5 mm in size) found in magnetite grains surrounded by titanite in a pegmatite from the Barton Hill mines yielded an age of 1028 ± 20 Ma (Chiarenzelli et al. 2008 – unpublished data).

The host rocks are much enriched in Na and/or K as well as in REE. The apatite also has very high REE concentrations. The chondrite-normalized REE distribution pattern for the apatite (Fig. 10) and the host rock (Fig. 11) are comparable. This trend could be interpreted in at least two ways: *(a)* synchronous introduction of REE in both the apatite and host rock and correlated probably with Na and/or K metasomatism or (b) the REE pattern of the host rock is strongly influenced by the REE pattern of accessory apatite. Apatite has small negative Sr and Eu anomalies and the $(Ce/Yb)_N$, $(Nd/La)_N$ and $(La/Yb)_N$ ratios indicate that the source was not a highly fractionated fluid, but was strongly enriched in both LREE and HREE.

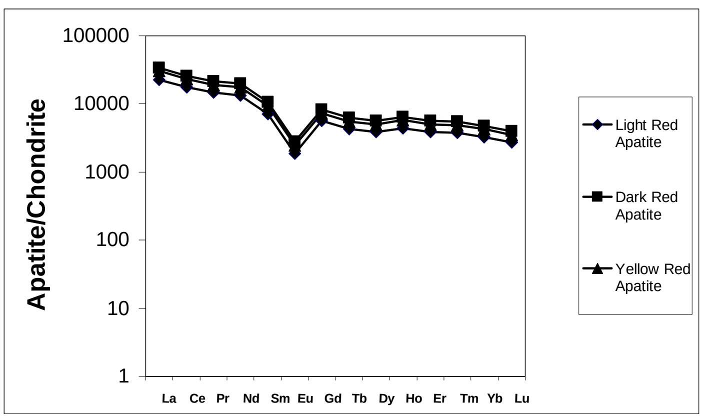


Figure 10. Chondrite-normalized REE pattern for fluorapatite (selected by color) from the Old Bed ore, Mineville iron deposit. Note the high concentration of the REE and the Eu anomaly. The apatite has the same chondrite-normalized REE pattern as the host rock but is enriched in REE by a factor of 10 or more.

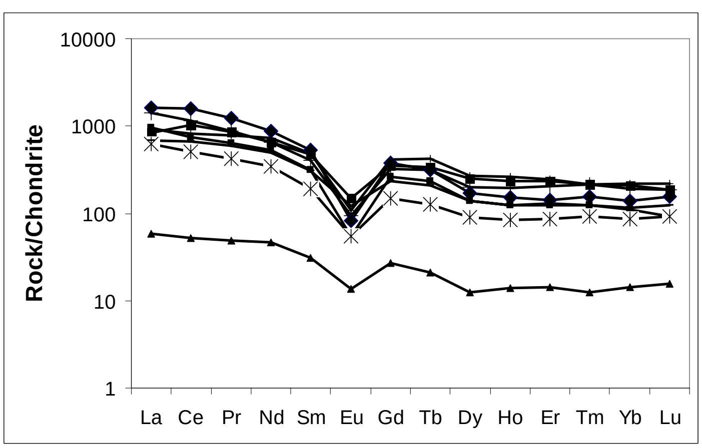


Figure 11. Chondrite-normalized REE pattern for the host rock (samples from Mineville and Barton Hill) of the magnetite and magnetite-apatite ores from the Mineville iron deposit. Note the enrichment in both LREE and HREE and the Eu anomaly.
- triangles: sample from Barton Hill;
- all other samples are from Mineville

The ore contains also small allanite-(Ce) crystals that have been partially replaced in a late event by the assemblage thorite – monazite-(Ce) – kainosite-(Y) – Y-dominant allanite. Results of twenty-two EMPA age determinations on secondary allanite yielded age populations of 240 Ma (primary population) and 165 Ma (minor secondary population), dating the allanite-(Ce) alteration event as Mesozoic. However, the apatite fission-track ages that centered on 113 Ma (Stone et al. 2006- unpublished data) suggest complex and/or multiple hydrothermal events during the Mesozoic or simply represent exhumation during the Mesozoic.

Craig (Crag) Harbor Mine

The mine is located on the Lake Champlain shore at about 15 m above the water. Emmons (1842) mentioned that "the vein is in hornblende … is twelve feet wide and dips southwest".

The host rock is a gabbro and the magnetite ± ilmenite ore is contained by a layer of pyroxene hornblendite. The amphibole (hastingsite) is dominant and associated with clinopyroxenes. Magnetite has spinel exsolution (Figure 12) along {111}. Ilmenite is subordinate, forms grains with sharp contact with or lamellae in magnetite. Apatite, as rounded grains, is also a component of the ore as well as pyrite that appears to be related to a later mineralizing episode. Almandine is a common metamorphic mineral in the gabbro, but not in the ore.

The magnetite-ilmenite ore from the Craig Harbor mine seems to be of igneous origin. It probably formed as result of liquid immiscibility between the iron-titanium oxide melt and the mafic silicate magma, but a late magnetite-ilmenite bearing pyroxene hornblendite intrusion into the gabbro cannot be excluded without further research.

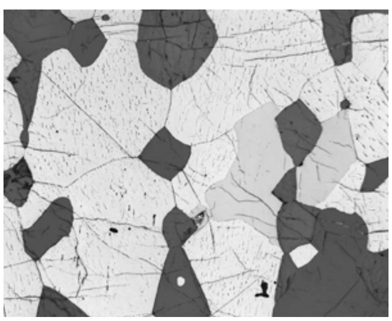
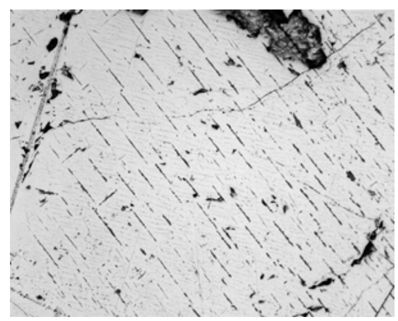


Figure 12. Magnetite-ilmenite ore from the Craig Harbor Mine. Left: magnetite (with spinel exsolutions), ilmenite (light gray) and clinopyroxenes (dark). Right: spinel exsolutions in magnetite along {111). Field of view: 3.5 mm.



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ROAD LOG FOR THE TRIP…(Fig. 13)

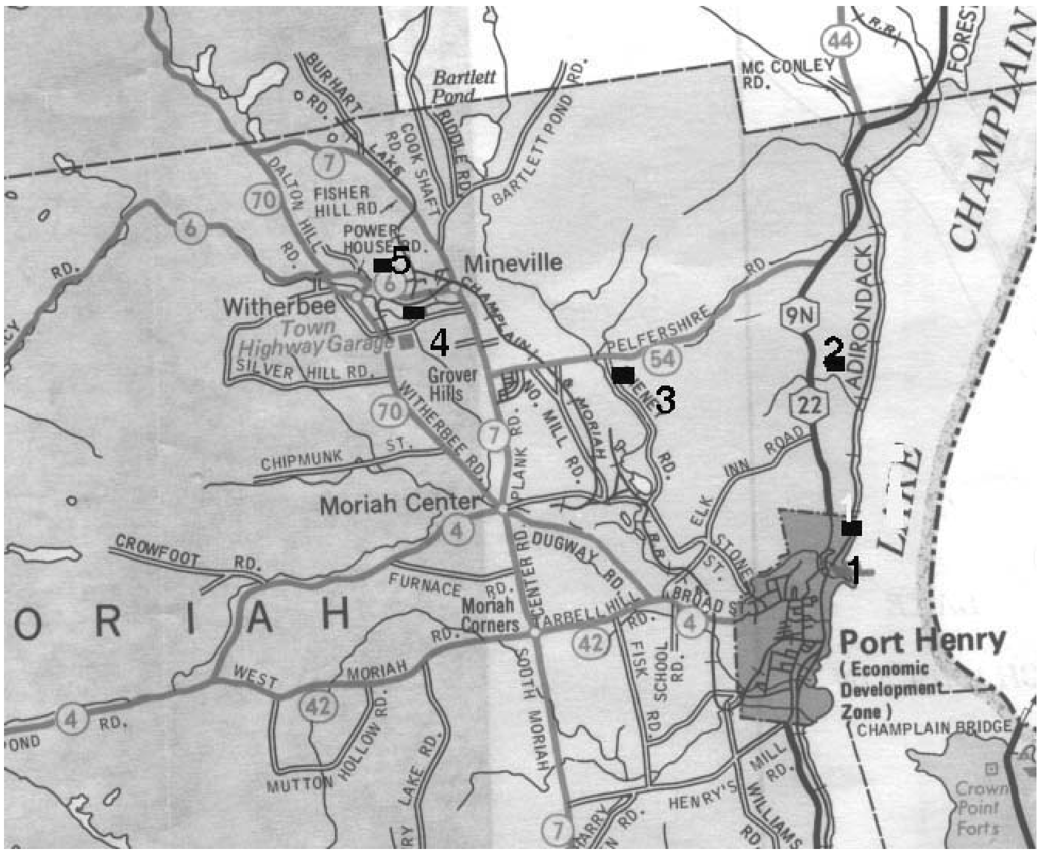

Figure 13. Road map of the field trip. 1. Craig Harbor Mine; 2. Cheever Iron Mine; 3. Pelfshire iron Mine; 4. Mineville group of mines; 5. Barton Hill group of mines.

The field trip starts at the Port Henry Boat Launch Site. The site is at the intersection of Dock and Velez lanes in Port Henry, Essex County. Follow the Amtrack railroad north but be careful because the Amtrack passengers and freight trains are coming fast and cannot be seen very easy due to the curves. Do not move or stay on the tracks. There is enough room to stand and walk safely on the side of the railroad. Also, be aware of poison ivy.

Stop 1. Craig Harbor Mine (N 44º 03’ 47.9”; W 73º 27’ 0.00”)

200m. Stop 1a. Upper Cambrian limestones with layers of chert. Small vugs and veinlets with pyrite are along the chert layers. Vertical faults cut the limestones.

100 m. Stop 1b. Craig Harbor faultline scarp. Beautiful view of the Craig Harbot faultline scarp. On the southern side of the fault there are Ordovician limestones and on the northern side there are Precambrian rocks. Here, a gabbroic intrusion is overlain by a metasedimentary sequence containing gneisses and marbles with deformed inclusions of gneisses, calcsilicates, metagabbro and quartzites. A very good description of the outcrop is in McHone (1987). The faultline zone was mentioned first by Emmons (1842); he noted inclusions of high-temperature minerals in the marble and considered them of igneous origin. Wiener et al. (1984) and McHone (1987) considered the marble to overlie the Eagle Lake gneiss. Roden-Tice (2002) reported that a sample from the gneisses below the marble yielded apatite-fission-tracks (AFT) age of 123 Ma.

200 m. Stop 1c. Craig Harbor Mine. The ore-hosted metagabbro displays micro- to coarse-texture: the coronitic character is obvious in places. The metagabbro contains garnet, hastingsite, phlogopite, plagioclase (oligoclase), clinopyroxene and orthopyroxene. The magnetite-ilmenite ore is in a pyroxene hornblendite layer (intrusion?) in the gabbro. Narrow veins of pyrite cut the iron-titanium ore. Whitney and McLelland (1975) and McLelland and Whitney (1980) studied the coronitic gabbros across the Adirondacks and determined that this texture formed in the granulite facies conditions of 8 kb and 800º C.

Cumulative mileage	Miles from the last point	Route description
1.5 miles	1.5 miles	Follow Rt22/9N toward north. Stop on the right side of the road in front of the metal gates. Walk around the gates and follow the unpaved Jeep road going north.

Stop 2. Cheever ore body (Port Henry mines): N 44º 04' 47.9"; W 73º 27' 10.1"

250 m. Stop 2a. There is a marble outcrop on the side of the Jeep road. The marble contains deformed inclusions of metapelites, calc-silicates and quartzites.

500 m. Stop 2b. On the left side of the road is the Witherbees-Sherman pit that opened the Cheever ore body. Fragments of felsic rocks with granitic to dioritic compositions containing layers of magnetite with sharp contact with the host are piled above the pit opening. Be careful and do not go very close to the rim. The gabbro outcrops on the right side of the road. Examine the texture, the mineral composition, the nature of the contact, and the gabbro outcrop.

500 m. Stop 2c. On the left side of the road there is a small mining work that opens the contact between the magnetite ore and the host rock. Examine the contact, the mineral composition of the ore, and the host rock. Coarse aggregates of amphiboles and clinopyroxenes are common in the ore.

250 m. Stop 2d. A large pit appears on the left side of the road. Take the left avoiding the pit and climb an easy slope on a deer trail heading north-north west. The trail goes between small pits and after 25 m you will reach a barbwire fence. Do not cross, just follow it along the trail. After 5 m another barbwire fence appears; go over and after 50 m you are at the Three Holes Mine. The host rock and the magnetite "layers" can be examined in the mine (the hanging wall and footwall are solid and there is no danger to going inside). Flash lights are not necessary because the Three Holes Mine was a very small operation and the day light is enough to help us to see the contact of the ore and to study the mineral composition of the ore and the host rock.

Go back to the parking place following the same Jeep road you hiked in along. Head north on Rt. 22.9N.

3.4miles	1.9 miles	Turn left onto Pelfshire Road.
4.4 mile	1 miles	Turn left onto Cheney road.
4.9 miles	0.5miles	Stop for the Pelfshire mines (pits).

Stop 3. Pelfshire (Pilfshire) mines (pits): N 44º 04' 30.8"; W 73º 29' 17.7"

The Pelfshire (Pilfshire) mines were mentioned by Smock (1889) and described by Kemp (1908). The ore and the host rock are similar to the Port Henry mines (Cheever ore body). The gabbro forms a ridge not far from the pits. The magnetite ore strikes north-south and dips 60º west; it is hosted by a felsic rock with transitions to more mafic compositions. Examine the characters of the gabbro, the host rock and the magnetite ore. The pits and dumps are the property of Rhodia Inc. (New Jersey). Ask for permission to visit and collect samples (Mr. Paul Tramblee of Mineville is the guardian of the property).

5.4 miles	0.5 miles	Turn back to the Pelfshire Road. Turn left on Pelfshire Road.
6.3 miles	0.9 miles	Turn left onto Swichback Road.
7.6 miles	1.3 miles	Turn right onto Titus Road.

8.3 miles	0.7 miles	Turn right onto Essex County 7.
10.2 miles	1.9 miles	Turn left onto Raymond-Wright Road.
11.2 miles	1 mile	Stop at the Firehouse. Park the car in the parking lot toward the picnic area.

Stop 4. Mineville group of mines: N 44º 05’ 22.5”; W 73º 31’ 30.5”

We will examine the rocks from the waste dumps of the Clonan, Joker and Bonanza shafts. Large fragments of magnetite-apatite, magnetite-apatite-pyroxene, magnetite, and magnetite-pyroxene ore display interesting textures. Examine the mineral composition and the texture of the ore and the host rock. The dumps are the property of Rhodia Inc. (New Jersey). Ask for permission to visit and collect samples (Mr. Paul Tramblee of Mineville is the guardian of the property). Turn left (SW) on the Raymond-Wright Road when you are done with this stop.

11.5 miles	0.3 miles	Follow the Raymond-Wright Road and turn right on the Barton Hill Lane
11.8 miles	0.3 miles	Park the car on the left side of the road. The mine is on the right side of the road.

Stop 5. Barton Hill mines: N 44º 05’ 34.6”; W 73º 32’ 05.2”

The Barton Hill group of mines is located half mile to the northwest from Mineville. Half mile north of the Barton Hill group is the Fisher Hill mine (actually on and under the Moriah Shock Correctional Facility). Here, we will examine the host rock with gabbroic character, its fabrics and the large body of pegmatite with magnetite and allanite-Ce cutting the gabbroic rock. The Lover’s Hole from the Barton Hill mines is the location where beautiful magnetite crystals came from at the end of the 19th century.

12.1 miles	0.3 miles	Turn right onto Powerhouse Road that continues as Essex County Route 6.
20.1 miles	8 miles	Turn right onto Route 9.
20.1 miles	0.1 miles	Turn left onto I-87.

End of the field trip

TIMING OF FLUID ALTERATION AND Fe MINERALIZATION IN THE LYON MOUNTAIN GRANITE: IMPLICATIONS FOR THE ORIGIN OF LOW-Ti MAGNETITE DEPOSITS

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INTRODUCTION

The Lyon Mountain granite (LMG), which crops out extensively in the northeastern Adirondack Highlands (Fig 1), is the host to numerous Kiruna type low-Ti magnetite deposits. The LMG experienced extreme metasomatic alteration by potassic- and sodic-rich fluids related to the Fe mineralization and during which time high field strength elements (HFSE) including Zr Y, U, and light and middle rare earth elements were highly mobilized during Fe mineralization.

This field trip will examine features associated with Fe mineralization, alkali metasomatism, the relative and absolute timing of the processes involved in the formation such deposits and the tectonic context in which these processes have occurred. Such processes are complicated and debate is encouraged.

GEOLOGIC SETTING

The LMG is a 1060-1045 Ma, syn- to late-tectonic granite that intruded the ~1150 Ma anorthosite-mangerite-charnockite-granite (AMCG) suite in the Adirondack Highlands (McLelland et al., 2001). Intrusion is syn-extensional between 1045 and 1037 Ma along the Carthage Colton shear zone in the western Adirondacks (Selleck et al., 2005). The Chateaugay mine is situated in the most northern and eastern region of exposed LMG (Fig. 1). By the early 1950s 15,000,000 tons of ore had been taken from the Chateaugay mine (Postel, 1952. This is one of several mines in and around the town of Lyon Mountain, New York. This Fe mineralized zone in the LMG lies near the northwestern flank of a large dome of Hawkeye granite. The Hawkeye granite intruded between 1103 and 1093 Ma (McLelland et al., 2001). Its age and characteristic quartz ribbon lineations provide a maximum age for the Ottawan orogeny in the Adirondack Highlands. A magnetic anomaly associated with Fe mineralization can be seen wrapping around the dome (Postel, 1952). This association suggests that either the mineralized zone was folded during subsequent doming/core complex formation, or that post-tectonic fluids penetrated into the LMG along a pre-existing structural interface with the Hawkeye granite. It is common throughout the LMG to find Fe mineralization in close proximity to the contacts of the LMG with other units, both igneous and sedimentary.

LITHOLOGY

At the Chateaugay mine and throughout the LMG three dominant lithologies persist; mesoperthite granite or granite gneiss, microcline granite and leucocratic albite/antiperthite granite

Mesoperthite granite

The granite consists mainly of mesoperthite, quartz, minor clinopyroxene and amphibole, biotite, minor apatite, titanite and zircon, and ubiquitous magnetite. This is the most common rock type in the LMG and probably was the original granite prior to alkali metasomatism. The unit is a pink, medium grained with varying amount of ferro-magnesian minerals. The unit may exhibit a gneissic fabric, which varies from locally penetrative to non-existent. Lamellae in the perthite grains are very coarse, sinuous, and free of micropores. These characteristics are consistent with perthites that have experienced metamorphism and deformation (e.g., Parsons et al. 2005).

Microcline granite

Microcline granite is representative of rocks in the LMG that have undergone extreme K metasomatism. Microcline granite consists of microcline as the dominant feldspar, quartz, minor clinopyroxene, disseminated magnetite, and minor titanite, apatite and zircon. The unit is pink to reddish pink and granoblastic. Any fabric is difficult to discern, as the main constituent minerals are equigranular. This suggests that these rocks experienced recrystallization after potassic fluid alteration if fluid temperatures were initially low or that fluid temperatures remained hot enough for triclinic microcline to form instead monoclinic orthoclase.

Leucocratic albite/antiperthite granite

These Na-rich granites are the result of rocks that have undergone complete (e.g., albite granite) or partial (e.g., antiperthite) albitization. Evidence that the albite granite and the antiperthite are the result of replacement is common. Both the perthitic granite and the microcline granite have been observed, at the outcrop as well as thin section scale, being crosscut and replaced by albite. Leucocratic albite granite is typically comprised of albite and quartz, minor amphibole, clinopyroxene, biotite, chlorite, magnetite, and minor titanite apatite and zircon. The unit is white to light gray and has a granoblastic texture. Fluorite has been observed in rocks that have experienced albitization.

LOW-Ti MAGNETITE DEPOSITS

Low-Ti magnetite deposits are part of larger class of deposits known as iron oxide (-Cu-Au) (IOCG) deposits. The characteristics of these deposits vary significantly from locality to locality. Olympic Dam in south central Australia contains significant amounts of Cu, Au and U, of significant economic proportions, while the dominant ore at Kiruna in northern Sweden, is magnetite and apatite (e.g. Reynolds, 2000; Harlov et al., 2002). The unifying characteristics of all these deposits are low-Ti Fe-oxides (magnetite and hematite, or pseudomorphic replacement of magnetite by hematite – so called martite), extreme hydrothermal alteration (Na, K, Si, seritization), and occur in extensional tectonic settings (Hitzman et al., 1992).

Low-titanium magnetite deposits in the LMG are associated with Na alteration of both the microcline and perthitic lithologies listed above (McLelland, 2002). The ores are comprised of magnetite, and/or hematite, and typically include quartz, apatite, feldspar, and clinopyroxene. The LMG ore bodies are associated with shear zones, the hinge regions of folds, and at the contact of the LMG with other units (Postel 1952, Whitney and Olmsted, 1993). The ore bodies generally conform to the foliation of the host granites but locally crosscut the gneissic fabric, where present, at a high angle. Barren zones nearly devoid of magnetite occur in the granite immediately adjacent to the ore bodies (Hagner and Collins, 1967). Outside of these barren zones disseminated magnetite occurs in the host granites.

Previous workers have discussed the origin of magnetite ores in the LMG. These include: 1) differentiation of immiscible Fe-rich fluids derived from the latest stages of granitic magmatism (e.g., Postel, 1952; Buddington, 1966); 2) preexisting mafic silicates that have broken down during deformation and metamorphism (Hagner and Collins, 1967); 3) eruption of Fe-oxide magmas and their interaction with evaporitic sediments (Whitney and Olmstead, 1993); and 4) brines that have interacted with the latest stages of pluton emplacement (McLelland et al., 2002).

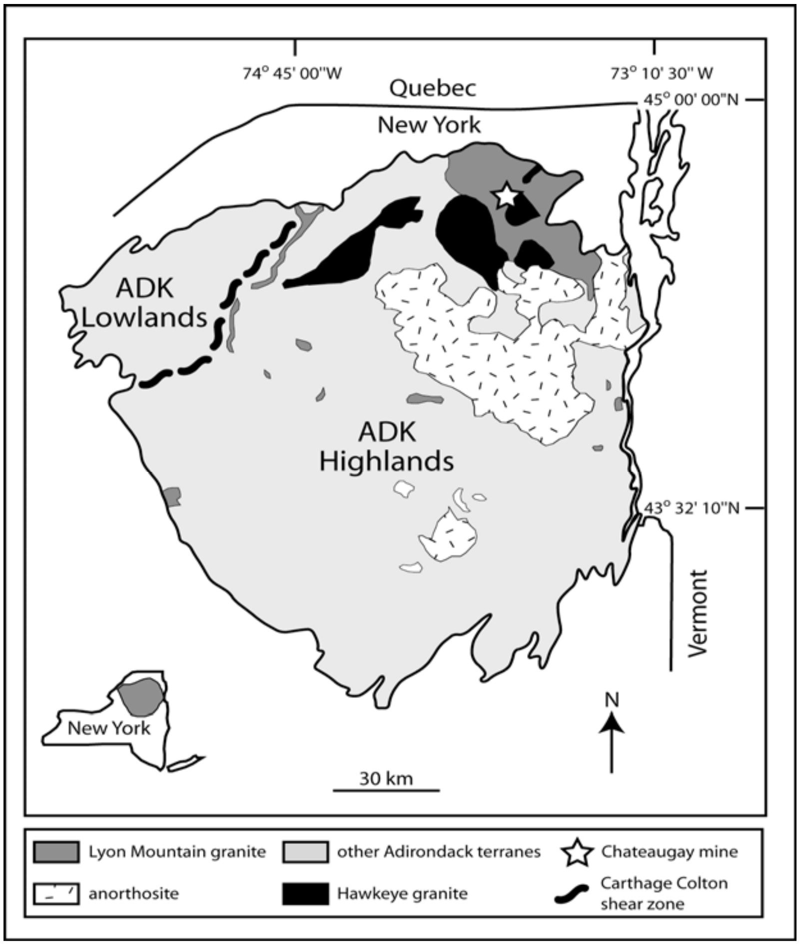

Figure 1. Generalized geologic map of the Adirondack Mountains.

ROAD LOG

This field trip has only one stop which is at the large open pit Chateaugay mine in Lyon Mountain, NY. To reach the mine from the parking area involves a short hike of approximately 200m. Accessing the mine is not difficult but extreme care should be taken while at the mine especially if the rocks are wet. Hiking boots are necessary since there is much loose rock in and around the mine some which we will traverse. Permission to visit the mine must be made in advance with the town office in Dannemora, New York. Hammers are welcome and flashlights may be useful for looking into adits and areas in shadow.

Mile

00.0	Exit 38N off of I87, from exit ramp follow routes NY 374W and NY 22N
00.2	Turn right at first light on route 22N.
00.7	Meet at Shell gas station.
	Left out of gas station parking lot on route 22S. Proceed back to traffic light, turn right onto route 374W. Outcrops of Beekmantown dolostone on left.
03.6	Cross route 190, continue west on route 374.
13.0	Clinton County Correctional Facility on right. Downtown Dannemora.
14.2	Outcrops of microcline granite on right side of road.
15.0	Outcrops of highly migmatized metasediments on left side of road

19.7 Chazy Lake on left. View of Lyon Mountain to the SW. This mountain is a large dome of Hawkeye granite
26.2 Just past Mobile gas station, turn left onto Standish Road
26.9 Left onto Power House Road
27.2 At "T" in road, turn left. Park on side of road just before gate to town sand pile

To reach the mine, proceed past the gate and around to the right around an old garage until you reach the first dirt road. Turn right on old dirt road. Follow road past a small old building and follow road to the right. After Approximately 30 meters the road forks. Follow the right fork. After only 3-5 meters there is a small overgrown road on the left goes downhill for 20 meters where it opens in the mine quarry. The area from the where the cars are parked to the mine is modified from time to time by logging operations, so the details above may change a bit. Total distance from the cars to the mine should not be more than a few hundred meters.

STOP 1: CHATEAUGAY MINE, LYON MOUNTAIN, NEW YORK

The Chateaugay mine is one of a series of large open pit and underground mines in the Lyon Mountain area. Mining began in the town of Lyon Mountain in 1867. By the early 1950s 15,000,000 tons of ore had been taken from the Chateaugay mine (Postel, 1952). The Lyon Mountain mines closed in 1967. Much iron remains but removal is not cost effective compared with cheaper foreign sources.

The strike of the quarry is approximately SW. The footwall and hanging wall of the ore body are perfectly exposed on the SE and NW walls of the pit, respectively. The ore follows the general trend of the pit and the gneissic layering, but locally crosscuts the gneissic banding (Gallagher, 1937). The local structure here appears to be that of a monocline with dip increasing towards the hanging wall. Overall the ore bodies here occur as "cigar-shaped" shoots in the limbs and core of a synclinal structure that plunges to the NE (Postel, 1952). These NE plunging structures are sometimes mimicked on a smaller scale by mafic and magnetite layers when observed in overhanging ledges in the quarry wall. Locally the ore may pinch and swell and may be openly folded.

Magnetite is the constituent ore mineral and is associated with pyroxene skarn and pegmatite and locally apatite and quartz. Magnetite rims pyroxene at both the outcrop and thin section scale. Late amphibole and chlorite are also associated with the pyroxene.

New age data from U/Pb zircon dating of various samples from the LMG and associated rocks will be discussed doing the field trip but are not listed here as they are in the process of being published elsewhere in a peer reviewed journal.

Mylonite

On the SE wall of the mine is a mylonitic shear zone (Fig. 2). The shear zone appears approximately parallel to the gneissic layering of the granite. "In-place" measurement is not possible as the mylonite is exposed high in the footwall of the mine and is not easily accessible. Blocks of the mylonite, however, lay at the base of the wall. Mylonitization is accompanied by partial melting of the felsic component of the host rock. Thin sections from this mylonite show that Fe mineralization is contemporaneous with shearing (Fig 3). Mineralogically, the mylonite contains remnant perthitic feldspar and microcline. The matrix consists of fine-grained albite, quartz, calcite, "strings" of magnetite and minor apatite and titanite.

Migmatites

In the hanging wall, on the northwest wall of the quarry immediately opposite the mylonite, is an approximately meter wide zone of migmatization. Here, the felsic component of the rock has melted leaving behind the more competent mafic rock to be deformed in a brittle manner. In places leucosomes are clearly rimmed by mafic selvages indicating that melting took place relatively in-situ and that these zones of melting are not intrusions into the LMG. The migmatite zone continues along the quarry wall. The felsic portion of this rock consists of quartz and coarse microcline. The mafic portion is dominantly cpx with late amphibole. Sugary-textured microcline granite borders the migmatized zone consisting of quartz, microcline and clinopyroxene.

Localized migmatite zones may also be evidence for fluid pathways. If this is true and the fluids are related to Fe mineralization then these fluids must have been hot enough to melt the felsic component of the rock. Fluid temperature estimates between 675 and 565 °C were determined from quartz-magnetite $\delta^{18}O$ fractionation elsewhere in the LMG (McLelland et al., 2002). Melting of granitic rocks in the presence of even small quantities of H_2O may begin at 600 °C (Robertson and Wyllie, 1971). Experimental data for granitic melts in the presence of F show that melt may be present as low as 550 °C (Manning, 1981). The presence of fluorite and F bearing alteration minerals and the previous temperature estimates by McLelland in the LMG suggests that partial melting during fluid alteration is a reasonable assumption.

Ore

The quarry steps down ~5 meters just beyond the migmatite locality. This step provides an excellent cross section view to the northeast of un-mined ore and the contact of the ore with the hanging wall and footwall (Fig.2). The ore here varies from "strings" or layers of "lean" ore to massive magnetite. The layers of magnetite in the "lean" ore are generally concordant with the gneissic fabric of the host, but locally these strings of magnetite are highly variable in orientation. Pegmatites containing amphibole and orange potassium feldspar are common near the edges of the ore as well as pyroxene skarn. There is a general transition from massive ore to layered magnetite + pyroxene skarn + pegmatite to barren granite (see below) to microcline granite with disseminated magnetite in the hanging wall (Fig.2). The contact of the ore with the footwall is not well exposed here. Further along in the quarry to the southwest a transition from ore to layered magnetite, to perthitic granite with pyroxene and disseminated magnetite occurs in the footwall. The ore commonly consists of magnetite, quartz, pyroxene and varying amounts of apatite and remnant perthitic feldspar.

"Barren" Granite

Disseminated magnetite is ubiquitous throughout the LMG. In the granite immediately adjacent to the ore body, however, there is commonly found a zone depleted in magnetite. These so-called "barren" zones are most likely a source for much of the magnetite in the ore body (the breakdown of Fe bearing pyroxene may be another). The barren zone can be traced along the length of the quarry just above the ore body and appears to be "sandwiched" between two zones of "lean" ore. There is possibly a second barren zone above the top "lean" ore but this is inaccessible and as such difficult to verify. The width of the barren zone is ~2 meters in most places. Approximately 50m beyond where the floor of the quarry steps down is an easily accessible place to view the barren zone on the NW side of the quarry. The rock is dominated by microcline and quartz with essentially no ferromagnesian minerals. Remnant pyroxene where present, usually occurs as inclusions in "forest" green amphibole. Late albite crosscuts microcline and there is some Ca alteration as well in the form of calcite. These zones depleted in magnetite are common around most of the ore deposits in the LMG (Postel, 1952). At the Palmer Hill mine this zone is up to 150m wide (Hagner and Collins, 1967).

Crosscutting Dike

Approximately 250 meters from the entrance to the quarry, located on the northwest side of the quarry, is a dike that crosscuts the gneissic fabric of the granite. The dike consists of quartz, plagioclase, magnetite, and minor sulphide and is approximately one meter wide. The composition of the plagioclase is oligoclase (McLelland, 2001). We will discuss the U-Pb age of this dike and its implications for Fe mineralization and deformation of the ore and host granites. A number of cross-cutting pegmatitic dikes are present throughout the quarry and range from plagioclase rich to syenite pegmatites to granite pegmatite and younger cross-cutting diabase dikes. These dikes do crosscut the ore but seem closely related, exclusive of the diabase dikes, to the mineralization and fluid alteration based on their chemistry and mineralogy.

Possibly related to these pegmatitic dikes are miarolitic cavities that also truncate the layering in the ore and the granite. These cavities are often up to a meter long. One is said to be large enough for a man to stand in (Gallagher, 1937). These cavities may contain very large crystals of orthoclase more than a foot in length, quartz, titanite, apatite, agerine/augite, amphibole and late nearly pure albite. It is not clear how these formed but for open spaces this large they could not have experienced much deformation. These cavities are difficult to find in the quarry but a number of them are present. One can be observed on the backside of an ore pillar at the southwest end of the quarry. It is associated with an orange granitic pegmatite dike that crosscuts the layering in the ore.

Ore Pillar

Most of the quarry we have walked through has been filled in after mining ceased for safety reasons. However at the far end of the quarry it is still possible to view a deeper section of the mine. Here, two ore pillars are still standing where it is possible to view the true thickness of the ore. Access is difficult due to the steep slope, but the pillars can be viewed from a distance. The massive ore is approximately 3 meters thick with lean ore above and below. You may also be able to observe the NE plunging lineation in the rocks that overhang the ore pillar as well as another crosscutting dike, the barren zone, and migmatite layer.

Another feature at the mine are discontinuous mafic layers. They consist of amphibole and/or biotite and plagioclase that take on a "salt and pepper" appearance. These mafic layers may exhibit gneissic banding internally and vary in thickness from a few centimeters to meters in width and are common throughout the LMG. During this study they were observed to be continuous beyond the limit of exposure. When exposures were better during active mining, these layers are reported to anastomose and pinch out, but can persist for over 100 meters (Gallagher, 1937).

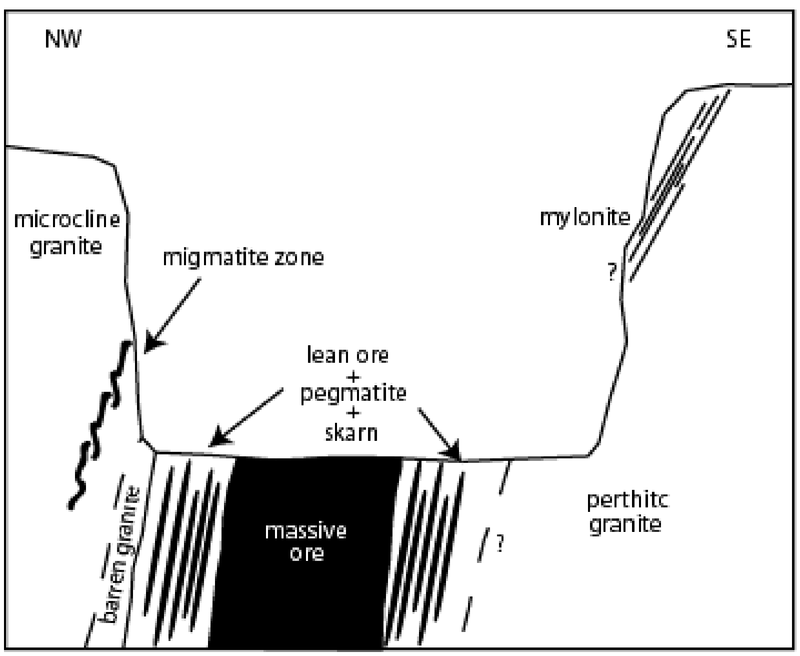

Figure 2. Generalized cross-section of the Chateaugay Mine with important features of this field trip labeled.

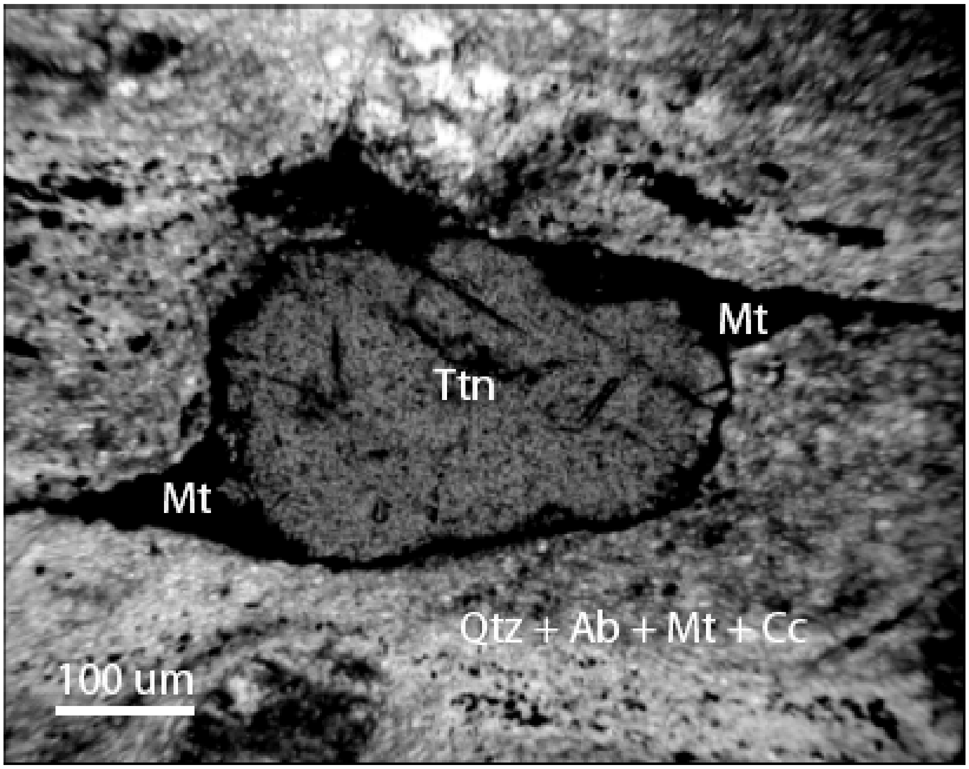

Figure 3. A plane polarized light photomicrograph from a mylonite in the Lyon Mountain granite. Photomicrograph shows magnetite growing during deformation. Mineral abbreviations after Kretz (1983).

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STRATIGRAPHY, SEDIMENTOLOGY AND DIAGENESIS OF THE POTSDAM FORMATION, SOUTHERN LAKE CHAMPLAIN VALLEY, NEW YORK

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INTRODUCTION

The Cambrian Potsdam Formation is widely exposed on the periphery of the Adirondack massif in northern New York State and adjacent Ontario and Quebec. As the basal unit of the Paleozoic sequence over the cratonic interior of eastern Laurentia, its distribution and character are strongly controlled by the geometery of the pre-Potsdam erosional surface, and patterns of rift-related faulting of the Iapetean margin and sea-level change. In the northern Lake Champlain Valley, the Potsdam is characterized by a basal arkosic unit, the Ausable Member, which is overlain by marine quartz sandstones of the Keeseville Member (Fisher, 1977) Recent study of key exposures in the Lake Champlain Valley region (Landing, 2007) have more firmly established age relationships of the Keeseville and Ausable Members of the Potsdam. Landing (2007) has also documented the existence of a Middle Cambrian unit, the “Altona” Formation, which underlies the Ausable Member of the Potsdam Formation in the northern Lake Champlain region.

The Potsdam Formation is approximately time-equivalent to the Danby and Winooski Formation of western Vermont (Mehrtens and Butler, 1987; Landing, 2007). The Potsdam is thickest in the northern Champlain Valley and adjacent southern Quebec (~250 meters), likely related to accumulation on the downthrown side of an Iapetan margin fault (Landing, 2007). In the area of this field trip, the total thickness reaches a maximum of 80 meters. A locally-developed basal arkosic unit may be correlative to the Ausable Member. The succeeding quartz-rich sandstones are referable to the Keeseville Member.

The Potsdam Formation acted as an aquifer for the transport of basinal brines and hydrothermal diagenetic fluids during burial. Significant diagenetic alteration of the original sediment has occurred, and the Proterozoic basement rocks beneath the Potsdam have been altered by hydrothermal fluids (Whitney and Davin, 1987). This alteration may be related to seismic pumping of fluids during Taconic (Late Ordovician), Acadian (Middle-Late Devonian) and/or Alleghanian (Carboniferous-Permian) tectonism on the Laurentian margin.

In the southern Lake Champlain Valley and Lake George Valley, the Potsdam records non-marine, tidal flat and nearshore marine depositional systems. This field trip will include exposures of a variety of sedimentary facies within the Potsdam Fomration, and will also examine relationships with underlying Proterozoic basement and burial diagenetic and hydrothermal features within the sandstone and basement rock.

STRATIGRAPHY AND SEDIMENTOLOGY OF THE POTSDAM FORMATION IN THE SOUTHERN LAKE CHAMPLAIN VALLEY

Lower Potsdam Formation (=Ausable Member?)

The Potsdam Formation, in the area of this field trip, includes a basal sequence of arkosic sandstone and conglomerate that reaches a maximum thickness of ~30 meters. The succeeding marine quartz sandstones of the Keeseville Member reach a maximum of ~50 meters. The basal arkosic unit is discontinuous in distribution, with the most extensive and thickest development limited to the area south of Ticonderoga Village along NYS Route 22 (Figure 1). Outside of this area, the Keeseville Member rests directly on Proterozoic basement. The basal arkosic unit is dark green to gray-green or yellowish green in color, massive to thick-bedded, irregularly laminated, cross-laminated fine pebble conglomerate and sandstone. Reverse and normal graded beds 5-20 cm thick are common, and trough cross-beds are present. This unit

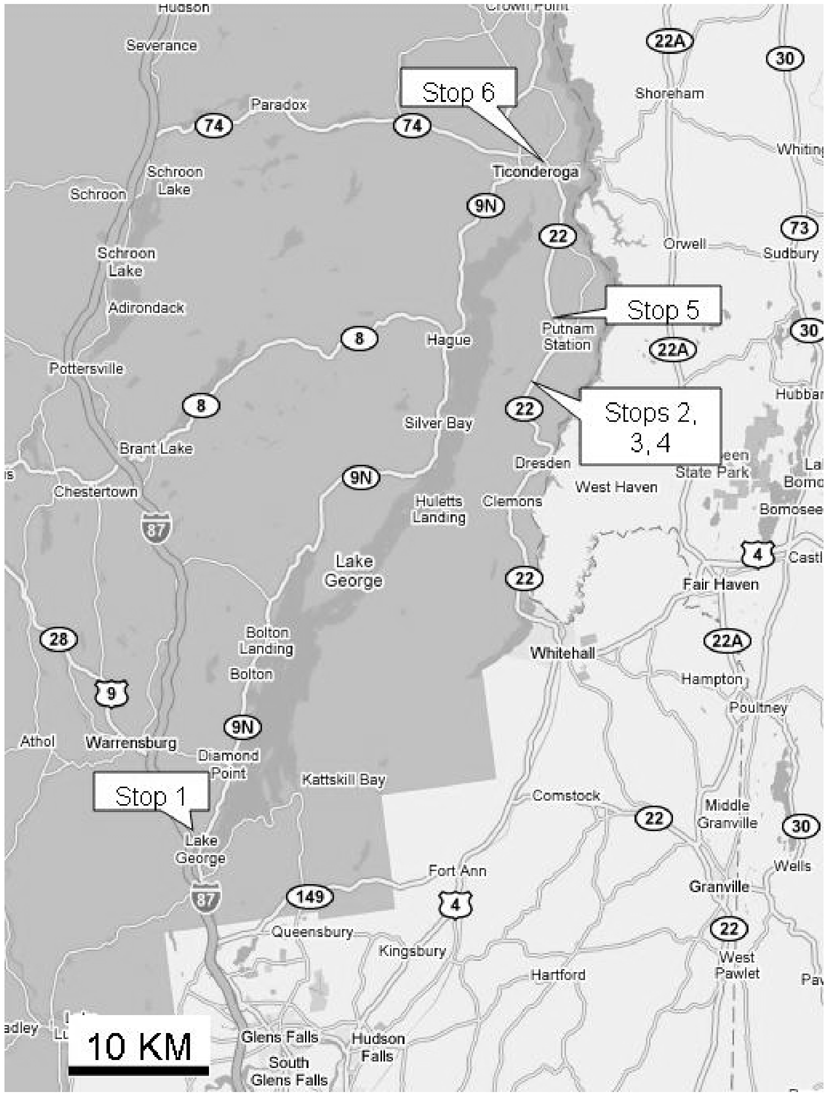

Figure 1 - Location Map
(from Google Maps)

resembles the Ausable Member of the Potsdam Formation in the northern Lake Champlain Valley. The depositional environments of the basal arkosic unit have been interpreted as alluvial fan and braided stream, based upon sedimentary textures, bedded thickness and primary structures (McCrae, et al, 1986). The source area of the sediment was local, since clasts of granitic and calc-silicate gneiss are common in the conglomerates of the basal unit.

The age of the basal arkosic unit in the southern Lake Champlain Valley is not well-constrained. Fisher (1977) proposed that the Ausable Member and related basal arkosic faciesof the Potsdam Formation might be late Proterozoic (Hadrynian) in age, based on the assumption that these facies accumulated in rift basins developed on the margins of the opening Iapetus Ocean. McRae, et al (1986) used paleomagnetic techniques to determine a poorly-cosntrained early Paleozoic depositional age for the Ausable member and other basal Potsdam Sandstone units in the northern New York State outcrop belt. However, the intense post-depositional alteration of the Potsdam Sandstone suggests that the primary depositional paleomagnetic signal was not preserved, and that the remanence measured is likely a diagenetic artifact. The recent description of a trilobite-bearing unit, the early-middle Middle Cambrian "Altona Formation", beneath the Ausable Member in the northern Lake Champlain Valley (Landing, 2007), requires a Middle Cambrian age for type Ausable Member. In the southern Lake Champlain Valley, quartz sandstones and carbonates of the marine Keeseville Member of the Potsdam contain a uppermost Middle Cambrian trilobite fauna (Landing, 2007). The marine facies of the Potsdam in this area contain local beds of coarse, locally derived clastics, and the contact between the basal arkosic facies and overlying Keeseville is apparently conformable, and represents a marine transgressive surface. These factors suggest relatively continuous deposition during the time interval represented by the basal arkosic unit and overlying marine Keeseville Member, and thus a middle-late Middle Cambrian time of deposition for the basal unit is suggested (Figure 2).

The green basal Potsdam facies contains abundant iron-rich chlorite (Selleck, 1997) that is consistent with diagenetic temperatures in excess of 200°C. The muddy matrix of the sandstones also contains illite and minor kaolinite. Quartz cement and overgrowths of authigenic feldspar on detrital grains are common. Titanium oxide (anatase) cements are present. Associated authigenic REE minerals (monazite, xenotime, allanite), sulfides (pyrite, galena, sphalerite), fluorite and barite suggest that hydrothermal fluids were transmitted through permeable basal sands, causing extensive alteration. Primary detrital minerals such as hornblende, bitotite, plagioclase, garnet and ilmenite are commonly altered to mixtures of chlorite, illite, anatase, quartz, K-feldspar and pyrite. Fluid inclusions in quartz veins which cross-cut the basal Potsdam in the study area have homogenization temperatures near 250 °C and salinities at or near halite saturation (Collins-Waite, 1991). Whitney and Davin (1987) suggested that hydrothermal alteration of Proterozoic basement gneiss in the Fort Ann, NY region was the result of explusion of connate brine from the overlying Paleozoic sediments during the Late Ordovician Taconic Orogeny. This proposition is supported by recent U-Th-Pb chemical age dating of monazite overgrowths in the basal Potsdam, which contain Late Ordovician (ca. 450 Ma) growth zones. Hydrothermal alteration of basement rocks in the area is also manifested by dolomitization of Proterozoic marble (Selleck, 1997). The dolomitization was driven by fault-related fluid flow and was likely contemporaneous with dolomitization of the overlying carbonate rocks in the Potsdam Sandstone and younger Beekmantown Group strata.

Upper Potsdam Formation (=Keeseville Member)

The quartz-rich sandstones of the upper Potsdam Formation are divisible into four lithologic packages which can be correlated over the southern Lake Champlain Valley region (Figure 3).

1. Lower stratified unit: Medium to fine-grained, parallel laminated and cross laminated subarkosic arenite. This unit has a sharp contact but conformable contact with arkosic facies below, often with a conglomeratic ravinement bed at the contact. Where the basal arkosic unit is absent, the lower stratified unit lies nonconformably on Proterozoic basement. Mudcracks and small, irregular burrows are present in silty carbonate-cemented sandstones near the base of the unit. Wave-ripple surfaces are common. Some beds are cemented by pyrite.

2. Lower cyclic unit: Medium to coarse sandstone and fine pebble conglomerate with decimeter-scale dolomite and dolomitic sandstone beds. Dolomitic beds may contain stromatolites, mudcracks and intraclast breccia. Bi-directional cross-stratification is prominent, forming trough cross-sets. Pebbles of locally-derived pegmatitic quarz and feldspar are common. Contact with the underlying lower stratified unit is sharp and marked by abrupt appearance of coarse, pebbly sandstone. Lower cyclic unit unconformably overlies Proterozoic basement in outcrops north of Fort Ann, New York.

3. Upper stratified unit: Medium bedded, coarse to medium-grained quartz sandstone with thin beds of dolomitic silty shale. Horizontal burrows (Rusophycos, Teichichnus) and vertical burrows (Diplocraterion, Skolithos) are present. Most beds are horizontally laminated with beds having overall lenticular form. Wave ripple lamination and wave rippled surfaces common. Contact with underlying lower cyclic unit is sharp, with uppermost beds of the underlying unit consisting of phosphatized, burrowed hardground. Bedsimmediately above the contact are dolomitic and contain abundant lingulid brachiopods and dolomitized shell debris.

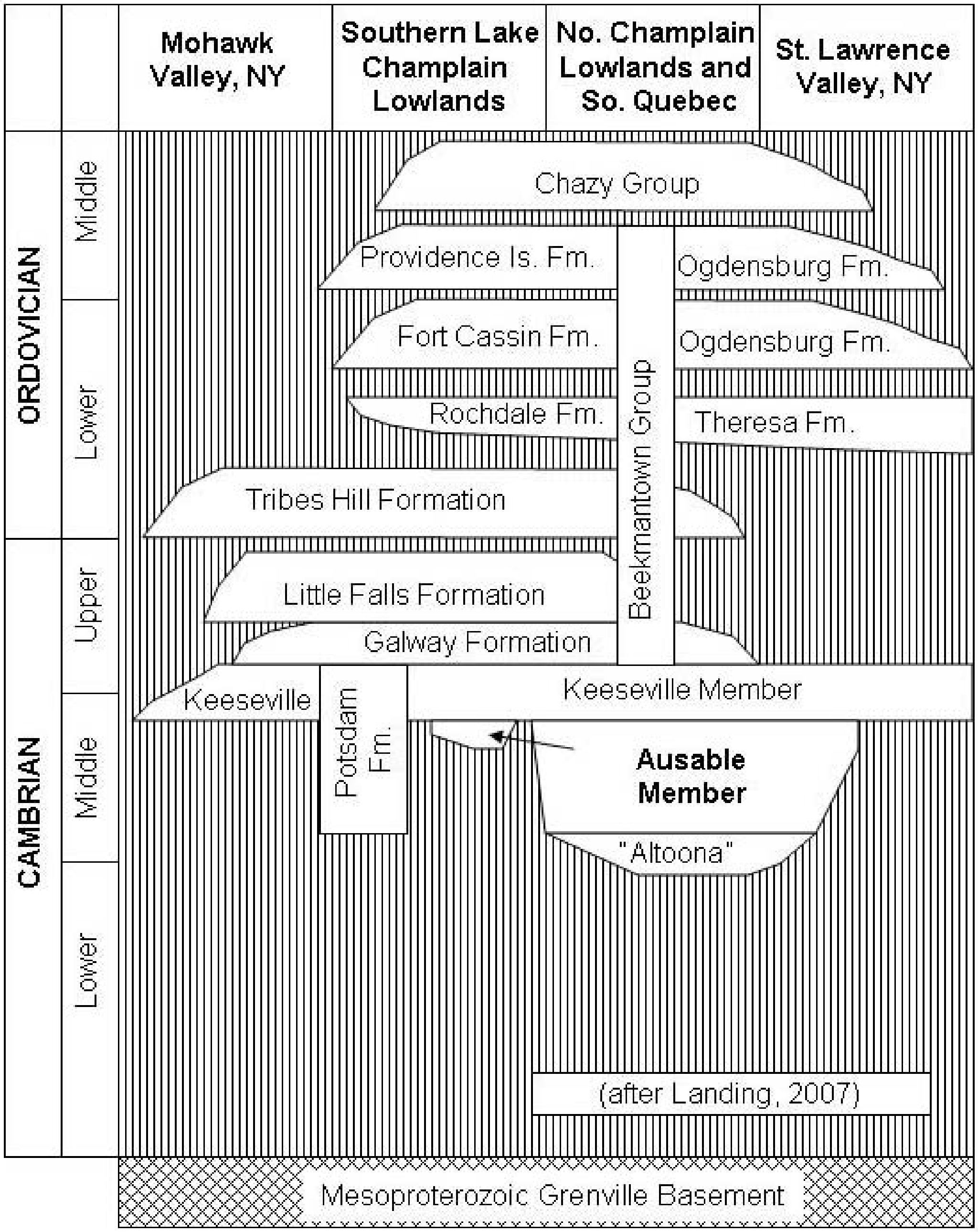

Figure 2 – Regional Stratigraphic Relationships. After Landing, 2007

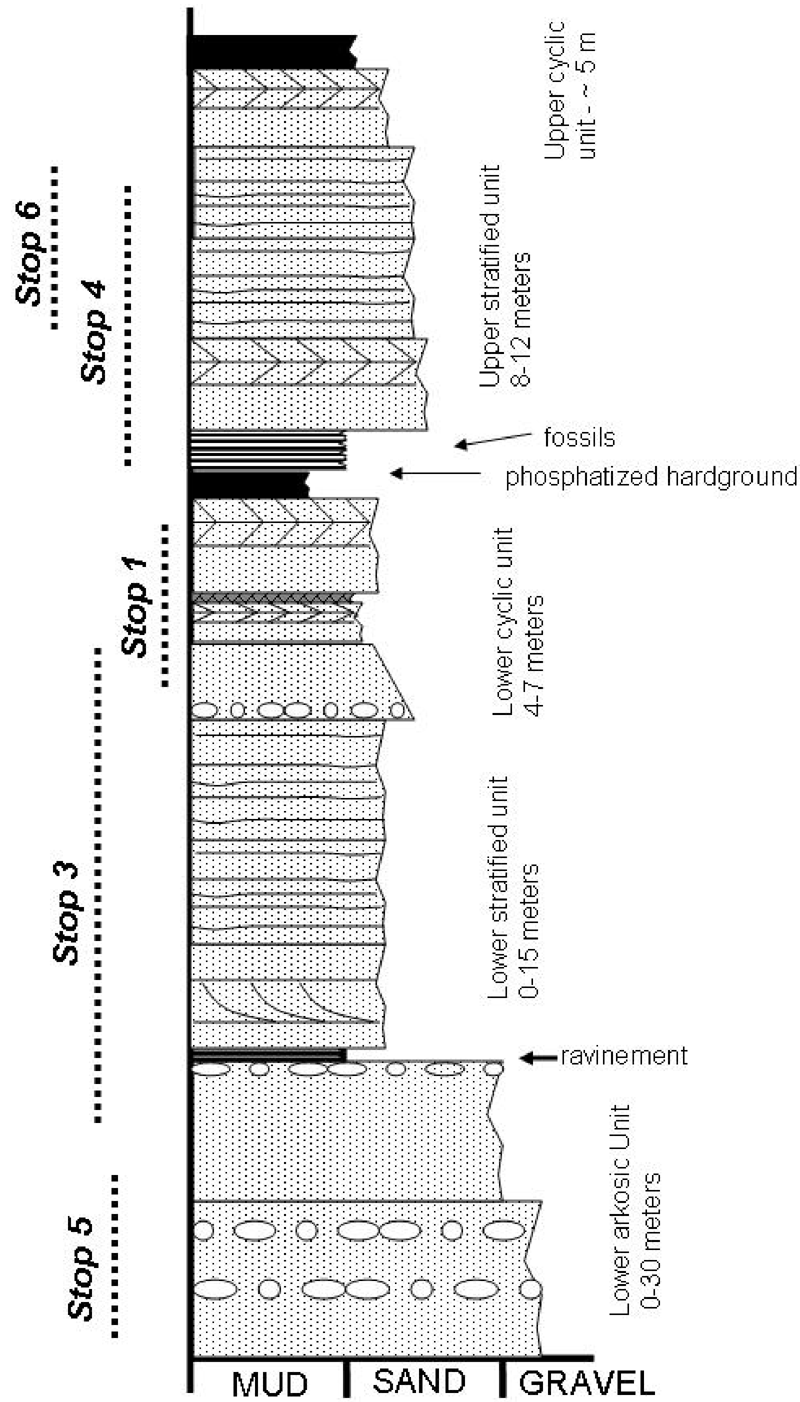

Figure 3 – Schematic composite columnar section for Potsdam Formation in Southern Lake Champlain Valley and Lake George Region

Dolomitic sandstone and calcareous siltstone forming the lowest portion of upper stratified unit exposures near Fort Ann, New York contain an uppermost Middle Cambrian fauna (Landing, 2007). The large trace fossil Climactichnites has been described from quarried slabs of the Potsdam Formation from Port Henry, New York (Yochelson and Fedonkin (1993). Although no specimens have been observed in place, older quarries in the Port Henry area are within the upper stratified unit of the Potsdam, and the slabs bearing Climactichnites at Port Henry most closely resemble the upper stratified unit. Yochelson and Fedonkin (1993) concluded that Climactichnites is restricted to upper Middle and Late Cambrian strata, consistent with the occurrence of uppermost Middle Cambrian trilobites in the upper stratified unit.

4. Upper cyclic unit: Thick to massive bedded, medium to fine quartz sandtone with decimenter-scale beds of burrowed, carbonate cemented fine sandstone. The burrowed units recur at 1-2 meter intervals. Vertical burrows are common, and mudcracks are present in carbonate-rich intervals. Contact with the underlying upper stratified unit is gradational and marked by appearance of burrowed, carbonate-cemented beds.

The uppermost beds of the upper cyclic unit of the Potsdam Formation contain progressively more carbonate up-section, and grade into the basal Galway Formation, which consists of dolomitic sandstone and dolomite. This contact can be observed in the exposures east of the Champlain Canal in the village of Whitehall, New York, where the uppermost Potsdam is succeeded by brown-weathering dolomite and dolomitic sandstone of the Galway Formation in the ledges immediately east of Skene Manor. In the village of Ticonderoga, basal beds of the Galway Formation consist of cherty dolomite with calcite-filled vugs and irregularly weathered brown dolomitic sandstone. These beds are exposed on the northwest side of the village near Mt. Hope Cementary, with white-yellow weathering ledges of upper Potsdam Sandstone found in backyards and scattered field exposures down slope and south of the cemetery.

Sedimentary Environments

The upper Potsdam Formation (=Keeseville Member) in the study area were deposited in shallow marine shoreface, foreshore, offshore subtidal shelf and tidal flat settings. The lower stratified unit records a regional sea level rise, with the shoreline transgressing over non-marine basal Potsdam, where present, and over Proterozoic basement elsewhere. Shoreline stabilization and renewed input of coarse clastics are recorded in the coarse-grained, rhythmically bedded tidalite facies of the lower cyclic unit. A second pulse of sea level rise is represented by the phosphatized hiatus surface which caps the lower cyclic unit, and is succeeded by offshore subtidal facies of the upper stratified unit. This unit then shallows into the tidal flat facies of the upper cyclic unit.

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ROAD LOG

Cum. Miles	Miles from Last Stop	(this road log begins at the exit from the parking lot of the Fort William Henry Hotel in Lake George Village)
0.0	0.0	Exit parking lot and turn right (north) on NYS Route 9N
0.8	0.8	Bear to right at light; continue on NYS Route 9N
1.2	1.2	Turn left at light on I87 access road.
1.4	1.3	Park on right next to low road cuts. UTM 604427E, 48110007N (UTM is NAD83)

Stop 1 – Lower cyclic unit of Keeseville Member, Potsdam Formation

The low outcrops by the roadside display gray-white quartz arenite of the lower cyclic unit of the Potsdam Formation with interbeds of reddish-brown dolomite and dolomitic sandstone. The quartz sandstone units are cross-laminated and trough cross-sets are present. The uppermost layer in the outcrop is comprised of domal stromatolites – algal accretionary structures – that formed in a shallow, muddy tidal pond. Modern stromatolites having this geometry are developed in coastal saline lagoons such as Lake Clifton in Western Australia. The stromatolitic laminations are very well preserved, and the grainy material between individual heads includes quartz sand and dolomitized bioclastic material. Coarse intraclast breccia is also found between individual heads. Note the coarsely crystalline calcite within the cores of some stromatolites. This may be a replacement of gypsum that grew as an evaporite mineral in the sediment that hosted the stromatolites.

Note the minor faults in the outcrop. Some of the patterns of faulting appear to mimic the major graben structure of Lake George and the boundary between the Adirondacks and Champlain Lowlands.

Cum. miles	Miles from last Stop	
1.3	0.0	Turn around and continue to intersection with Route 9N.
1.4	0.1	Turn right (south) onto NYS Route 9N and continue south through Lake George Village
3.9	2.6	Interstate 87 entrance to right; stay on Route 9N south
6.2	4.9	Interesection with NYS Route 149; turn left (east) onto Rt. 149
10.6	9.3	Intersection with Route 9L; continue east on Rt. 149
17.9	16.6	Intersection with NYS Route 4 in Fort Ann, New York; turn left (north) on

		Rt. 4. Continue north
21.0	20.7	Intersection with NYS Route 22; continue north on Rts. 4+22
27.7	26.4	Intersection of Rt. 4 and Rt. in Whitehall; continue north on Rt. 22. Skene Mountain to east includes exposures of Potsdam Formation and overlying Beekmantown Group carbonates
41.9	40.6	Park on right next to low outcrop of marble. UTM 626390E, 4841026N

Stop 2 – Proterozoic marble with Paleozoic dolomitization features

The roadcut on the east side of Route 22 displays dolomitized Proterozoic basement marble. Note the very coarse crystal size of the unaltered calcite marble and the pseudomorphic dolomite in the dolomitized portions. Graphite with the calcite marble is not altered in the dolomitized marble. The coarse dolomite crystals contain mm-scale voids that are sometimes partially filled with tiny 'saddle' dolomite rhombs, and terminated calcite and quartz crystals, which resemble miniature Herkimer 'diamonds'. Note the vertical fractures in the marble which served as conduits for the dolomitiziting fluid. This is a good exposure to examine the effects of enhanced permeability, produced during the dolomitization process, on the development of a 'dolomitization front'. Alteration of calcsilicate minerals (clinopyroxene, brown phlogopite mica) in the dolomitized marble may have served as a source of magnesium for dolomitization. The relict altered minerals are now represented by chlorite and illite pseudomorphs. Pyritiferous, cherty, chlorite-bearing lithologies on the west side of the highway are also the result of hydrothermal alteration accompanying the dolomitization.

Saddle dolomite crystals within vugs in the dolomitized marble contain aqueous 2-phase fluid inclusions with homogenization temperatures of 160-210°C. Ice melt temperatures of these inclusions indicate salinity near halite saturation; similar high salinity inclusions are also found in sparry dolomite in the overlying Beekmantown Group carbonates (Collins-Waite, 1991). Stable isotopic analysis of the calcite and dolomite suggest interaction of the marble with dolomitizing fluids that were isotopically enriched relative to seawater, and that the fluids were strongly rock-buffered (Selleck, 1997).

The broad valley to the north of this road cut is underlain by Potsdam Formation; Stop 3 is visible approximately 2 km to the north along Route 22. The local geology suggests that the marble exposure here was close to the overlying Potsdam, and thus fluids within the Potsdam were readily available to invade the marble along vertical fractures. Enhancement of permeability within the marble by dolomitization increased the access of fluids to the basement rocks.

Cum. miles	Miles from last Stop	
41.9	0.0	Continue north on NYS Route 22
43.8	2.1	Park on right adjacent to southern end of outcrop. UTM 627686E, 4843461N

Stop 3 – Lower Potsdam Sandstone at Firehouse Road

These road cuts expose the upper part of the basal arkosic unit, the succeeding lower stratified unit and the lower cyclic unit of the Potsdam Sandstone. We will examine the basal arkosic unit in more detail at Stop 5. The contact between the basal arkosic unit and the overlying lower stratified unit is well-displayed on both sides of the road. Note the abundant large clasts at the contact. This layer is interpreted as a lag deposit on a ravinement surface formed by the reworking of previously deposited gravel and sand by the encroaching marine shoreline during sea level rise. The immediately overlying marine sandstone contains rare mudcracks, best seen in the silty layers less than a meter above the contact. Rare vertical burrows are sometimes visible in outcrop, but common in slabbed samples of the sandstone 2-3 meters above the contact. Note the dominantly horizontal stratification and tabular foreset cross-stratification. Trough cross-sets are also present. This facies is interpreted as a shallow subtidal foreshore deposit.

Continuing up-section at this outcrop, the base of the lower cyclic unit is marked by reappearance of coarse pebbly sandstone that documents shallowing to an intertidal depositional setting; availability of local Proterozoic basement source rocks is indicated by pebbles of vein quartz and feldspar. Well-developed bipolar cross lamination is clearly seen on vertical faces. The coarse pebbly sandstone intervals are capped by medium sandstones and carbonate beds at approximately one meter intervals. The carbonate beds consist of ferrroan dolomite with late calcite veinlets and vugs; and are interpreted as accumulations of carbonate mud on shallow upper intertidal flats. In other outcrops of this facies in the area, the carbonate beds contain stromatolites, mudcracks and intraclast breccia. Note the buckling and truncation of some of the carbonate beds. Sulfide mineralization is common in the carbonate, with pyrite, sphalerite and galena present. The dolomitization and sulfide mineralization are assumed to be related to the same hydrothermal event(s) which caused the dolomitization of marble seen at the last stop.

Cum. miles	Miles from last Stop	
43.8	0.0	Continue north on Rt. 22
44.3	0.4	Park on right side of Route 22. UTM 628278E, 4844038N. Be careful crossing the highway!

Stop 4 – Upper Potsdam Sandstone at Firehouse Road

The upper part of the lower cyclic unit and the succeeding upper stratified unit of the Potsdam Sandstone are exposed in the outcrops on the west side of Route 22. Scattered low outcrops to the south allow for a nearly complete section to be constructed in these exposures. The lowest beds on the southern end of the roadcuts are dolomitic sandstones of the lower cyclic unit. A phosphate mineralized hardground with large burrows is developed at the top of these beds. This surface, interpreted as a sediment starvation interval that records the onset of a period of sea level rise, marks the contact between the lower cyclic unit and the upper stratified unit. The succeeding beds of the upper stratified unit consist of sandy dolomite, dolomitic shale and silty sandstone. The dolomite beds contain fragmentary lingulid brachiopods, dolomitized bioclasts, vertical and horizontal traces including Rusophycos and Teichichnus, intraclast breccias and wispy phosphatized surfaces. This interval is apparently correlative to the beds near Fort Ann, New York that have yielded a uppermost Middle Cambrian trilobite fauna. Diligent searching at this outcrop has not recovered identifiable trilobite remains, but it is a likely target for further study.

The upper stratified unit coarsens rapidly above the basal beds with coarse bipolar cross-stratified sands documenting storm or tidal current effects. Medium bedded sandstone with thin silty shale interbeds form the bulk of the exposures. Note the lensoid character of the bedding in the middle and upper portion of the outcrop. The uppermost part of the outcrop is thicker-bedded and slightly coarser sandstone. The upper stratified unit facies is interpreted as a lower shoreface and foreshore subtidal deposit, documenting a deepening event followed by shallowing. Exposures in the forested area immediately north and west of this outcrop are within the lower portion of the upper cyclic unit, and record further shallowing to tidal flat facies. We will not examine these exposures on this trip. Note the prominent fault-line scarp to the west and north of the highway. We will cross this fault as we proceed north to Stop 5.

Cum. miles	Miles from last Stop	
44.3	0.0	Continue north on Rt. 22

45.8 1.5 Park on right side of Route 22. Be careful crossing the highway!

Stop 5 - Basal arkosic unit of the Potsdam Formation

The angular non-conformity separating banded Proterozoic calcsilicate gneiss and the basal arkosic unit of the Potsdam Formation is very well-displayed in road cuts on both sides of Route 22. The gneiss below was deformed in the Ottawan Orogeny, ca. 1050 Ma. The Potsdam here is ca. 520 Ma.

Medium to thick-bedded arkosic wacke, and arkosic arenite with abundant quartz and feldspar pebbles form the basal arkosic unit at this stop. Coarser lenses of pebble conglomerate occur discontinuously above the basal contact, and are sharply overlain by sandstone. These lenses represent an earlier phase of gravel deposition. The main body of sandstone contains normal and reverse graded beds. This facies was deposited by mass flow and sheet flood processes in a high-gradient fluvial system.

The gray-green color of the sandstone is due to iron-rich chlorite (25-29 wt. % total FeO). The chlorite formed as a diagenetic replacement of hornblende, garnet, biotite and plagioclase. Chlorite also occurs as an integranular cement with quartz, K-feldspar and anatase. Chlorite occurs as tabular crystals within fluid inclusions in vein quartz at this outcrop, suggesting that the vein-mineralizing fluids were similar to those causing the general diagenetic alteration. Chemical data indicate that the diagenesis of the basal arkosic facies involved exportation of sodium, magnesium and calcium from the original sediment. This is consistent with petrographic observations, notably conversion of plagioclase, hornblende, biotite and garnet to Fe-chlorite. Chemical data also require importation of potassium and iron by diagenetic fluids.

The magnesium lost for the primary detrital minerals during burial diagenetic alteration may have served as a source of magnesium for dolomitization of underlying marble (Stop 2) and overlying marine carbonates (Stop 3). Similar Fe-chlorite alteration occurs in Proterozoic basement rocks and these could also have served as a source of magnesium for dolomitization. The widespread occurrence of these features in the vicinity of faults in this region suggest that seismic pumping of diagenetic/hydrothermal fluids was an important process during fluid alteration of the basement and overlying sedimentary strata. Other metals (Cu, Pb, Zn) in the original depositional minerals were also mobilized and exported via hydrothermal fluids, documented by scattered, non-economic occurrences of galena, sphalerite and chalcopyrite in the Potsdam Formation sandstones and in carbonates of the Beekmantown Group.

The diagenesis of the basal arkosic unit also resulted in precipitation of REE-bearing minerals. The yttrium phosphate mineral xenotime is found as scattered grains within authigenic chlorite and as authigenic overgrowths on detrital zircon grains. Monazite occurs as ‘meshwork’ replacement of detrital allanite and as overgrowths on detrital monazite grains. Thorite is present as detrital grains and micron-scale crystals within authigenic quartz. The source of the REE was the original sediment, which contained detrital grains of allanite, which is rather common in the adjacent Proterozoic granites and REE from breakdown of minerals such as garnet, ilmenite and magnetite. Detrital grains of ilmenite and magnetite in the basal arkosic unit are often altered to intergrowths of chlorite and anatase. Authigenic pyrite, sphalerite, galena and chalcopyrite are common in the green sandstones at this locality.

Collins-Waite (1991) reported fluid inclusions with 200-300° homogenization temperatures in vein quartz hosted by the sandstone at this locality. Fluids in the inclusions are halite-saturated. The quartz crystals hosting the inclusions show strong inclusion zoning patterns, suggesting a series of pulses of fluid flow and quartz crystal growth.

Cum. miles	Miles from last Stop	
45.8	0.0	Continue north on Route 22
49.7	3.9	Intersection with Route 74 and Montcalm Street; turn left (west) onto Montcalm Street
50.1	4.3	Park near picnic area on right. UTM 627148E, 4856272N

Stop 6 – Potsdam Sandstone in Ticonderoga Village at LaChute River Falls

The upper stratified unit of the Potsdam Formation forms the water fall on the LaChute River in Ticonderoga Village. The hydroelectric plant excavations in the mid-1990's provided an opportunity to inspect the underlying carbonate-bearing lower cyclic unit. The overlying upper cyclic unit can be observed in scattered outcrops up slope to the north of the river exposures. Basal beds of the Galway Formation are present further up slope, near the road adjacent to Mt. Hope Cemetery. The sandstone slabs used in construction of the building at the hydroelectric site are from the upper stratified unit of the Potsdam. Note that some of these bear the trace fossil Protichnites.

End of trip – to return to Interstate 87, follow Rts. 22 and 74 north from last intersection; continue west on Route 74 to I87; to return to Lake George Village, continue on Montcalm Street though Ticonderoga Village; take Rt. 9N south to Lake George.