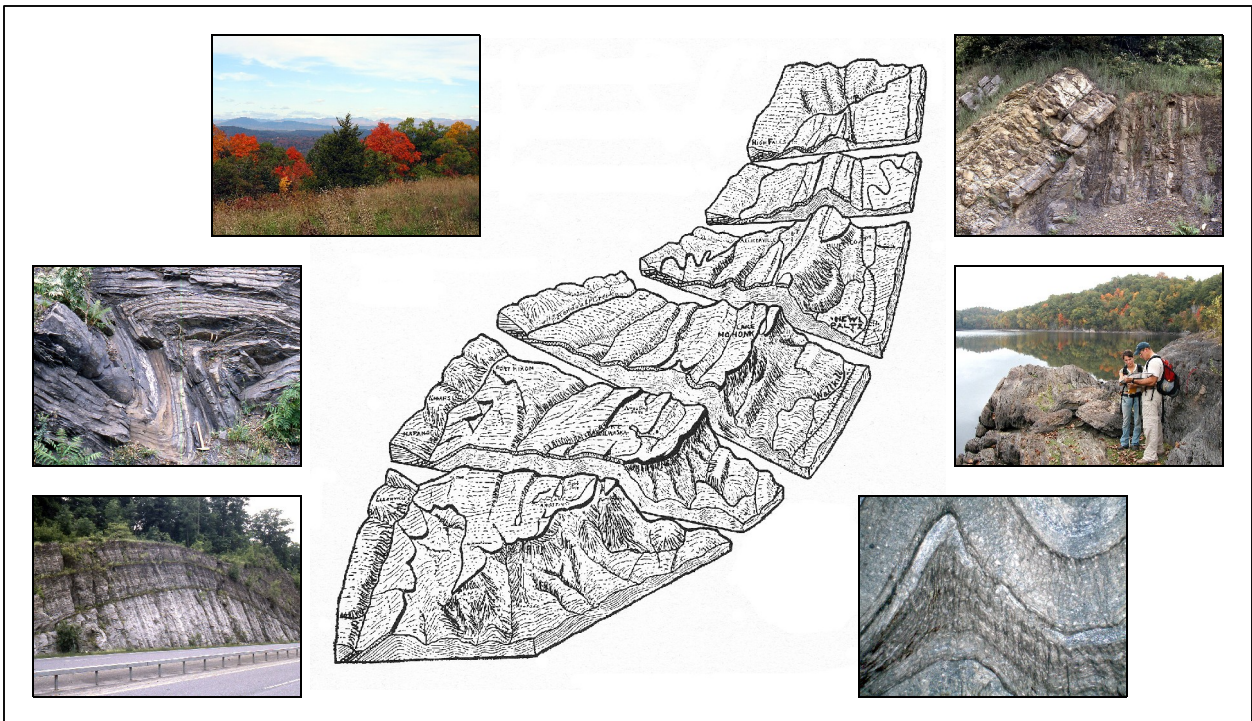


New York State Geological Association
81st Annual Meeting

Field Trip Guidebook



SUNY New Paltz
New Paltz, New York
September 25-27, 2009



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Hosted by
SUNY New Paltz
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STATE UNIVERSITY OF NEW YORK
NEW PALTZ

Edited by
Frederick W. Vollmer

Published by the
New York State Geological Society
www.nysga.net
ISBN 1061-8724

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Dedicated to
Russell H. Waines
1927 - 2009

Ph.D. Paleontology, UC Berkley 1965
Professor of Geology, SUNY New Paltz
1963 – 2006, Retired Emeritus
Editor of 1967 and 1987 NYSGA Guidebooks

Cover Images

Top: Sky Top cliff on the Shawangunk Ridge, folded Silurian Shawangunk quartz pebble conglomerate unconformably overlying Ordovician Martinburg shale along the Taconic unconformity.

Center: 'Stereogram of the Shawangunk Mountain', from N.H. Darton, 1894, Plate 12.

Clockwise from one o'clock:

1. Taconic unconformity, Catskill, NY, Silurian Rondout overlying Ordovician Austin Glen.
2. Mapping in the fold-thrust belt at Fourth Lake, former Williams Lake property.
3. Photomicrograph of crenulated slate, Dutchess County metamorphic sequence.
4. Anticline in Devonian Helderburg Group limestones, Kingston, NY.
5. Kink folds in Austin Glen associated with Taconic Allochthon emplacement.
6. View of the Catskill Mountains, Devonian Red Bed molasse sequence.

Photographs by F. W. Vollmer

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Structures of the Hudson-Valley Fold-Thrust Belt in the Appalachian Foreland of Eastern New York

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TRIP OVERVIEW

Exposures of a thin-skinned, foreland fold-thrust belt crop out west of the Hudson River in the central Hudson Valley of New York State (Figure 1a, b). This Hudson Valley fold-thrust belt (HVB), which involves Middle Ordovician through lower Middle Devonian strata, is generally less than 4 km wide. Thus, first-order structures (e.g., fault-related folds, detachment faults and duplexes) of the belt are small enough to be seen in their entirety within individual outcrops. The small size of the belt also makes it possible to examine along-strike variations in fold-thrust belt structural architecture that reflect variations in the thickness and mechanical properties of pre-deformational strata and/or in the amount of shortening (Marshak, 1983; 1986a; Burmeister and Marshak, 2003; Burmeister, 2005). This field trip will visit selected exposures of structures in the HVB to examine the structures of the belt at all scales. The stops illustrate the relationships between deformation style and stratigraphy.

In addition to examples of fold-thrust belt structural architecture, stops on the trip will also provide: (1) exposures of the Taconic angular unconformity (Rodgers, 1971); (2) examples of tectonic cleavage, veins, and slip fibers (Marshak and Engelder, 1985); (3) classic Lower Devonian North American faunas (Chadwick, 1944; Goldring, 1943); (4) examples of shallow-marine carbonate facies (Rickard, 1962; LaPorte, 1969); (5) an example of an orocline (Marshak and Tabor, 1989; Marshak, 2004); and (6) a context for considering continuing debates concerning the timing (Alleghanian vs. Acadian) of regional orogeny in the New England Appalachians, and the nature of the transition between the central Appalachians and the northern Appalachians (Marshak, 1986a; Marshak and Tabor, 1989; Marshak and Bosworth, 1991).

This trip proceeds from north to south along the belt. Highlights include the Route 23 roadcuts near Catskill, evidence for overprinting of deformation features near Kingston, and structural complexities of the Helderberg escarpment. This guidebook is a modification of a one that the lead authors have prepared for earlier field trips sponsored by the Geological Society of America and NYSGA.

GEOLOGICAL CONTEXT

Introduction

The Hudson Valley fold-thrust belt (HVB) is a narrow band of thrust faults and folds that verge generally westwards, toward the foreland of the Appalachian orogen. First-order folds range up to about 100 m in amplitude and 250 m in wavelength. The HVB is clearly a post-Taconic feature, because its structures involve the post-Taconic unconformity (an angular unconformity that truncates folds involving Middle Ordovician turbidites) and the Silurian through lower Middle Devonian strata above. Because of erosion, all that remains of the HVB is a 2 to 4 km-wide miniature valley and ridge province that lies between the Hudson River on the east, and the foothills of the Catskill Mountains on the west. Outliers of Silurian/Lower Devonian strata exhibiting HVB deformation crop out east of the Hudson River (at Mt. Ida and Becraft Mountain), indicating that the belt once extended further to the east into the region of the Taconic Mountains. But because the HVB structures strike parallel to Taconic structures, HVB deformation cannot be identified in regions where only Middle Ordovician and older strata are exposed.

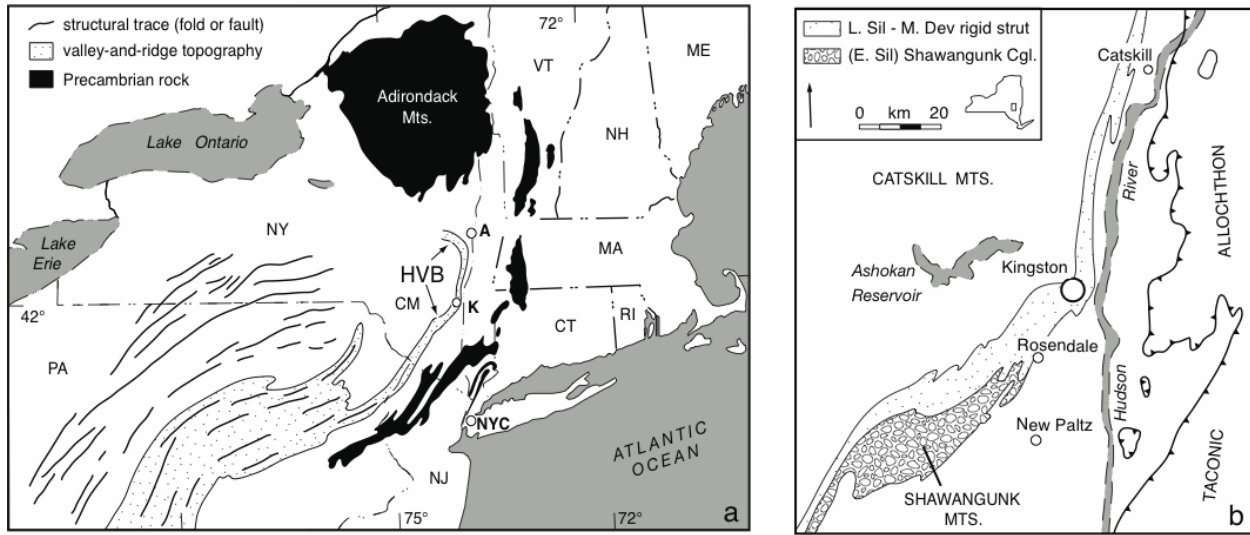


Figure 1. (a) A map of New York State showing the distribution of fold-thrust belt structures. The stippled belt represents the area of structurally controlled valley-and-ridge topography. Note that the broad Pennsylvania Valley and Ridge Province merges with the very narrow Hudson Valley fold-thrust belt, and that very gentle folds occur in the foreland of the Pennsylvania Valley and Ridge (in Pennsylvania and south-central New York), but not in the Catskill Mountains. HVB = Hudson Valley fold-thrust belt; K = Kingston; A = Albany. (b) Detail of the New Paltz to Catskill area, illustrating the pinch out of the Shawangunk Conglomerate, and the regional change in structural trends that takes place in Kingston.

In the region north of Kingston, structures of the HVB trend north-south to $N10^{\circ}E$, whereas in the region south of Kingston, HVB structures trend about $N30^{\circ}E$, and the belt becomes progressively wider as deformation propagates into the foreland (Figure 2). In New Jersey and Pennsylvania, the HVB merges with Pennsylvania Valley and Ridge Province. This map-view curve is one of the four major curves of the Appalachians (i.e., two salients and two recesses; Figure 3), and effectively marks the center of the syntaxes between the Northern Appalachians and the Central Appalachians. The change in trend within Kingston has been interpreted as an "intersection orocline" in which Alleghanian structures of the central Appalachian foreland overprint and rotate Acadian structures of the northern Appalachian foreland (Figure 4; Marshak and Tabor, 1989).

Structures of the HVB are significantly smaller than those of Pennsylvania Valley and Ridge or other foreland fold-thrust belts. This contrast is due to the thinness of the stratigraphic sequence involved in the HVB. In effect, the belt is a "fold-thrust belt in miniature". The scale of the HVB puzzled 19th century geologists-Davis (1882; 1883), who referred to the belt as the "Little Mountains," could not understand why strata in the HVB, a region in which elevations are less than 100 m, was so much more deformed than the Catskill Mountains, in which elevations reach 1200 m. Because of the proximity of the HVB to many colleges and universities, its outcrops are many student field trips and mapping projects every year. Several field guides to belt are available (e.g., Sanders, 1969; Marshak 1986b; 1989; 1990; Burmeister and Marshak, 2002).

Stratigraphy

The strata encountered on the field trip includes Middle Ordovician flysch as well as Late Silurian through lower Middle Devonian shallow marine strata (Figure 5a). At Rosendale, the Middle Ordovician Martinsburg Shale, a thick sequence of interbedded greywacke and shale, underlies the Taconic unconformity. Its correlative, the Austin Glen Formation, underlies the Taconic unconformity between Kingston and Catskill. This sequence represents an accumulation of turbidites deposited on the margin of North America just prior to collision with the offshore Taconic volcanic arc. Subsequent to the Taconic orogeny, which folded the Austin Glen and Martinsburg strata, a long period of erosion produced the post-Taconic unconformity. A sequence of Late Silurian clastic and carbonate strata were deposited on the unconformity. Near Rosendale, this sequence includes the Late Silurian Shawangunk Conglomer-

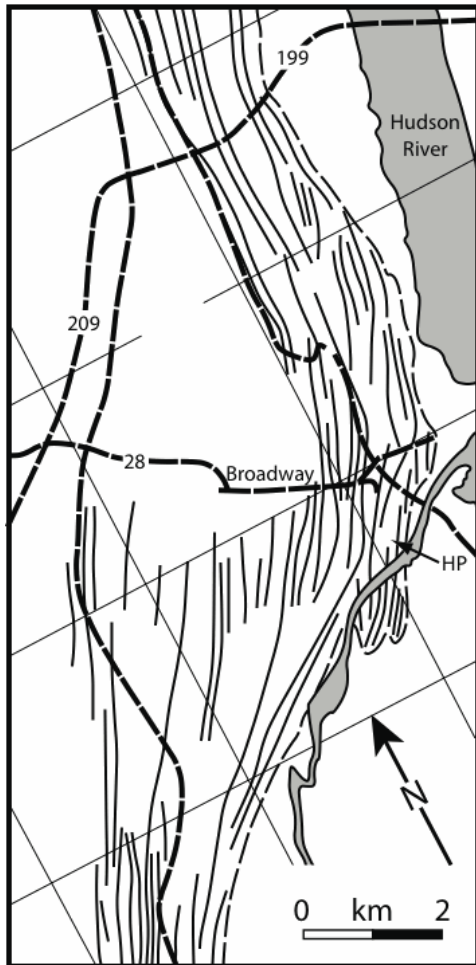


Figure 2. Sketch map of structural grain (fold and fault traces) in the Kingston area, showing the change in structural trend that occurs at Kingston (Marshak and Tabor, 1986). Note also that the belt gets wider south of Broadway Avenue. HP = Hasbrouck Park.

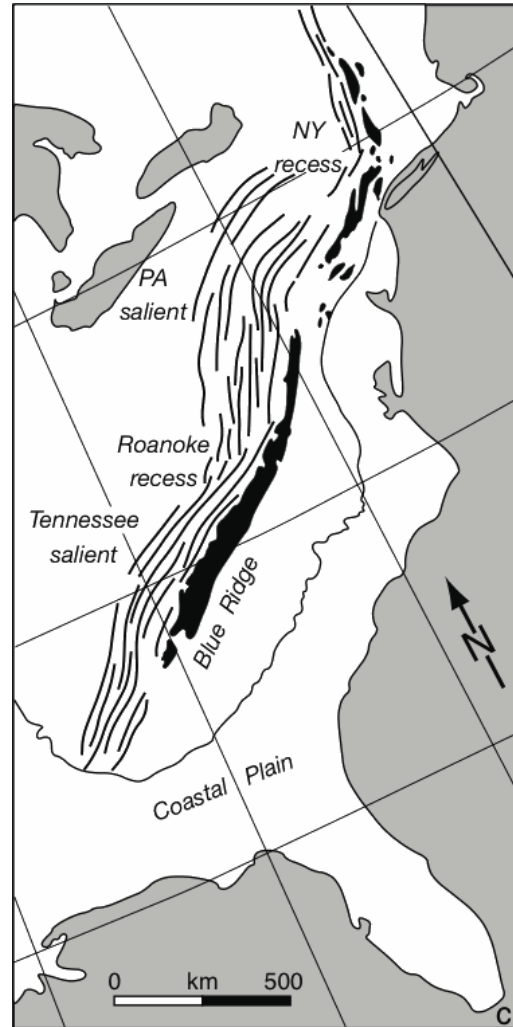


Figure 3. Sketch map of structural traces illustrating the salients and recesses of the Appalachian fold-thrust belt. The heavy lines are representative traces of folds and faults.

ate, the High Falls Shale, the Binnewater Sandstone, and the Rondout Formation (Waines and Hoar, 1967). Lower units in the Silurian section thin to the north and pinch out north of Rosendale, so that by the latitude of Kingston, only the Rondout Formation remains (Figure 5b). The Rondout Formation continues to thin northwards to Catskill, where it is represented by only 1 to 2 m of sandy, dolomitic limestone.

The Devonian Helderberg and Tristates Groups overlie the Rondout Formation. These groups include deposits indicative of successive transgressions of a shallow sea (Wanless, 1921; Waines and Hoar, 1967; Laporte, 1969; Sanders,

1969; Rodgers, 1971; Marshak, 1986a; Epstein and Lytle, 1987; Marshak and Tabor, 1989; Marshak, 1990). The only significant non-carbonate unit in this sequence is the Esopus Shale. Above the Tristates Group lies the Onondaga Limestone, which is the youngest carbonate unit to be deposited prior to the deposition of the Catskill clastic wedge. Deformation features characteristic of the fold-thrust belt deformation in the Hudson Valley are visible in the Bakoven Shale and Mt. Marion Formation, units of the Hamilton Group that directly overlie the Onondaga Limestone, but cannot be found in younger units (Murphy and others, 1980).

Regional Structural Architecture of the HVB

Model cross sections of the HVB suggest that shortening in the fold-thrust belt occurred above regional-scale detachment faults (Marshak, 1986a; Burmeister and Marshak, 2002). From Kingston to the north, there appear to be four detachment horizons (Figure 6). Area balancing of cross sections suggests that the lowest of these, called the Austin Glen detachment, lies at a depth of about 500 m below ground level. This detachment may represent a reactivated Taconic thrust. Displacement on it resulted in folding of the post-Taconic unconformity, and the development of folds with amplitudes of tens of meters that are cored by Middle Ordovician strata. Ramps rising from this detachment locally place Ordovician strata over Silurian strata; examples cropped out in quarries north of Kingston. The next higher detachment, the Rondout detachment, lies just above the post-Taconic unconformity. Where exposed, this detachment is a zone of west-verging mesoscopic folding, intense cleavage development, and local duplexing in the Rondout Formation. Ramps rising from this detachment cut across Lower Devonian strata. Detachments, with associated deformation also occur at the base of the Esopus Shale, and in the Bakoven Shale. From Kingston to the south, there appear to be additional detachments below the Austin Glen detachment, and the Rondout detachment does not appear to exist. Deformation in the fold-thrust belt south of Kingston appears to be associated with two detachment faults within Ordovician strata. Thus, detachments in the HVB appear to ramp up section from east to west, and from south to north.

Deformation associated with the fold-thrust belt along the western margin of the Hudson Valley dies out westward. The occurrence of cleavage duplexes in the Bakoven Shale (Nickelsen, 1986) and of spaced cleavage in the Mount Marion Formation suggests that west-directed displacement associated with the HVB occurs at least a few kilometers west of the westernmost fold of the HVB (Marshak and Bosworth, 1991).

Stratigraphic units in the Hudson Valley can be categorized in terms of their mechanical character (Figure 3b). Specifically, the Late Silurian through Middle Devonian clastic and carbonate sedimentary rocks can be considered to be a mechanically rigid "strut" (Burmeister and Marshak, 2002; Burmeister and Marshak, 2003; Burmeister, et al., 2003; Burmeister, 2005). The Siluro-Devonian strut in the central Hudson Valley is roughly 150 m thick near Rosendale. (Of note, its along-strike equivalent in the Pennsylvania Valley and Ridge Province to the south is more than 1000 m thick.) This strut thins northwards between Rosendale and Catskill. This strut is sandwiched between thick, relatively weak, shale sequences: the Middle Ordovician Martinsburg and Austin Glen Formations below, and the Middle Devonian Esopus Shale above. This distinct mechanical stratigraphy controls the dimensions of structures and the positions of detachments - as units thin, fold amplitude and thrust spa-

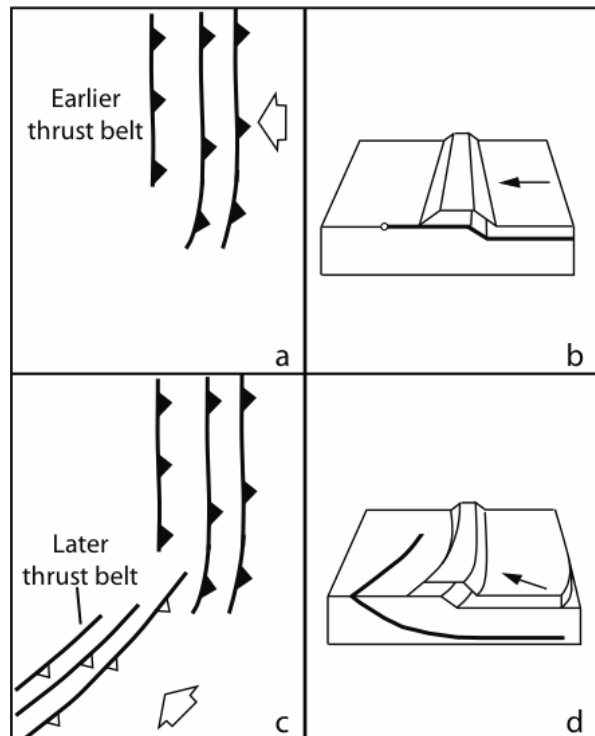


Figure 4. Block diagrams showing the concept of an intersection orocline. The NE-SW-trending thrust belt is younger than the N-S-trending thrust belt. Displacement on a younger thrust bends the structures in its hanging wall (Marshak and Tabor, 1986).

Middle Devonian	Hamilton Gp.	Dmm	Mount Marion Formation (siltstone; sandstone)
		Dbk	Bakoven Shale (black, organic-rich shale)
		Don	Onondaga Limestone (light grey limestone with chert)
Lower Devonian	Tristates Group	Dsc	Schoharie Formation (argillaceous limestone)
		De	Esopus Shale (grey shale and siltstone)
	Helderberg Group	Dg	Glenerie Formation (silicified limestone)
		Dpe	Port Ewen Formation (argillaceous limestone)
		Da	Alsen Formation (argillaceous limestone)
		Db	Becraft Limestone (pinkish grey crinoidal lime grainstone)
		Dns	New Scotland Formation (argillaceous limestone)
		Dk	Kalkberg Formation (argillaceous limestone; black chert layers)
		Dc	Coeymans Limestone (grey fossiliferous limestone)
		Dm	Manlius Limestone (grey fossiliferous and laminated micrite)
Upper Silurian		Sr	Rondout Formation (sandy, dolomitic limestone; natural cement)
		Sb	Binnewater Sandstone (tan sandstone)
		Sh	High Falls Shale (red shale/siltstone)
Middle Ordovician		Oag	Austin Glen / Normanskill (turbidites)

~ 30 m

— Post-Taconic unconformity —

Figure 5. (continues)

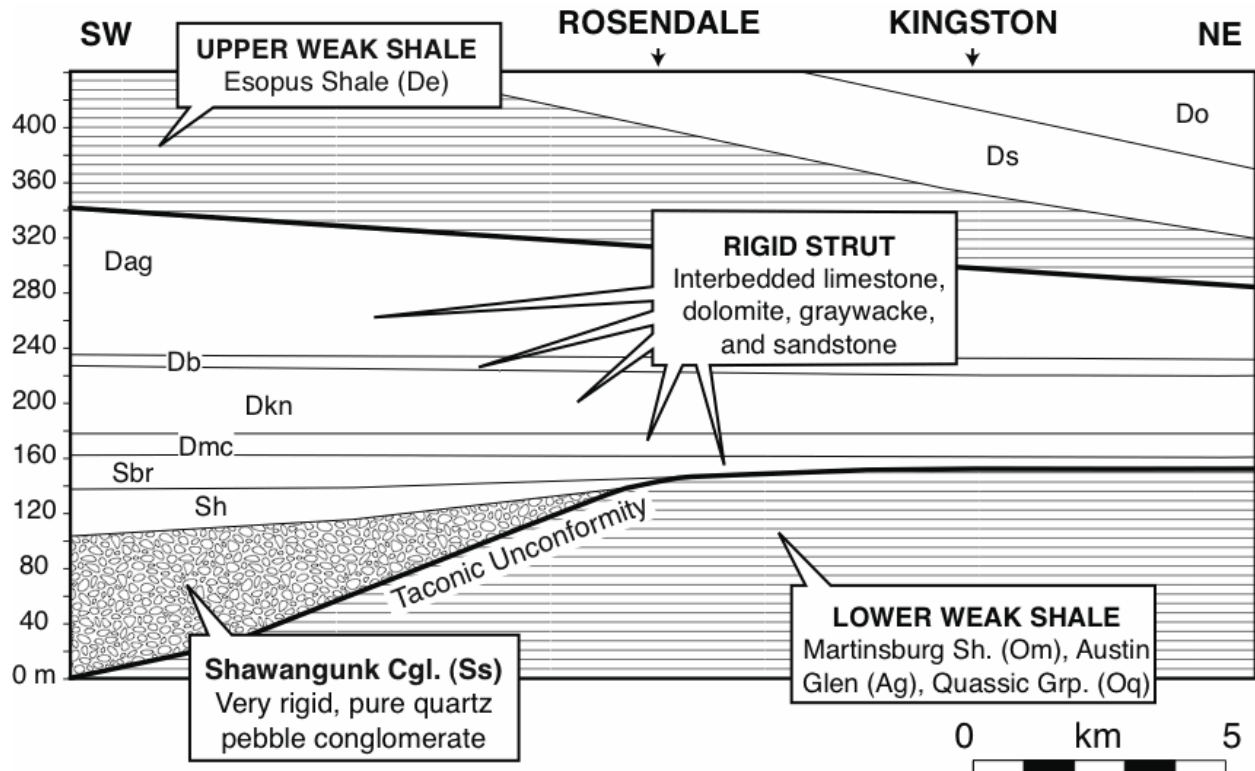


Figure 5. (continued) (a) Composite stratigraphic column of the Hudson Valley fold-thrust belt. Thicknesses are approximate. (b) Simplified north-south trending fence diagram, illustrating how Silurian strata thicken to the south starting at a latitude just north of Kingston (Burmeister, 2006).

cing decreases. Accomodation between different zones of the fold-thrust belt appears to take place by lateral ramping.

The age of deformation in the Hudson Valley remains enigmatic. It is possible that the belt is Acadian (Devonian), because of the location of the belt between the Acadian foreland basin, represented by redbeds of the Catskill Mountains, and the Acadian orogen of New England. However, the belt could also be Alleghanian (late Paleozoic), because of an apparent continuity with structures of the central Appalachians and the proximity of the belt to the foreland of a region that was remobilized during the Alleghanian orogen. Deformation age cannot be stratigraphically constrained because the belt involves Lower Devonian strata, but dies out to the east of the first exposures of Acadian foreland strata.

LIST OF STOPS

Optional Stop A - Abandoned Quarry near Fuerra Bush. Exposure of a duplex along the Rondout Detachment, and a large Ramp anticline. This stop provides excellent exposures of an impressive 10 m-high duplex involving the Rondout Formation and the Manlius Limestone along the Rondout Detachment. Within the duplex, there are numerous ramps spaced at 0.5 to 2 m. The duplex has a flat roof. The quarry also exposes of a first-order ramp anticline involving the Manlius Limestone, Coeymans Limestone, and Kalkberg Formation. The quarry has been cut into the Helderberg Escarpment, a 10 to 50 m-high cliff that marks the eastern limit of Siluro-Devonian strata in the HVB west of the Hudson River.

Stop 1 - Exposures along Route 23 Near Catskill: A cross section of the entire HVB. Stop 1 includes several roadcut exposures along State Route 23, approximately 2 km northwest of Catskill. The outcrops are numbered in succession from west to east (Figure 7); those containing the letter N border the north side of the road, and those containing the letter S border the south side. For logistic reasons, we start the trip at the east end of the fold-thrust belt and work our way westwards, toward the foreland. These exposures show the following:

Roadcut 1A exposes the post-Taconic angular unconformity and the Rondout detachment. The unconformity dips moderately westward, parallel to the strata above, and cuts across steeply dipping beds of Austin Glen Formation. Displacement on the detachment resulted in formation of west-verging (i.e., down-dip verging, in this outcrop) folds, and in cleavage that has rotated counter-clockwise, relative to bedding (Marshak, 1986a, b). A homoclinally dipping sequence of the lower Helderberg Group (Manlius, Coemans, and Kalkberg Formations) lies over the Rondout Formation. The Manlius contains small black ostracods, the top of the Manlius is a layer of thinly laminated micrite, the Coeymans contains the white, beaked brachiopod *Gypidula coeymanensis*, and the base of the Kalkberg is marked by a distinctive black chert layer.

Here, as throughout the Hudson Valley fold-thrust belt, development of cleavage is lithologically controlled; cleavage occurs primarily in rocks containing greater than 10 percent clay (Marshak, 1983). Also, slip fibers develop on bedding planes due to both detachment faulting and flexural slip. On the west limb of folds, slip due to flexural slip can be distinguished from that due to detachment faulting by the sense of slip indicated by imbrication of fiber sheets and by shear rotation of cleavage adjacent to slip surfaces. Flexural slip displacement is always "top-up." Slip on a west-verging, but now west dipping detachment, is "top-down" in its present (post-folding orientation).

Outcrop 1B provides another exposure of the lower Helderberg Group in the western limb of the Tollgate Syncline, a first-order fold that underlies State Route 23B between 1A and 1B. In this outcrop, the section was thickened as a result of movement on two well-exposed, out-of-the-syncline forethrusts (see Dahlstrom, 1970). The lower fault brings the Manlius Formation over the Kalkberg Formation.

Outcrops 1CN and 1CS provide additional exposures of the Taconic Unconformity and of the Helderberg Group

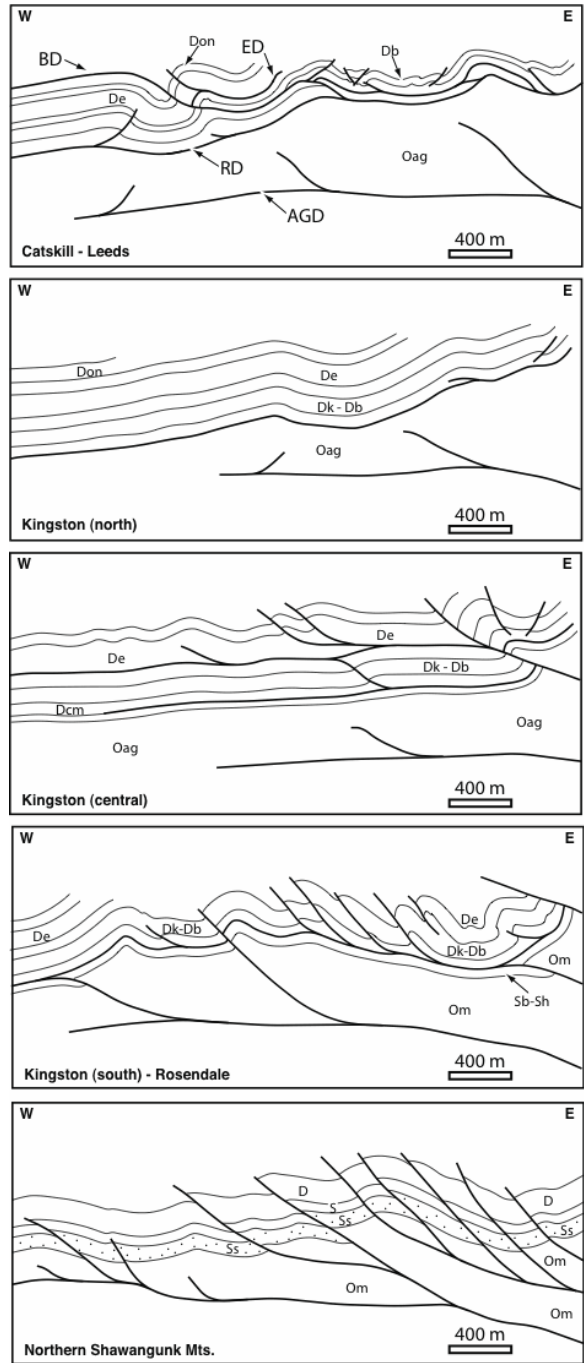


Figure 6. Model cross sections of the Hudson Valley fold-thrust belt at different latitudes (north at the top, to south at the bottom). Oag = Austin Glen; Om = Martinsburg; Sr = Rondout; Sb-Sh = Binnewater and High Falls; Dm-Dc = Manlius and Coeymans; Dk- Db = Kalkberg, New Scotland, and Becraft; De = Esopus; Don = Onondoga (Marshak and Tabor, 1986; Burmeister, 2006).

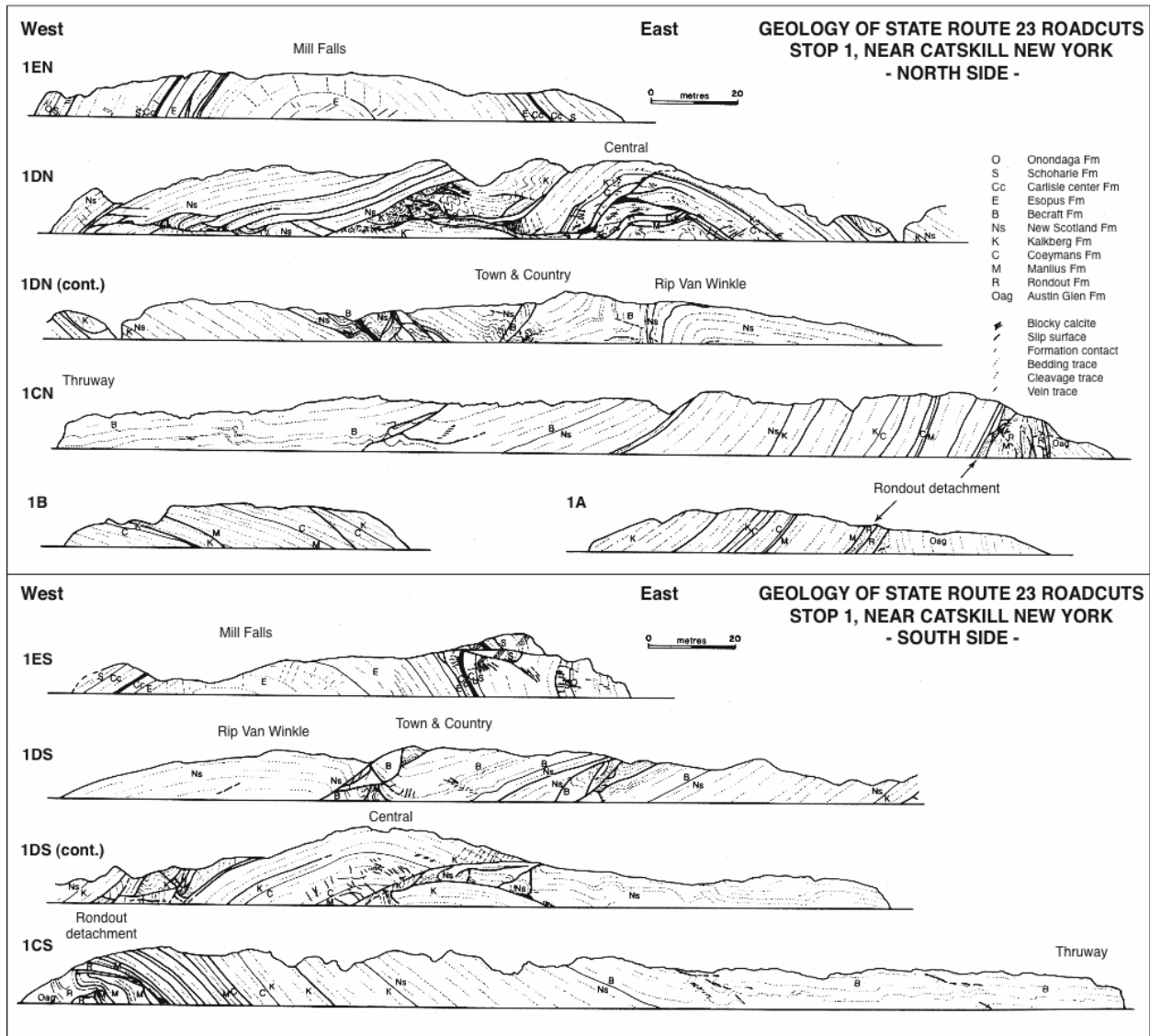


Figure 7. Sketches of the structures exposed in the Route 23 roadcuts, west of Catskill (Marshak, 1984). The major folds are named.

(through the Becraft Formation). Of particular note in these outcrops are the complex faults and folds in the Rondout and lower Manlius Formations, manifestations of movement on the Rondout detachment. Many bedding-plane slip surfaces, which developed during flexural-slip folding and are coated with sheets of calcite slip fibers, occur in outcrops 1CN and S3. In the Kalkberg Formation, some of these slip surfaces are bounded by zones of nearly slaty cleavage. At the northwest end of outcrop 1DN, numerous mesoscopic folds, as well as two back-thrusts, occur within the Becraft Limestone.

Outcrops 1DN and 1DS are the most spectacular of the State Route 23 outcrops. Together, these exposures (which include rocks of the Manlius Formation through Becraft Limestone) display, from southeast to northwest, ramp faults with hanging wall anticlines (Rip van Winkle anticline), out-of-the-syncline forethrusts and backthrusts, folded ramps and flats (in the Central anticline), and zones of tectonic cleavage intensification (on the northwest limb of the Central anticline). Of particular note are the examples of fault bends (Suppe, 1983) at which bedding-parallel

flats join cross-strata ramps. The Central anticline appears to be composed of a stack of fault-bounded horses (see Boyer and Elliott, 1982), one of which is internally deformed throughout by mesoscopic folds. Structures of outcrops 1DN and 1DS do not directly correlate across the highway, illustrating how rapidly structural geometry can change along strike in the HVB. The contrast in structural geometry between outcrops on opposite sides of the road may reflect the occurrence of a lateral ramp in the interval that was excavated during construction of the highway. We can also see variations in strain magnitude, indicated by cleavage intensity, as well as lithologic control on the degree of cleavage development. Studies of these rocks provide interesting constraints on the volume-constant vs. volume-loss models of cleavage development (Bhagat and Marshak, 1990).

Outcrops 1EN and 1ES expose the Esopus and Schoharie Formations, and the base of the Onondaga Limestone. These units are arched around the Mill Falls anticline. Nearly slaty cleavage occurs within the lower Esopus Shale. The upper few meters of the Esopus Shale are composed of beds of finely laminated mudstone and siltstone that have been crinkled into tiny folds. A sub-horizontal fault is present at the top of outcrop 1DS. Time permitting, we will visit the deformation front of the fold-thrust belt as it is exposed along Catskill Creek in the village of Leeds, approximately 1 km north of State Route 23.

Stop 2 - Roadcut along Route 23A, east of the Thruway Bridge: Exposure of the Esopus detachment. Stop 2 exposes disharmonic folds that formed in association with movement on a detachment fault between the Esopus and Glenerie Formations. This outcrop inspires considerable debate about the origin of the deformation. Is the structure tectonic, or is it a consequence of slumping that was penecontemporaneous with deposition? The association of fractures, cleavages, and shear zones with the folds suggests that the deformation is tectonic. Deformation likely occurred at this horizon because of the viscosity contrast between the Esopus and Glenerie Formations. The lowermost Esopus Shale is composed of alternating beds of siltstone and shale, and such a sequence is susceptible to buckling. The overlying Esopus Shale is more homogeneous and contains closely spaced to slaty cleavage. The detachment at the contact may reflect differential shortening between the Esopus and Glenerie Formations.

Optional Stop B - Roadcut on Route 209: Cleavage duplexes in the Bakoven Member of the Union Springs Formation. This stop exposes the Bakoven and Stony Hollow Members of the Union Springs Formation, the basal unit of the Hamilton Group, in the steep escarpment and road cut along City View Terrace, beneath the Skytop Motel. Mesoscopic-scale duplexes and shear zones in the upper 10 m of the Bakoven Member provides an opportunit-

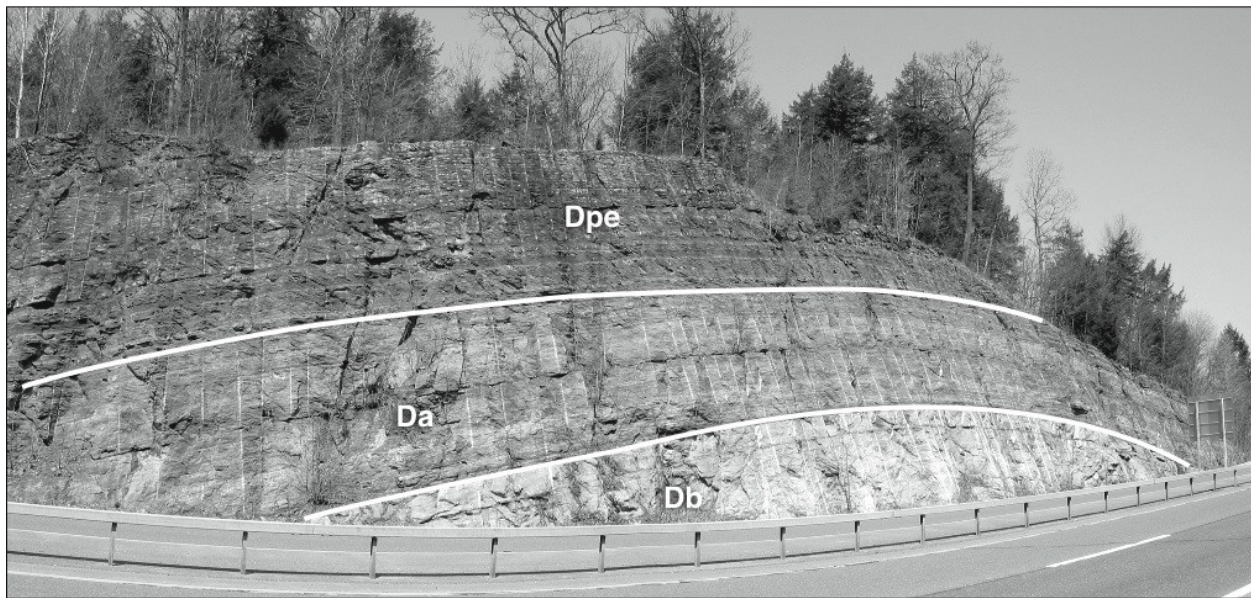


Figure 8. Photograph of roadcut exposure showing a first-order anticline involving Becraft Limestone (Db), Alsen Formation (Da), and Port Ewen Formation (Dpe) along the north side of State Route 209/199, north of Kingston.

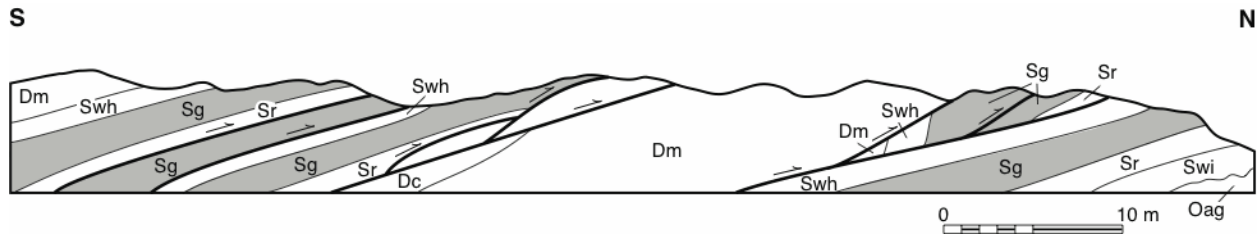


Figure 9. Schematic cross section with slight vertical exaggeration of road cut exposure along west side of State Route 32 just-south of intersection with State Route 199. Note duplication of Silurian and Devonian strata in series of imbricate thrust sheets and horses. Oag = Austin Glen Formation; Rondout Formation [Swi = Wilbur Member; Sr = Rosendale Member; Sg = Glasco Member; Swh = Whiteport Member (Swh)], Dm = Manlius Limestone; and Dc = Coeymans Formation. Figure redrawn after D.B. McEachran (1985 - UIUC M.S. thesis).

ity to examine deformation associated with the stratigraphically highest detachment of the HVB (Bosworth, 1984; Nickelsen, 1986; Ver Straeten and Brett, 1995). Nickelsen referred to the structures in this outcrop as "cleavage duplexes," implying that the planar laminations within the shear zone is slaty cleavage not bedding. An alternative view is that the laminations are bedding planes that have rotated almost 90°, due to shear. The kinematics of these zones remains unclear.

Stop 3 - Roadcuts along State Route 199/209, just north of Kingston: Exposures of large anticlines and synclines. Stop 3 exposes large, slightly asymmetric open folds with no visible thrust faulting or mesoscopic folding. Five major, map-scale fold hinges cross State Route 209 between Routes 9W and 32 (McEachran, 1985; Marshak and Tabor, 1989; Marshak, 1990). The most prominent of the folds at Stop 3 is a large anticline involving the upper Becraft Limestone, the Alsen Formation, and the Port Ewen Formation (Figure 8). Note the differential development of cleavage in these units. The Becraft Limestone contains little to no cleavage, whereas a strong, southeast-dipping cleavage is developed in the Alsen and Port Ewen Formations. These folds are characteristic of structural styles in the northern arm of the Kingston Orocline, in that they trend roughly N015°E and lack structural complexity (Marshak and Tabor, 1989).

Stop 4 - Roadcuts along Route 32 (Flatbush Ave) north of Kingston: Lateral ramping in the Rondout Formation. Stop 4 is along the west side of State Route 32 (Flatbush Ave.), 0.3 miles south of US 199 overpass (Figure 9). Stop 4 exposes four major thrust faults duplicating the Rondout and Manlius Formations in a series of imbricate thrust sheets along the west side of State Route 32 (Figure 7; Waines and Hoar, 1967; McEachran, 1985; Marshak, 1990). Thrust faults in this outcrop strike roughly N040E. Calcite slip fibers on the fault surfaces suggest a transport direction of N060E-N070E. In places, these faults have a flat-on-flat geometry (i.e., thrust faults are bedding parallel in both the hanging wall and footwall), suggesting large displacements. These thrusts occasionally appear to cut down-section due to the obliquity of the road cut face to the orientation of the structures and the transport direction (hanging walls are thrust westward into the outcrop). These faults are untraceable west of the road cut, but may ramp laterally up-section and die out to the north.

Optional Stop C - Route 9W Roadcut in Onondaga Limestone, north of Kingston: Cross-cutting cleavage. At this exposure, in the center of the Kingstone orocline, a bedding surface of Onondagaga limestone at the top of a roadcut on the east side of the Highway, appears to contain two distinct cleavage trends (Figure 10). This relationship is evidence that two distinct segments of the fold-thrust belt cross cut in the Kingston area.

Stop 5 - Hasbrouck Park, Kingston: The foreland edge of an antiformal stack along the Helderberg Escarpment. Stop 5 at Hasbrouck Park provides a cross-sectional view through the leading edge of a duplex structure involving horses from the Rondout through Kalkberg Formations (Figure 11; Marshak and Tabor, 1989; Marshak, 1990). The basal thrust fault in this duplex, the Hasbrouck thrust, cuts laterally up-section along the sloping footpath between the parking lot and Stop 5. The Hasbrouck thrust fault reappears north of Delaware Avenue, where it places the Manlius Formation over the Alsen Formation. A tear fault with a trace roughly coincident with Delaware Avenue

may extend to the west of the lateral ramp. The complexity of the structural relationships at Hasbrouck Park is characteristic of the geology along the Helderberg Escarpment near Kingston.

Optional Stop D - Exit Ramp from 9W to Delaware Avenue: Fly Mountain Thrust. At this locality, we see an exposure of a thrust fault that places Esopus Shale against Onondaga Limestone. The fault surface is decorated with calcite fiber slip lineations. This fault changes trend, south of Kingston.

Optional Stop E - Callanan Quarry: Backthrusts along the Helderberg Escarpment. The northwest wall of the quarry (the high wall below the railroad line) is a spectacular exposure of complex structural relationships, including backthrusts, out-of-the-syncline faults, and duplex structures that are characteristic of the Helderberg Escarpment near Kingston (Fig 12). The Upper Becraft Limestone provides a distinctive marker horizon.

Optional Stop F - Rosendale Landfill: Exposure of Ordovician strata in the core of the Hickory Bush anticline. Outcrop at this stop exposes a complete, east-dipping sequence of Middle Ordovician through Middle Devonian strata in the northern wall of the Rosendale Landfill and Recycling Center. Here, the Martinsburg Shale is strongly deformed by complex brittle faulting. The Shawangunk Conglomerate is little more than a 5 to 10 cm thick lag deposit of characteristic milky white quartz pebbles overlying the Taconic unconformity. The Shawangunk Conglomerate, a silica cemented quartz-pebble conglomerate, thickens dramatically to the south, where it becomes a mechanically rigid strut in the pre-deformational stratigraphic section involved in fold-thrust belt deformation in the northernmost Valley and Ridge Province. Overlying the Shawangunk Conglomerate are the High Falls Shale and Binnewater Sandstone. The Rosendale and Whiteport Members of the Rondout Formation are quarried at this location, leaving the Glasco Member, which contains beautiful Halycities chain corals. The Rondout Formation is overlain by the Manlius Limestone, which is truncated by a fault.

The east-dipping strata exposed here form the eastern limb of the Hickory Bush anticline. The involvement of Martinsburg Shale in the fold core suggests this structure developed as the result of slip along a detachment at depth in the underlying Ordovician strata. The scale of the Hickory Bush anticline suggests that the underlying thrust is a master fault in the fold-thrust belt at this latitude and ramps directly from the lower of two detachment horizons. A smaller thrust fault exposed at the eastern end of the exposure is the westernmost fault in a complex imbricate fan of thrusts that ramp out of the Rondout Formation and cut through Hickory Bush Hill (the large hill southeast of the landfill).

Optional Stop G - Rail Trail, near Fourth Lake: Limb of Hickory Bush anticline. Walk southwest along the rail trail from the parking lot along Hickory Bush Road for roughly 350 m until you encounter a cut exposing northwest dipping strata. The exposed sequence includes High Falls, Binnewater, Rondout, and Manlius Formations (Figure 13). Along the northwest side of the trail, the High Falls Shale is difficult to distinguish, but it is overlain by a complete thickness of the Binnewater Sandstone. The Rosendale and Whiteport members of the Rondout Formation were quarried at this location. Climb onto the embankment on the south side of the rail trail using the small path through the trees located just north of the northernmost quarry opening. The embankment was



Figure 10. Photo of a bedding surface in the Onondaga Limestone, in north Kingston, illustrating two different cleavage orientations. The compass points due north. In the intersection orocline model, the earlier cleavage trends about N10°E, and the later cleavage dates about N30°E. The earlier cleavage is better developed, and may have continued to be an active locus of pressure solution even after the second cleavage starts to form, thus explaining ambiguous cross-cutting relationships.

once a tramway that serviced the cement quarries in this area. Watch your step and be careful to stay away from the large openings atop the cement kilns that are hidden in the underbrush.

Optional Stop H - Former Williams Lake Hotel property: Exposures of an asymmetric anticline in an abandoned room-and-pillar mine. Abandoned cement mines at Stop G provide unique exposures of a series of large, asymmetric folds in the High Falls, Binnewater, Rondout, and Manlius Formations on the property of the former Williams Lake Resort. Watch your step—abandoned cement kilns and sunken shed foundations are scattered along this tramway and are often difficult to see. The floor of the "Chop Shop" mine in the Rosendale Member of the Rondout Formation is formed by the contact with the underlying Binnewater Sandstone, whereas the contact with the overlying Glasco Member forms the ceiling. Because of a recent change in ownership, it is not clear if permission can be obtained to visit this property.

Stop 6 - Snyder Estate: Historic mine in the footwall of the laterally ramping Century thrust fault. This outcrop is an extensively mined exposure of the Rondout Formation in the footwall of the Century thrust fault. As units in the mechanically rigid Siluro-Devonian strata thin along strike to the north, the Century thrust ramps laterally up section. South of Rondout Creek, the Century thrust juxtaposes Ordovician and Late Silurian strata, but ramps up section into Middle Devonian strata north of Rondout Creek. The abandoned mine at this stop is one of the oldest in the Rosendale natural cement region and is an example of a classic, two-tiered Rosendale cement mine. Dolostone

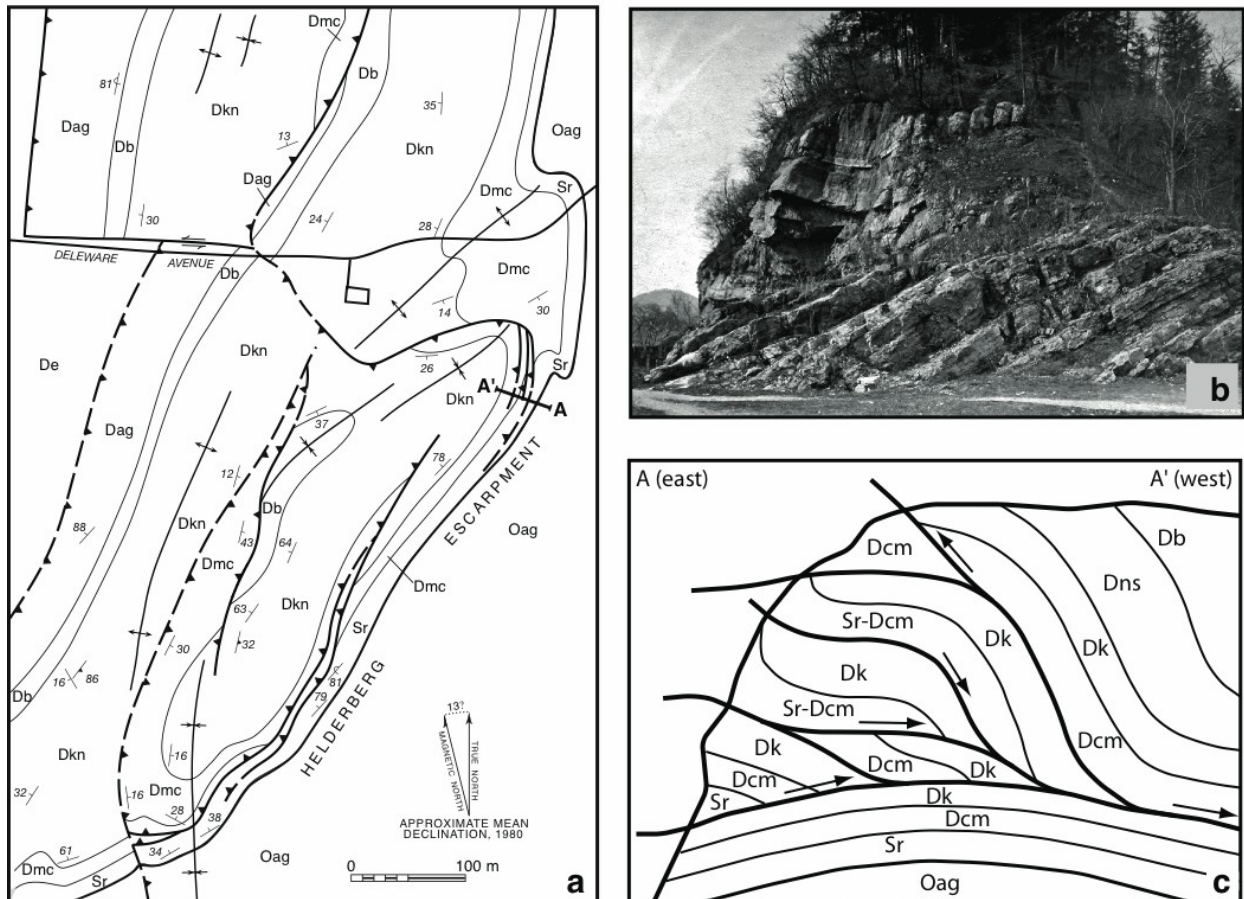


Figure 11. (a) Simplified geologic map of Hasbrouck Park (after Marshak and Tabor, 1989). Note complex structural relationships along eastern map margin. (b) Photo by G. van Ingen in Cairnes (1920) shows the NE corner of the escarpment in the park before regrowth of the forest. (c) A schematic cross-section above (not to scale), showing an imbricate thrust interpretation along Line AA'.

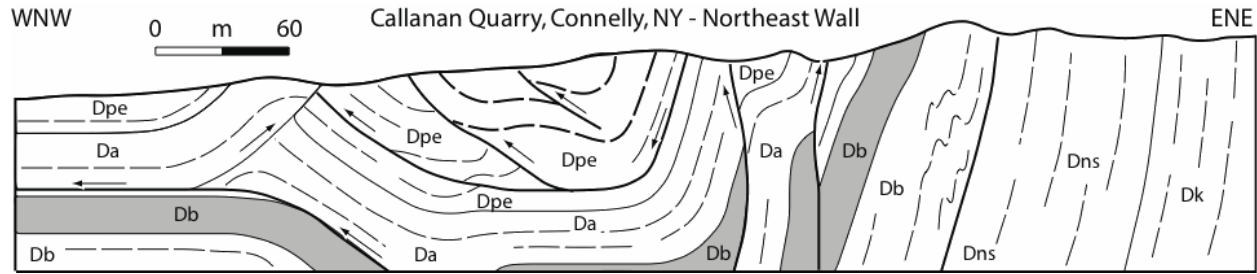


Figure 12. Sketch of the northeast highwall in the Callanan quarry in Connelly, south of Esopus Creek. Note the backthrusts and the steep dips. (From Marshak and Tabor, 1986, modified from an unpublished version by P. Kwizinski, UIUC M.S. student.)

was extracted for the production of natural cement from the Rosendale and Whiteport members of the Rondout Formation, which were historically referred to as the upper and lower cements, respectively. The Glasco Member, historically referred to as the middle ledge, is left un-quarried between the dolomitic layers. There are many remnants of the cement industry on the historic Snyder Estate property, including the ruins of cement kilns, a screening house, and many tramway roadbeds. Use caution when exploring these ruins and be careful to stay away from the large openings atop the cement kilns.

Optional Stop I - Rondout Creek, Lawrenceville. Optional Stop I can be reached by continuing west 0.7 miles on State Route 213 from the entrance to the Snyder Estate. Pull into the public parking lot along the left (south) side of the road for longer stays at this location. This stop affords a good view of the Kalkberg and New Scotland Formations in the Lawrenceville anticline. This anticline is the westernmost major, map-scale fold exposed in the Rosendale region and appears in numerous historical photographs taken of the Delaware & Hudson Canal in this region. This exposure is often obscured by vegetation during the spring and summer months.

Optional Stop J - Route 213, High Falls: Exposure of a bedding-parallel slip surface in Shawangunk Conglomerate. Optional Stop J provides an excellent exposure of the Shawangunk Conglomerate. The contact with the overlying High Falls Shale lies near the top of the road cuts. Rocks exposed at the eastern end of this road cuts lie in the footwall of a thrust fault that strikes roughly perpendicular to the trend of State Route 213. The hanging wall of this fault is exposed in the next set of road cuts east of R8 along State Route 213. A distinct, bedding-parallel slip surface exposed along the north side of State Route 213 at this stop contains a cleavage duplex composed of powdered quartz. This duplex suggests a top-to-the-west sense of shear. However, the lack of piercing points prevents a quantification of the amount of slip that has occurred on this sur-

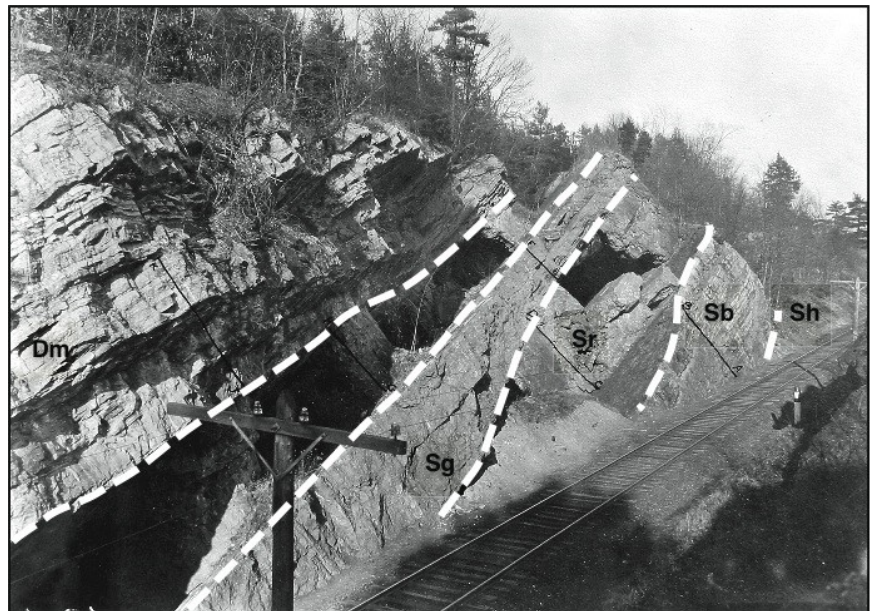


Figure 13. Annotated photograph of a cut along the rail trail, near Fourth Lake, showing the west limb of the Hickory Bush anticline, and the remnants of abandoned roof and pillar cement quarries. The photograph was taken by G. van Ingen, before reforestation, and was reprinted by Osborn (1921).



Figure 14. Panoramic photograph along the north bank of Rondout Creek and High Falls Park (Central Hudson power plant) in High Falls, NY. This is the westernmost fold at this latitude.

face. For this reason, it is unclear if the slip on this surface was the result of flexural-slip folding (suggesting little lateral displacement between bedding layers) or if it is a segment of a thrust fault exhibiting a flat-on-flat geometry (suggesting large lateral displacement).

Stop 7 - Waterfall at High Falls: Exposure of an east-verging asymmetric anticline at High Falls. Stop 7 can be reached by continuing west on State Route 213 for 0.4 miles before turning right (north) into High Falls Park/Central Hudson Power Station. Follow paths north down the hill to the bank of the Rondout Creek. Stop 7 exposes an asymmetric anticline in the Binnewater Sandstone and High Falls Shale along the Rondout Creek, northwest of High Falls (Figure 14). To the southwest of High Falls, Rondout Creek flows north, roughly parallel to the strike of gently west-dipping strata. At High Falls, Rondout Creek bends sharply to the east and begins cutting perpendicular to strike, down-section towards Rosendale. The highest waterfall is over an outcrop of the Rondout Formation. Downstream, smaller waterfalls occur in the Binnewater Sandstone and High Falls Shale, and are followed by minor rapids in shallowly east-dipping Shawangunk Conglomerate. The asymmetric anticline in the Binnewater Sandstone and High Falls shale at Stop 7 is the westernmost mesoscopic structure in the fold-thrust belt near Rosendale, and is best exposed along the north bank of the creek. Note that the axial surface of the fold dips to the west, which is a vergence opposite to the regional trend. The geometry and scale of this fold suggest that it is a fault-propagation fold above a blind thrust fault. The fault underlying this structure is most likely a west-dipping back thrust associated with detachment horizons in the Ordovician strata at depth.

ROADLOG

Optional Stop A - Abandoned quarry near Fuerra Bush (Lat 42°33'27.37"N; Long 73°51'55.04"W). This quarry is now owned by the South Bethlehem Police, and is used for firearms training. It cannot be entered without permission. The high walls are very unstable and dangerous, and should only be observed from a distance. The best exposure of the duplex is at the base of the north high wall. To reach the quarry, take the Selkirk Exit (Exit 22) off the NYS Thruway. After the tollgates, turn right onto Route 144. Proceed southwest on 144 for about 0.4 miles, then turn right and head west on Rte 396 through Selkirk, to South Bethlehem (about 2 miles). Continue on Rte 396 for another 0.5 miles west of South Bethlehem to the junction with Old Quarry Road (County Rte. 102). Turn right and head north on Old Quarry Road for about 1.5 miles. There will be a highway department shed on the left, and a parking lot. The quarry is up the hill behind the shed, and can be reached by walking about 10 minutes along a trail.

Stop 1 - Route 23 Roadcuts, northwest of Catskill (Lat 42°14'20.76"N, Long 73°53'9.41"W to Lat 42°14'54.51"N, Long 73°53'46.13"W). To reach the Roadcuts, take the Catskill Exit (Exit 21) off the NYS Thruway. At the end of the tollgate access road, turn left (southeast) onto State Route 23B heading toward Jefferson Heights and Catskill. Continue on State Route 23B for 0.2 mi (0.3 km) to the junction with State Route 23. Park here. To reach outcrops 1A and 1B, park on the west shoulder of 23B just north of the entrance ramp that leads onto State Route 23 heading northwest. Depending on logistics, it may be easiest to walk the length of the highway to Catskill Creek. Drivers can pick up participants on the shoulder of Route 23 near Catskill Creek. **WARNING!** These highways are very busy,

and vehicles are traveling very fast. Outcrop 1A is along the exit ramp from State Route 23 northwest leading to 23B, and outcrop 1B is along the entrance ramp from 23B onto 23 northwest. To reach outcrops 1CN-1EN and 1CS-1ES, drive onto State Route 23 heading northwest toward Cairo.

Stop 2 - The Esopus Detachment, along Route 23A. (Lat 42°11'52.94"N, Long 73°55'1.45"W). These outcrops can be reached by continuing west from 1EN/1ES on State Route 23. Turn left (south) on State Route 47. Turn right (west) at the intersection with Old Kings Road and continue to the intersection with State Route 23A. Turn left (east) on State Route 23A and pull off the road to the right (south) after 0.1 miles onto the abandoned Thruway ramp. The exposure is a roadcut along the abandoned ramp.

Optional Stop B - Bakoven Shale Exposures, City View Terrace Road, Kingston (Lat 41°56'58.75"N, Long 74°2'31.29"W). Take the Kingston Exit (Exit 19) from the NYS Thruway. Follow the roundabout to Route 28. Head west on Route 28 across the Thruway, and then across Route 209. At 0.3 miles west of the Route 28 bridge over Route 209, turn right off of Route 28 onto Forest Hills Drive. Head north on Forest Hills Drive for about 50 m, and turn right to head east on City View Terrace Road. Pull off on the shoulder. The exposures at Optional Stop B are along the north side of City View Terrace Road.

Stop 3 - Route 209/199 roadcuts, north of Kingston. (Lat 41°58'27.37"N, Long 73°59'4.88"W to Lat 41°58'34.00"N, Long 73°58'30.84"W). To reach these outcrops, take NYS Thruway Exit 19. Leave the Thruway and pass through the tollgate. At the end of the tollgate access road, bear right (west) onto State Route 28. Proceed west on Rte 28 for 0.3 miles and bear right to take the entrance ramp onto Route 209 north. Proceed north on Rte 209 (toward the Rhinebeck Bridge) for about 3.7 miles. The road curves and heads east. When it crosses Route 9W, it becomes Route 199. Park along the south shoulder of Rte. 199/209 about 0.7 miles east of the State Route 9W underpass. [*Note:* If you are heading to Stop 3 after Optional Stop B, simply backtrack from B back to the junction with Route 209 to pick up the route described above.]

Stop 4 - Roadcuts exposing lateral ramps in the Rondout (Lat 41°58'12.84"N, Long 73°58'18.10"W). To reach these outcrops from Stop 3, continue east on Rte 199 to the exit for Route 32 (the last exit before the Kingston-Rhinecliff Bridge over the Hudson River). At the end of the ramp, turn left and head south on Rte 32 (= Flatbush Avenue). Proceed about 0.5 miles south on Route 32 and pull off on the shoulder to the right. The exposures are roadcuts on the west side of the road.

Optional Stop C - Cross-Cutting Cleavage in the Onondoga Limestone, Kingston. (Lat 41°56'33.88"N, Long 73°59'19.02"W). To reach this stop from Stop 4, proceed drive south to southwest along Rte 32 (Flatbush Road) for about 2.2 miles. You will go 0.2 miles past the entrance to Route 9W south (part of a divided highway), to the junction with non-divided Route 9W north (East Chester Street). There is a convenience store at the corner. Park where convenient, then walk north along the east side of Route 9W (East Chester Street) to the first roadcut. Climb to the top of the roadcut and look for bedding planes.

Stop 5 - Hasbrouck Park, Kingston (Lat 41°55'31.44"N, Long 73°58'32.57"W, and Lat 41°55'21.91"N, Long 73°58'41.63"W). Turn south off of Route 32 and proceed south for 1.1 miles on Route 9W south (part of a divided highway). Take the Delaware Avenue Exit. At the end of the exit ramp, turn left and proceed east on Delaware Avenue, on a bridge across Route 9w. You need to take the right turn immediately after the bridge to stay on Delaware Avenue. Proceed for 0.4 miles on Delaware Avenue to Hasbrouck Park Road. (The park entrance road is small and not well marked; it occurs before Delaware Avenue goes over the Helderberg Escarpment.) For the first part of this stop, proceed south on Hasbrouck Park Road for 100 m, and turn left into the small parking lot. Follow the trail from the parking lot to the outcrop in the woods, about 100 m to the east of the lot. For the second part of this stop, drive to the viewpoint at the south end of the park. First, look at the view to the south. Then, walk back north following the trail through the woods to where there is an access trail over the escarpment. The exposures are old roof and pillar quarries. **WARNING!** The trails are steep and slippery, and the cliffs have loose rock.

Optional Stop D - Exit Ramp from Route 9W to Delaware Avenue (Lat 41°55'39.86"N, Long 73°59'12.78"W). Backtrack from Hasbrouck Park toward Route 9W. Park on Delaware Avenue before the bridge over 9W, and walk to a viewpoint where you can see the roadcuts along 9W and the exit ramp from 9W to Delaware Avenue.

Optional Stop E - Callanan Quarry, Connelly (south of Kingston) (Lat 41°54'29.71"N, Long 73°59'47.00"W). This stop can be reached by driving south on US 9W from Kingston, crossing Rondout Creek, and continue into Port Ewen (about 0.7 miles south of the bridge). In Port Ewen, turn right (west) onto Salem Street (County Road 25). Continue on Salem Street across railroad tracks and around a sharp bend to the right. The entrance to Callanan Quarry is approximately 2.2 miles west of the intersection of Salem Road and US 9W. The Quarry cannot be entered without permission, hardhats, etc..

Optional Stop F - Rosendale Landfill (Lat 41°52'39.33"N, Long 74°04'06.41"W). This stop can be reached by continuing south on State Route 32 from Bloomington (Figure 1). Turn right onto Kallop Road 0.8 miles south of the New York State Thruway overpass and then bear right onto Hickory Bush Road at the first three-way intersection. Continue north on Hickory Bush Road for 1.0 mile and pull onto the dirt parking lot along the left side of the road. This stop is along the north wall of the Rosendale landfill and recycling center and permission to visit must be obtained from the Town of Rosendale.

Optional Stop G - Rail Trail exposure near Fourth Lake (Lat 41°52'22.68"N, Long 74°04'25.01"W). This stop can be reached by walking 350 south along the Wallkill rail trail from the dirt parking lot on Hickory Bush Road. Warning! The exposure is an abandoned roof and pillar mine. There are steep drop offs and obscured openings.

Optional Stop H – Former Williams Lake Hotel property (Lat 41°52'52.46"N, Long 74°04'50.11"W).

Stop 6 - Snyder Estate (Lat 41°50'28.70"N, Long 74°05'51.05"W). This stop can be reached by driving south on Hickory Bush Road for roughly 0.8 miles and turn right (west) onto Breezy Hill Road. At the first stop sign, turn left (south) onto Binnewater Road/County Route 7. Turn right (west) at the intersection with State Route 213 in the Town of Rosendale. Continue east on State Route 213 for 0.2 miles before turning right (north) into the driveway of the Snyder Estate. Proceed up the driveway and across the bridge before parking in the gravel lot on the right.

Optional Stop I – Rondout Creek Exposures, Lawrenceville (Lat 41°50'23.72"N, Long 74°06'23.35"W).

Optional Stop J - Roadcuts in Schwangunk, Old Route 213 (Lat 41°49'36.33"N, Long 74°07'32.19"W). Exposures at Stop 10 are along both north and south sides of State Route 213, 1.8 miles west of the Snyder Estate. Turn right onto Old Route 213 or Bruceville Road to park, because the road shoulders at the stop are low and generally muddy.

Stop 7 - Waterfall in the High Falls Shale, High Falls (Lat 41°49'45.07"N, Long 74° 7'56.90"W). To reach this outcrop, take Route 213 to High Falls. The outcrop is a stream cut in Rondout Creek, viewed from a state historic site, on the north side of Route 213, about 0.4 miles northwest of the village of High Falls, and about 1.4 miles southeast of the junction between Route 213 and Route 209.

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Hydrothermal Alteration, Mass Transfer and Magnetite Mineralization in Dextral Shear Zones, Western Hudson Highlands, NY

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INTRODUCTION

The Reading Prong is a Grenville basement massif that forms part of the spine of the Appalachians, connecting the Blue Ridge and Green Mountain provinces, in Pennsylvania, New Jersey, New York, and into Connecticut (Figure 1). Known as the New Jersey Highlands and the Hudson Highlands in New York, it hosts thousands of iron deposits, the first discovery of which was as early as 1730 (Lenik, 1996). These deposits were extensively mined throughout the 18th and 19th centuries (Lupulescu and Gates, 2006).

Earliest investigations of the iron districts in the Highlands focused on the ore deposits (*e.g.* Kitchell, 1857; Cook, 1868; Wendt, 1885; Ruttman, 1887), and generally accepted an origin from metamorphosed sediments. Rodgers (1840), however, proposed that the magnetite deposits were of igneous genesis. Colony (1923) later proposed that the ore deposits were the result of magmatic replacement of country rocks. More recent investigators found that the deposits were either the result of magma driven replacement or hydrothermal processes (*e.g.* Sims and Leonard, 1952; Hotz, 1953, 1954; Sims, 1958; Hagner *et al.*, 1963; Buddington, 1966; Collins, 1969; Baker and Buddington, 1970; Foose and McLelland, 1995; Martinko and Gates, 2000). Gundersen (1984, 1985, 2000) proposed that they formed in a back arc setting, but Volkert (2001) proposed that the magnetite deposits were formed in an extensional setting, prior to metamorphism. It is evident that there are multiple modes of emplacement for the magnetite bodies of the Hudson Highlands (Lupulescu and Gates, 2006), as there are multiple types. Gundersen (2000) and Puffer (2001) acknowledge several types, including vein deposits created by remobilization of magnetite into faults and fractures, as well as deposits that are related to plutonic rocks and hydrothermal activity (Gundersen, 2004).

This guide will focus on two, kilometer scale shear zones that host several hydrothermally mineralized veins containing massive magnetite bodies. The NE trending dextral shear zones formed in crystalline rocks of the western Hudson Highlands late in the Grenville orogenic cycle. The veins are exposed at several abandoned magnetite mines within Harriman State Park. New modeling for the mode of formation of these deposits is discussed, following an overview of the bedrock geology, and brief description of each of the veins. New geochemical evidence from the wall rock-vein contact will also be explored. This guide will then conclude with an overview of the current working model for the formation of these iron oxide deposits.

BEDROCK GEOLOGY

The origin of the bedrock of the Hudson and New Jersey Highlands has been the subject of much controversy over the past two centuries. Earliest interpretations agreed that the gneisses were of meta-sedimentary origin (*e.g.* Rodgers, 1840; Kitchell, 1857; and Cook, 1868), but later Spencer (1904, 1905, 1909) and Berkley (1907) proposed a plutonic origin. Bayley (1910) proposed that they were derived from both plutons and metamorphosed sediments. Later workers also concurred with this hypothesis (*e.g.* Sims and Leonard, 1952; Hotz, 1953, 1954; Sims, 1958; Hagner *et al.*, 1963; Buddington, 1966; Collins, 1969; Baker and Buddington, 1970; Foose and McLelland, 1995). More recent geochemical investigations found that some of the rocks are of plutonic origin (Gundersen, 1984, 1985, 1986, 2001; Volkert and Drake, 1999; Gates *et al.*, 2001, 2003, 2004; Puffer and Gorrington, 2005). However, others have recognized that many of the quartzofeldspathic and amphibole-pyroxene gneisses show striking similarities in major-element chemistry to volcanic rocks (Helenak, 1971; Jaffe and Jaffe, 1973; Drake, 1970, 1984; Murray, 1976; Grauch, 1978; Kastellic, 1979; Gundersen, 1984, 1985; Gates *et al.*, 2001, 2003, 2004; Puffer and Gorrington, 2005).

Currently, several models exist for the formation of these rocks. Gundersen (1984, 1985, 2001, 2004) proposed that many of these gneisses formed in an extensional backarc marginal basin whereas others have a bimodal, volcanic origin. In contrast, Volkert and Drake (1999) interpret many of the igneous gneisses a result of fractionation of a single parent diorite and the metasedimentary units as unconformable sequences. Gates *et al.* (2001) propose formation of a volcanic pile with a volcanoclastic apron in an island arc or marine magmatic arc setting, characterized by layered intermediate and mafic gneisses and associated plutons, for the genesis of these rocks.

Prior geologic mapping in this region sub-divided the abundant quartzofeldspathic gneisses based upon the individual ferromagnesian minerals (Dodd, 1965; Dallmeyer, 1974). However, considering that 80% of these rocks are quartz-feldspar gneisses (Gates, 2003), this guide will use the system first conceived by Gundersen (1986), and later adapted by Gates *et al.* (2001), using a type of sequence stratigraphy for metamorphic rocks. Units are grouped into lithofacies based on various rock types, to define quartzofeldspathic, metasedimentary (calc-silicate, pelite, and psammite), and metavolcanic (plagioclase-amphibole-pyroxene) assemblages (Figure 2).

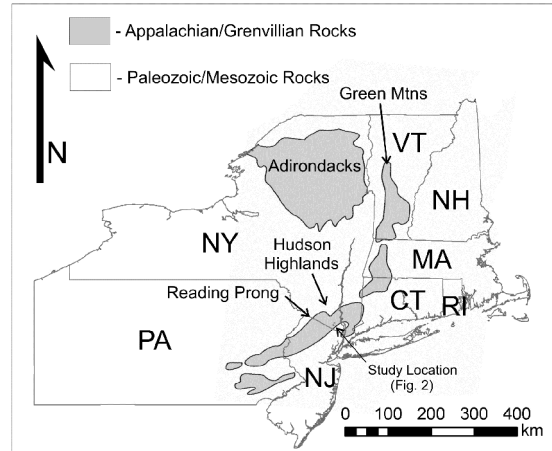


Figure 1. Map of the Northeastern United States showing distribution of Grenville rocks, the Reading Prong, Hudson Highlands, and study location (Figure 2).

Metavolcanic Gneiss

The metavolcanic unit consists of strongly banded sequences of inter-layered dark mafic and gray intermediate gneisses, interpreted to represent rocks with volcanic protoliths (Gates *et al.*, 2001). Compositional banding ranges in thickness from 5cm to 5m with varying quantities of each rock type. Mafic assemblages are composed primarily of medium to coarse grained amphibole, plagioclase, clinopyroxene and hypersthene, and minor magnetite locally. Intermediate bands contain medium to coarse grained plagioclase and quartz, with minor amounts of amphibole, biotite, clinopyroxene and hypersthene. This unit also contains localized interlayers of quartzite, marble, and calc-silicate gneiss, as well as migmatites. The contacts with the quartzofeldspathic unit are interstratal gradational.

Quartzofeldspathic Gneiss

The quartzofeldspathic gneiss ranges from massive to layered quartz-plagioclase gneiss to quartz-K-feldspar-plagioclase gneiss with lesser amounts of clinopyroxene, hypersthene, amphibole, and/or phlogopite, layer pending. Minor amounts of magnetite and garnet can also be observed locally. Compositional layering is defined by the proportions and species of ferromagnesian minerals present. Closest to the contact with the metavolcanic unit, it is also locally interlayered with quartzite and mafic gneiss. Locally, this unit also contains apparent fining upwards sequences defined by an increase in the amount of mica and decrease in the layer spacing, showing sharp contacts between sequences (Gates *et al.*, 2001). However, it is difficult to interpret such relict sequences in granulite terranes. Gradational contacts with the metavolcanic lithofacies, composition and mineralogy, and internal compositional layering suggest that this unit represents a volcanoclastic sequence (Gates *et al.*, 2001).

Metasedimentary Gneiss

Throughout the western Hudson Highlands there are belts of rock considered to have sedimentary protoliths including meta-pelitic, meta-psammitic, calcsilicate gneisses, quartzite and marble. Belts of rock may contain all or some of these lithologies interlayered at the scale of meters to 100's of meters (Gates *et al.*, 2001). The calc-silicate gneiss is quartzofeldspathic containing salite, K-feldspar, apatite, sphene, scapolite, and amphibole. Centimeter-scale quartzite layers and discontinuous layers of diopside marble also appear in this unit. The meta-pelite consists of interlayered biotite-garnet gneiss with medium to coarse quartz, plagioclase, K-feldspar, with cordierite and sillimanite locally (Gates *et al.*, 2001).

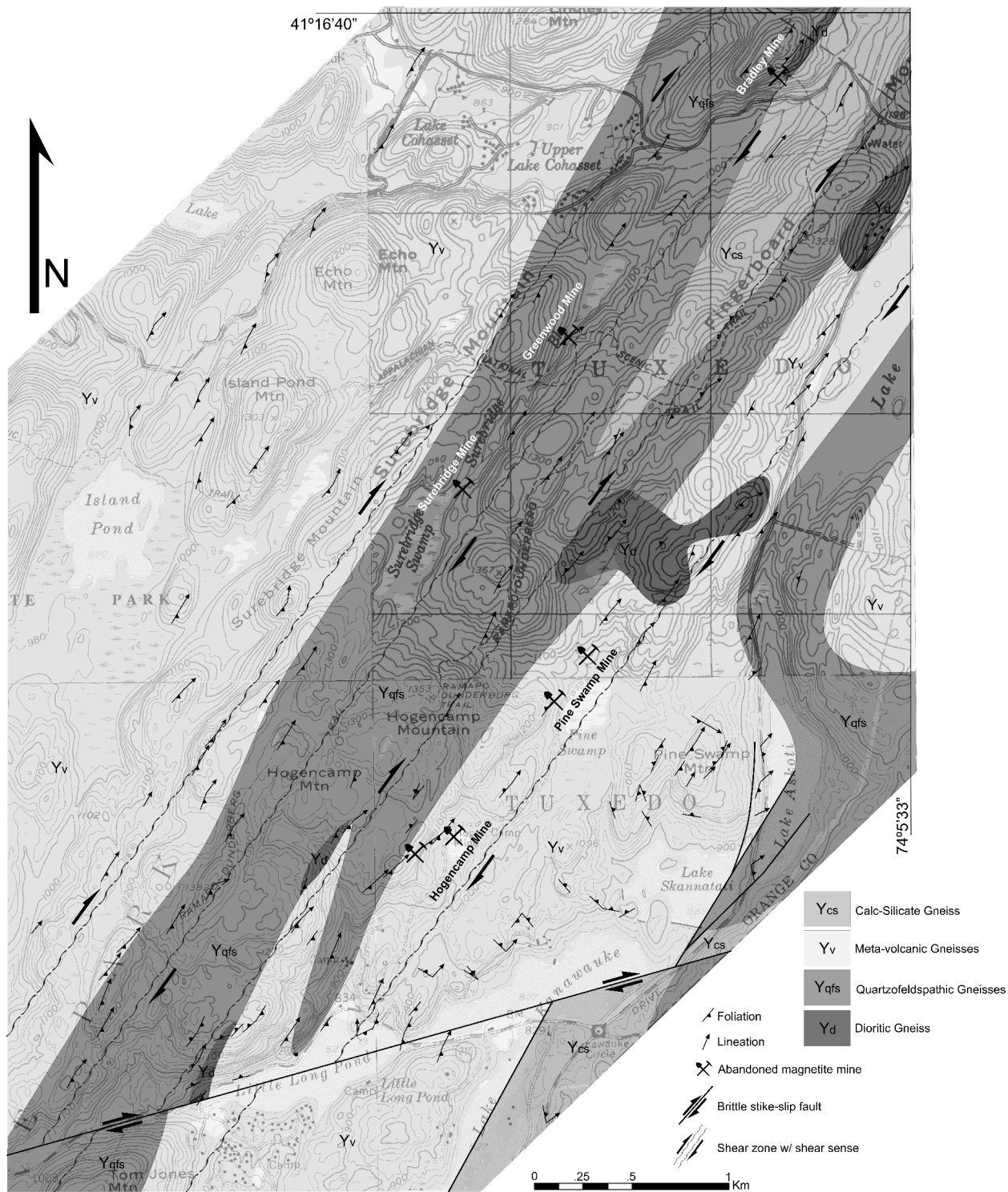


Figure 2. Geologic map of the field trip area, located within Harriman State Park, NY.

Lake Tiorati Diorite

Coarse to very-coarse grained black and white diorite dikes and bodies containing plagioclase, pyroxene, amphibole, and minor biotite locally, are found throughout the field area. Some small bodies appear concentrated in certain areas, possibly indicating larger bodies at depth (Gates *et al.*, 2001). The diorite also grades to pyroxene-poor, anorthositic compositions locally. Textures vary from coarse granoblastic to foliated and mylonitic with type II S-C fabrics (Lister and Snoke, 1984), exhibiting dextral shear sense (Figure 3). Locally, the diorite contains country rock xenoliths, showing intimate, ductile contacts which are partially melted, forming a rind of coarse pegmatite granite around them. Granite also fills fractures in the diorite that opened after crystallization, but while the granite was still liquid, which implies emplacement at depth (Gates *et al.*, 2001).

Pegmatites

Two generations of pegmatites occur throughout the field area. The earliest dikes are white and contain K-feldspar, quartz, muscovite and minor garnet locally. They are concordant to semi-concordant to the gneissic foliation, commonly boudinaged and containing internal fabrics and deformed grains. Thickness of these dikes ranges from 1cm to 1m. The later pegmatitic dikes are pink and very coarse grained, containing K-feldspar and quartz with muscovite, amphibole, magnetite, pyroxene, titanite, and/or garnet locally, depending on the rock intruded. They are highly discordant, commonly within brittle faults and contain xenoliths of faulted country rocks. They show minor to no deformational fabrics, and thickness ranges from 1m to 10m. They are also associated with small granite bodies (Gates *et al.*, 2001).

MINERALIZED ZONES

Two small, sub-kilometer wide shear zones formed within a 35-km-wide anastomosing dextral strike-slip shear system in the western Hudson Highlands (Gates, 1995). Each one contains a concordant to slightly discordant, late stage dilational, brittle fracture zone where hydrothermal fluids interacted with local country rock and deposited mineralized veins (Figure 4). Both NE-trending zones are defined by steeply dipping foliations, penetrative mineral lineations, and type II S-C mylonites with other dextral kinematic indicators (Figure 3). The vein-wall rock contact is sharp and semi-concordant to the mylonitic foliation. However, on the small-scale it appears slightly discordant, crosses foliation, and erodes into the wall rock.

The geometry of the vein deposits is characterized by three distinct zones sub-parallel to the wall rock boundary (Figure 5). The unaltered wall rock, grades into a 1-2cm “bleached zone” that is lighter in color and marked by alteration of the original minerals and formation of new minerals. The bleached zone is in direct contact with the vein deposit. The vein proper is characterized by two distinct zones, a layered sequence along the wall rock and a core of massive mineral assemblages. The layered sequence is characterized by distinct, dark colored bands of fine-coarse grained, pyroxene, amphibole, and/or biotite-rich assemblages, which range in thickness from 2-10 centimeters and have semi-gradational to sharp contacts. The core of the veins consists of massive deposits, characterized by very coarse, randomly oriented Fe/Mg-rich assemblages including magnetite and also containing late stage interstitial cementing minerals. The magnetite deposits in the core are characterized dominantly by massive magnetite, with minor am-

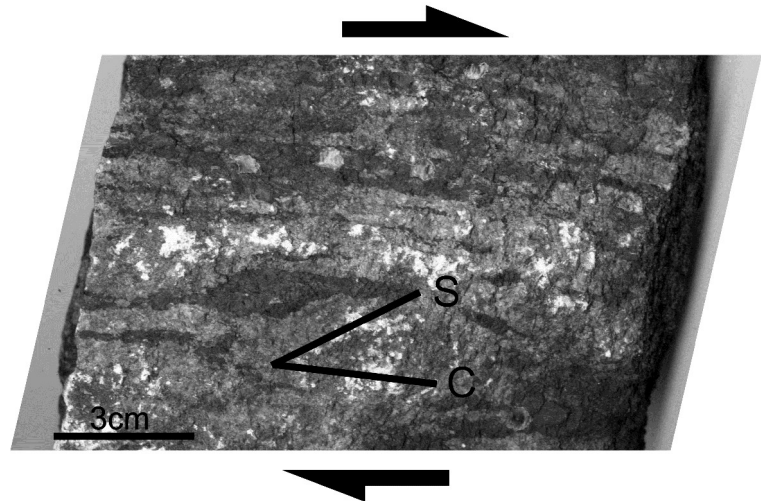


Figure 3. Hand sample of the meta-diorite, with type II S-C fabrics and σ porphyroclasts showing dextral shear sense.

phibole and pyroxene gangue minerals. The thickness of the core ranges from 1-10 meters. Thickness of the entire zones ranges from 2-15 meters and from tens of meters, to one kilometer in length. The much narrower zones that connect the magnetite deposits are thinner and typically composed of randomly oriented to aligned clinopyroxene with minor magnetite, phlogopite and/or quartz. The zones are commonly intruded by late pegmatite dikes that contain the mineralized rock as xenoliths.

Bleached zone mineral assemblages, vein material, and late-stage cementing minerals vary with location within each mineralized zone. In areas of Ca-rich country rock, clinopyroxene/calcite rich mineral assemblages dominate the bleached zone and throughout the vein. In areas of quartzofeldspathic country rock, amphibole/quartz assemblages dominate layered and massive vein material, whereas localities with sulfide-rich country rock contain orthopyroxene/sulfide rich vein assemblages.

Southeastern Shear Zone

Hogencamp Mine. The Hogencamp Mine lies in the southern part of the southeastern mineralized zone, where the sheared wall rock is dominated by metavolcanic gneisses, with calcsilicate gneiss and marble locally. It is characterized by a series of 1–10 meter horizontal and vertical mine shafts and open pit mines where the magnetite ore was extracted from the vein deposit. These mines can be traced along strike for up to one kilometer. The mineralized zone that hosts Hogencamp Mine is roughly 6 kilometers long and ranges in thickness from 2-10 meters at the mine locations to as little as one meter in the narrow zones connecting the massive deposits. The bleached zone is characterized by calcite and scapolite, and retrogression of pyroxene to amphibole, with phlogopite, and minor apatite locally. The layered portion of the vein is composed of hornblende- and orthopyroxene-rich layers, also containing calcite and phlogopite closest to the semi-gradational bleached zone contact where observed (Figure 5). The central massive portion of the vein deposit is characterized by substantial magnetite ore and gangue minerals of euhedral crystals of clinopyroxene cemented by late-stage, interstitial calcite (Figure 6 A).

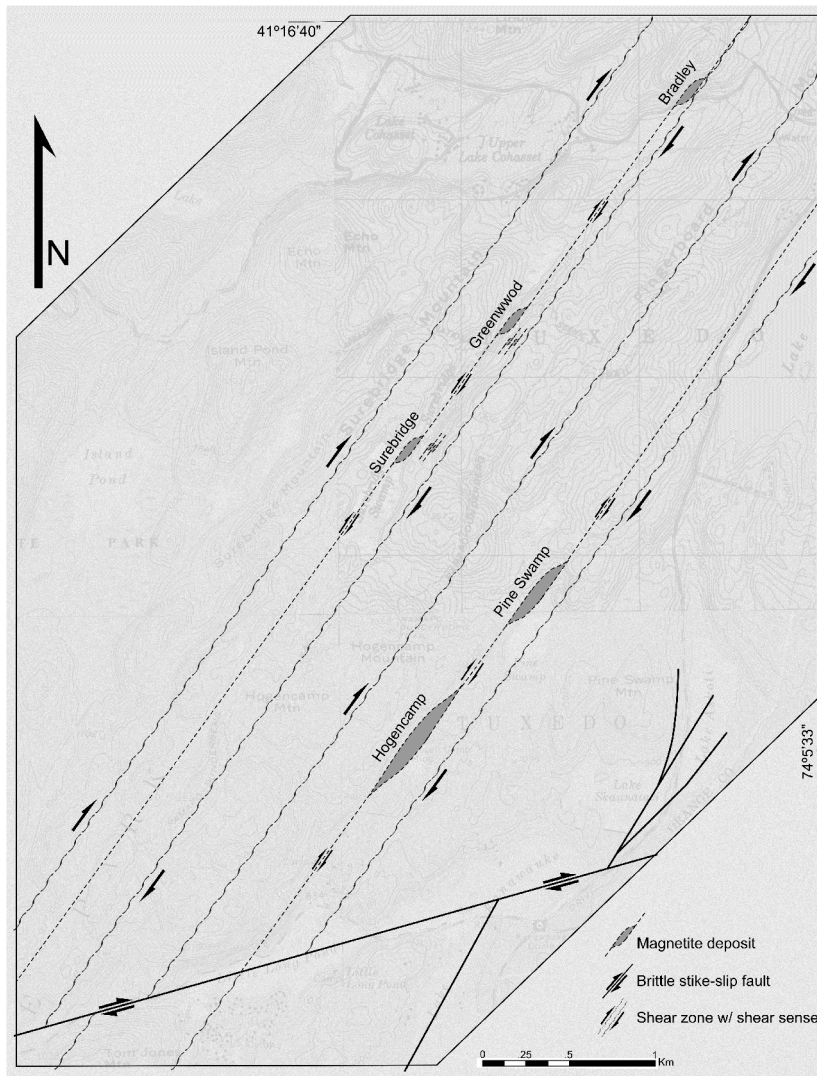


Figure 4. Map showing location and orientation of the vein deposits.

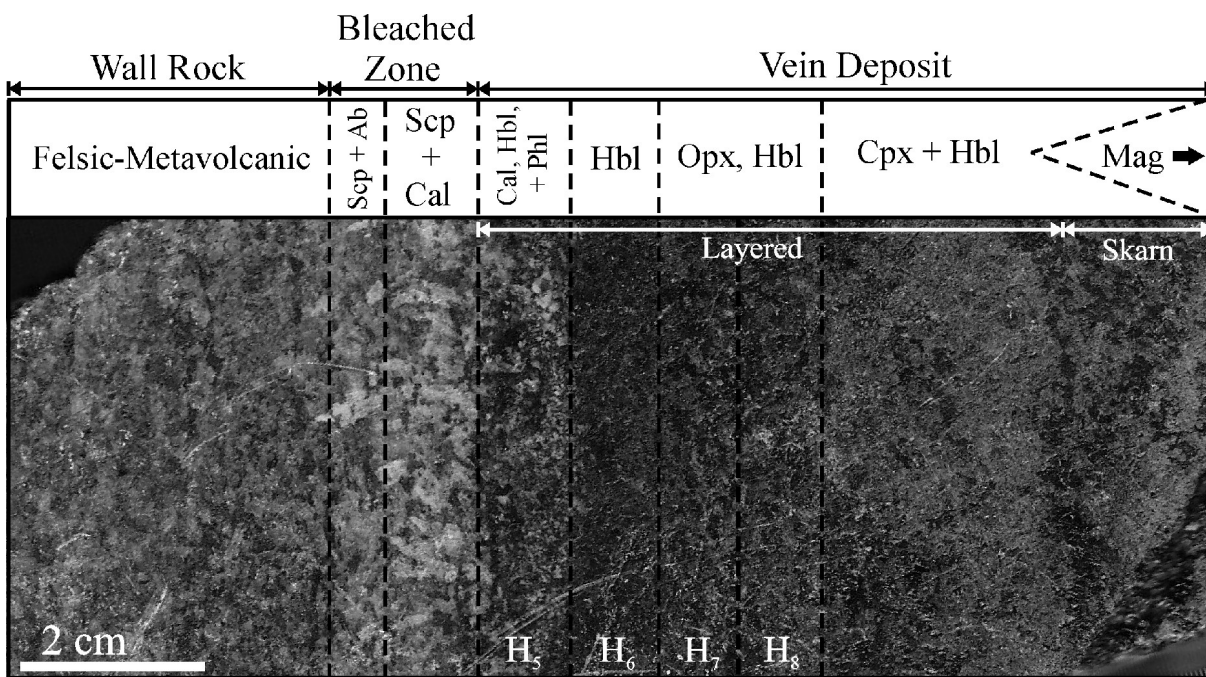


Figure 5. Hand sample from Hogencamp mine, showing banded mineral assemblages across the wall rock, bleached zone, vein deposit contact. Also shown are the layered and massive vein zones and layers H5-H8.

Pine Swamp Mine. Pine Swamp Mine, lies along strike, about one kilometer to the NE, of the Hogencamp Mine deposit. Pine Swamp is also characterized by a 5-meter wide horizontal mine shaft and meter scale open pit mines. However, unlike Hogencamp, the vein which hosts Pine Swamp Mine is only several hundred meters long, with minor semi-concordant veins in the northern extent of the deposit. This portion of the vein lies primarily in sulfide-bearing quartzofeldspathic gneiss and metavolcanic gneiss country rock, which varies among interlayered mafic and intermediate gneisses. The bleached zone is primarily defined by retrogression of pyroxene to amphibole, but also contains minor amounts of scapolite, and apatite locally. Orthopyroxene and amphibole dominate the thin layered vein and thick massive sequences including the magnetite deposits. Massive minerals are cemented by late stage sulfide minerals, mainly pyrite and pyrrhotite.

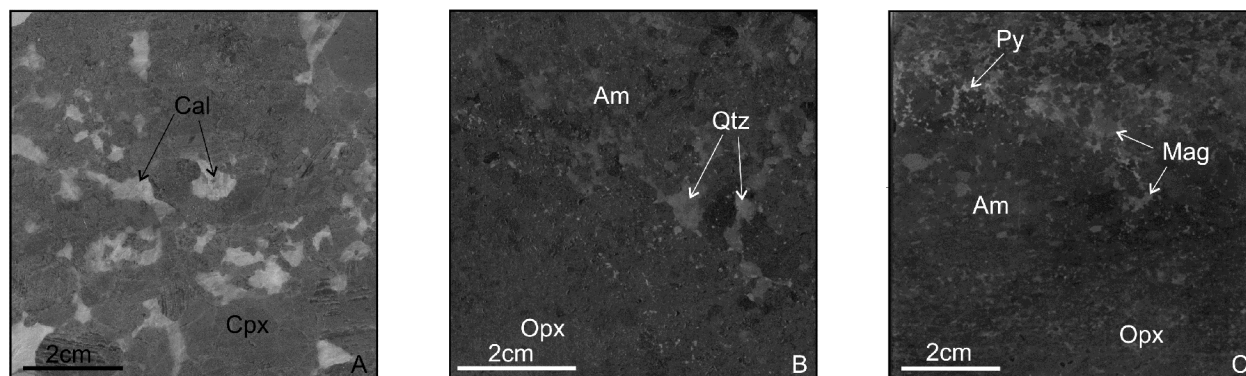


Figure 6. Hand samples of massive vein material from Hogencamp (A) and Greenwood Mines (B & C), showing late-stage interstitial cementing minerals; (A) calcite in massive clinopyroxene, (B) quartz in amphibole, orthopyroxene, and minor magnetite, and (C) pyrite in amphibole, magnetite, and layered orthopyroxene.

Northwestern Shear Zone

Bradley Mine. Bradley Mine is the northern-most deposit in the northwestern shear zone. Bradley Mine hosts a several meter discontinuous body of diopside marble. This marble is composed of medium-coarse grained calcite, fine-medium diopside, which occurs in cm-scale irregular aggregates, and it includes minor amounts of garnet locally. The vein deposit is a few hundred meters in length and varies in thickness from 2 to 10 meters. The bleached zone here is characterized by scapolite, diopside, and sericite, followed by a narrow zone of layered clinopyroxene and very fine grained micas. Massive intergrowths of clinopyroxene cemented by calcite, and magnetite appear in the core of the deposit.

Greenwood Mine. Greenwood Mine lies in the middle of the northwestern shear zone, between the Bradley mine to the NE and the Surebridge mine to the SW. It occurs within the quartzofeldspathic gneiss unit and the deposit is characterized by a series of sub-parallel veins. The main deposit ranges in thickness from 2-10 meters and several tens of meters in length. The two sub-parallel veins are a few hundred meters to the SE of the mine and are 2-4 meters thick and 10-20 meters long. The bleached zone is absent in all cases with the vein in direct contact with the wall rock. Most wall rock-vein contacts are relatively sharp, and lead directly into a layered orthopyroxene-rich zone, with minor amphibole. The massive zone contains magnetite and amphibole/orthopyroxene-rich deposits cemented by interstitial quartz, or less commonly pyrite and, pyrrhotite (Figure 6, B & C).

Surebridge Mine. The Surebridge Mine is located at the southern end of the northwestern shear zone. Country rock is sulfide-bearing quartzofeldspathic gneiss, but also contains mafic metavolcanic gneiss. The Surebridge Mine is characterized by a roughly 5 meter wide by 30 meter long central vein deposit, and two, narrower, sub-parallel secondary vein deposits. These secondary vein deposits are about 100 meters to the east of the main vein. The much narrower veins are about 2-3 meters thick and up to 20-30 meters in length. They can currently be observed as two, 5 meter deep pits, from which the iron ore was extracted. The narrow bleached zone at Surebridge is composed primarily of amphibole with only minor scapolite and sericite locally. The layered vein sequence in this locality is only a few centimeters in thickness and dominantly composed of amphibole and magnetite with minor pyroxene and quartz. Massive minerals are pyroxenes and amphibole, followed by magnetite in the core, with sulfides and some quartz as a late cementing agent.

GEOCHEMISTRY

Samples were collected at each of the mines, across the wall rock-bleached zone boundary, for chemical analysis. They include one each from the northern and southern Hogencamp and Pine Swamp deposits, and one each from the Bradley and Surebridge deposits. Small, cm-scale samples were removed from the unaltered wall rock and from the bleached zone at each locality and pulverized for chemical analysis. These samples were analyzed for bulk oxides

	<u>Hogencamp</u>				<u>Pine Swamp</u>				<u>Bradley</u>		<u>Surebridge</u>	
	<u>North</u>		<u>South</u>		<u>North</u>		<u>South</u>		Wall	Bleach	Wall	Bleach
	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach	Wall	Bleach
SiO ₂	68.30	55.96	57.48	53.06	56.06	53.29	66.18	63.48	43.88	43.01	69.46	59.39
TiO ₂	0.43	0.31	1.10	2.02	0.95	1.40	0.72	0.78	0.13	0.11	0.36	1.00
Al ₂ O ₃	12.98	14.05	16.04	14.89	15.92	14.59	13.80	15.56	17.23	19.09	13.84	13.97
Fe ₂ O ₃	4.51	9.69	8.46	10.86	8.29	10.02	6.64	8.09	12.66	11.54	6.82	11.28
MnO	0.16	0.23	0.05	0.05	0.05	0.05	0.15	0.17	0.15	0.17	0.09	0.16
MgO	1.73	3.80	4.75	6.06	4.83	6.19	2.36	2.37	3.85	3.27	2.07	4.20
CaO	3.86	11.93	7.90	9.40	8.13	9.24	4.89	5.48	19.79	21.89	4.39	7.13
Na ₂ O	3.11	4.23	5.08	4.41	5.01	4.35	4.23	4.67	0.83	0.15	4.36	3.89
K ₂ O	5.34	0.57	1.63	1.12	1.34	1.17	1.61	2.47	0.18	0.07	0.79	0.95
P ₂ O ₅	0.11	0.24	0.22	0.18	0.20	0.19	0.16	0.17	1.49	0.68	0.06	0.29
Total	100.53	101.03	102.71	102.06	100.78	100.50	100.74	103.24	100.19	99.98	102.24	102.25

Table 1. Bulk oxide geochemical results (in wt.%) for samples from select mines. Wall rock compositions are an average from several analyses.

and minor elements using a JY ULTIMA-C, ICP-OES at the ICP OES & MS Laboratory at Montclair State University, New Jersey. Wall rock chemistries were averaged at each sampling location to reduce variations due to compositional heterogeneity in the cm-scale samples of gneiss (Table 1) (Grant, 2005). Bulk rock chemistries were also acquired in cm-scale bands across layered vein material from Hogencamp Mine (Table 2, Figure 7). The locations of the bands are shown in Figure 5.

Bulk oxide chemical analyses results are shown in Table 1. Wall rock samples from Hogencamp Mine varies from felsic to intermediate in composition, containing about 58-68% silica, moderate amounts of alumina (13-16%) and iron (4.5-8.5%), and varying amounts of potash (1.6-5.3%), soda (3-5%), and calcium (3.8-7.9%). Pine Swamp wall rock samples show similar bulk chemistries to Hogencamp Mine. Silica ranges from 56-66%, alumina varies from about 14-16%, and iron ranges from 6.6-8.2%, whereas soda and potash appear relatively consistent between sample locations. Larger differences in calcium (4.8-8.1%) and magnesium (2.3-4.8%) are also observed. Calc-silicate wall rock at Bradley is very high in calcium (19.8%), alumina (17.2%), and iron (12.6%), and very low in silica (43.9%), soda (0.8%), and potash (0.2%). Surebridge Mine samples are from sulfide-bearing quartzofeldspathic gneiss, characterized by almost 70% silica, with significant amounts of alumina (13.8%), iron (6.8%), calcium (4.4%), and soda (4.4%), and minor quantities of magnesium (2%) and potash (0.8%).

Bulk chemical composition of the layered vein material from Hogencamp Mine resembles that of a mafic to ultramafic igneous rock (Table 2). Silica progressively decreases from 48-44.8% into the vein, whereas magnesium increases from 6.1-13.3%. Iron shows a small net loss from 13.1-11.6%, inward whereas alumina and soda decrease progressively away from the wall and increase into the innermost analyzed layer, from 11.2-7.5% and 1.9-0.7% respectively. Potash decreases close to the wall rock from 0.7-.3%, but increases to 1.1% in the innermost layer. Similarly, calcium and phosphorus progressively increase but then decrease in the innermost layer from 17.6-21.4% down to 18.6%, and 0.5-1.3% down to 1.0%, respectively. Titanium remains relatively stable around 0.2-0.3% across all layers (Figure 7).

The geochemistry of the wall rock and bleached zone samples were compared to constrain elemental gains and losses into the metasomatic fluids within the zone. However, comparison of two sets of geochemical data may lead to misinterpretations without the knowledge of the relationship between composition and volume changes that accompany the processes (Gresens, 1967). To that end, Grant's Isocon analysis (1987), after Gresens (1967) equation for metasomatic alteration was applied to the bulk oxide geochemistry results and density calculations. In doing so, constant alumina was generally assumed.

Hogencamp Vein				
	H ₅	H ₆	H ₇	H ₈
SiO ₂	48.00	47.13	45.98	44.76
TiO ₂	0.22	0.32	0.32	0.32
Al ₂ O ₃	11.18	7.93	7.49	8.62
Fe ₂ O ₃	13.06	13.95	12.10	11.59
MnO	0.25	0.27	0.20	0.17
MgO	6.08	8.59	11.73	13.25
CaO	17.58	20.45	21.42	18.58
Na ₂ O	1.86	0.91	0.65	0.91
K ₂ O	0.67	0.28	0.50	1.06
P ₂ O ₅	0.49	0.79	1.30	1.02
Total	99.42	100.61	101.69	100.27

Table 2. Bulk oxide geochemical results (in wt.%) from layered vein material.

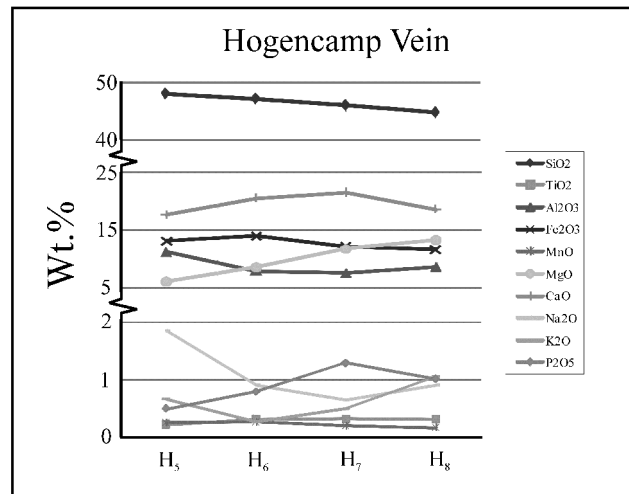


Figure 7. Plot of bulk oxides (wt%) for data in Table 2, Figure 5, showing layered vein material progression.

The results are shown in Figure 8. Samples from both northern and southern Hogencamp and Pine Swamp deposits reveal significant gains in CaO, MgO, and Fe₂O₃, with considerable losses in K₂O, and SiO₂ to a lesser extent (Fig-

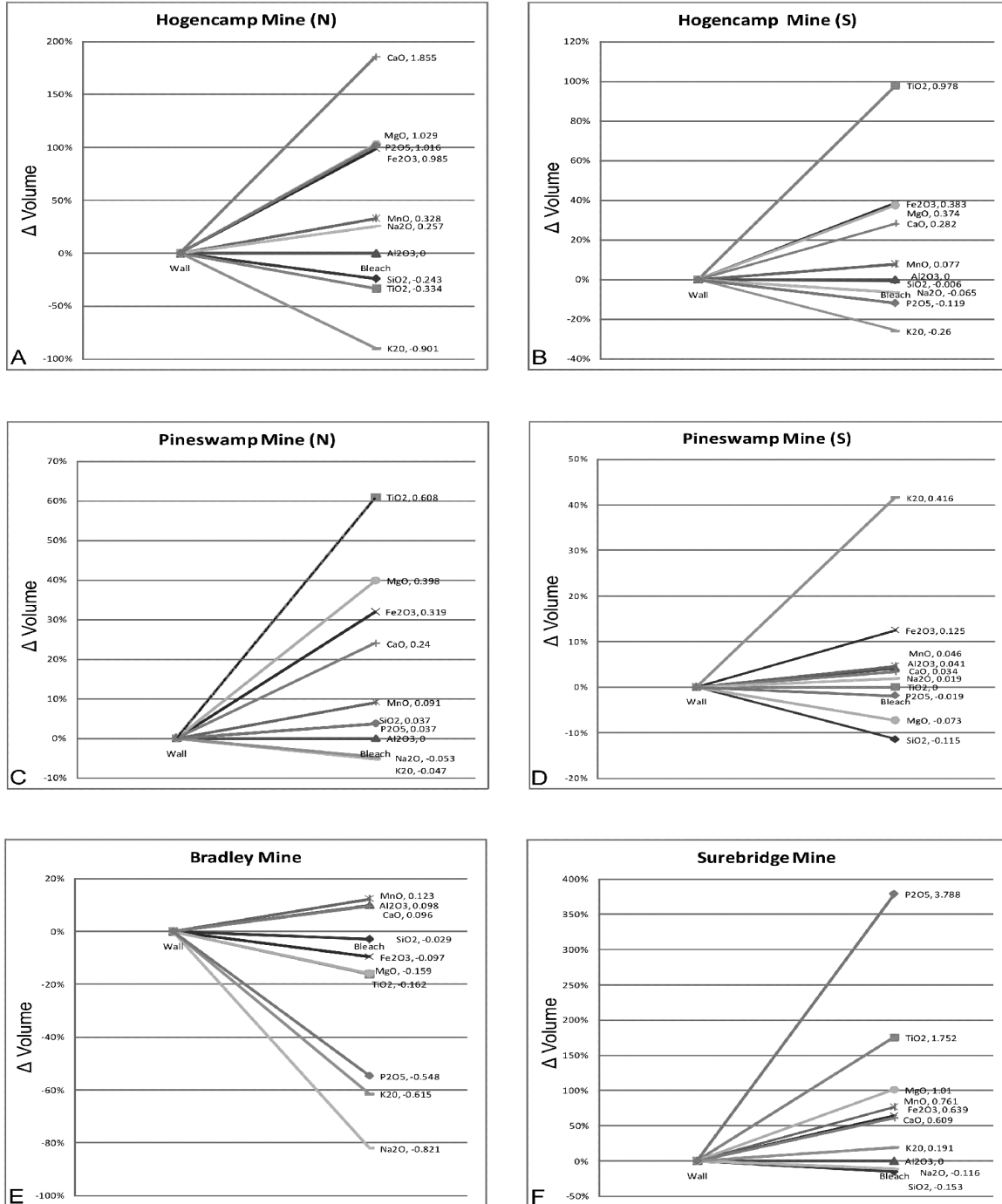


Figure 8. Volume percentage change in oxides for samples from select mines, across the wall rock-bleached zone boundary, as determined by Grant's isocon analysis (1987).

ure 8, A-D), into the bleached zone. Mass transfer analysis from the Bradley deposit shows small gains in CaO and Al₂O₃, and large losses in Na₂O, and K₂O, and lesser amounts of MgO and Fe₂O₃ (Figure 8, E). Geochemical modeling from the Surebridge deposit reveals significant gains in MgO, Fe₂O₃, and CaO, with only minor losses of SiO₂ and Na₂O (Figure 8F).

DISCUSSION

A volcanic pile formed about 1.2 Ga in an island arc or marine magmatic arc setting characterized by layered intermediate and mafic rocks, and associated plutons and volcanoclastic sediments (Gates *et al.*, 2001). This sequence underwent granulite facies metamorphism, about 1,050 Ma associated with the Ottawa phase of the Grenville orogeny (Gates *et al.*, 2003). Locally, anatexis produced migmatites, granite sheets, and the early pegmatites (Gates *et al.*, 2003). Subsequent diorite intrusions occurred around 1,008 Ma, either the result of delamination at the end of the first event, or the early dilational stages of the next event (Gates *et al.*, 2003). This second event is characterized by dextral strike-slip movement during a period of rapid uplift and unroofing at approximately 1,008 Ma to 924 Ma (Gates and Krol, 1998). Thick anastomosing zones of mylonite formed, overprinting previous features. Offset reached upwards of 100's of kilometers (Gates *et al.*, 2003). Strike-slip shearing continued during rapidly decreasing temperatures, resulting in the shear zones crossing the brittle-ductile transition and becoming dilational (Gates *et al.*, 2003).

These dilational structures began to open at small angles to the shear zone boundaries, resulting in sub-parallel, pull-apart structures (McCaig, 1987), connected by brittle faults. Hydrothermal fluids flushed into the faults, and chemically equilibrated with the composition of the local wall rock. These reactions produced the bleached zones in the wall rocks. In this way, the fluids were chemically buffered locally, and varied in composition along strike. During continued deformation, these now buffered metamorphic fluids were transported along strike likely through such mechanisms as seismic pumping (Sibson *et al.*, 1975). Alteration in the bleached zone and precipitation in the vein was driven by chemical or physical changes encountered as the fluids were flushed along strike. This was controlled by the stability or instability of individual minerals in response to changes in P, T, pH, or solution composition (Eugster, 1986).

Acid-base reactions are important in metamorphic and ore-forming environments, because in many mineral alteration reactions, Al is conserved in the solid, and HCl is produced (Eugster, 1986). This is the case for the alteration of K- and Na-feldspars to mica (Hemley, 1959). The addition of SiO₂ and S₂ into solution may also drive pH levels of the fluid to higher acidity, whereas introduction of Ca drives fluids to more basic values. Such positive feedback, of varying pH solutions from addition and subtraction of certain chemical species into and out of solution, may have driven the earliest stages of the bleached zone formation as fluids fluxed along the zone. Furthermore, redox, acid-base, and exchange reactions are the principal mechanisms by which metals are acquired by hydrothermal fluids (Eugster, 1986). Acids released during early wall rock reactions may have aided in the mobilization of metals which contribute to the later massive deposits, where Fe is preferentially mobilized during more oxidizing conditions (Eugster, 1986). Ultimately, chemical interaction primarily drove much of the early precipitation and alteration seen in the bleached zone and layered vein sequence, whereas decreasing pressure gradients primarily caused precipitation of the later massive mineral assemblages.

Higher pressures and acidity permitted higher levels of iron and magnesium to remain in solution in the narrow faults connecting the mineralized zones. Wall rock in these areas likely exhibited lithostatic pressure, and buffered the pH of fluids to more acidic values (Eugster, 1986). Dropping from lithostatic pressure, to hydrostatic pressure at the extensional segments of the fault, where fluids were also buffered to higher pH, led certain chemical species to become supersaturated and prompted precipitation.

Influx of Fe/Mg-rich fluids into the dilational fractures resulted in the deposition of the early pyroxene/amphibole-rich assemblages. Locally buffered fluids may have interacted in some localities, resulting in early clinopyroxene-rich layered and massive mineral assemblages in areas of CaO enrichment, amphibole-rich in areas of SiO₂ enrichment, and orthopyroxene-rich in localities dominantly increased only in Fe₂O₃ and MgO. Layered vein sequences were likely formed in the earlier volume confined conditions, whereas the conditions favored formation of massive

assemblages as dilation continued, volume increased and pressures began to drop. In this way, the main driving force switched from a chemical to pressure controlled precipitation mechanism, forming the massive ferromagnesian and magnetite assemblages, central to the deposit.

Inflowing fluids progressively became extremely Fe-rich, magnetite ore forming fluids, as these dilational structures continuously opened, and the fluids continued depositing the massive magnetite ore bodies. Residual, depleted fluids or fluids later buffered from local country rock, played a role in the late-stage, interstitial, cementing minerals found throughout the vein deposits. Calcite, quartz, and sulfides cement occurred in areas of calcium, silica, and sulfide-rich country rock, respectively. Granitic pegmatites intruded as fault activity waned, concentrated along the faults, suggesting a genetic relationship (Gates *et al.*, 2003). Fe-rich, dark pink potassium feldspars are evidence of intrusion along the same iron-oxide forming fluid conduits.

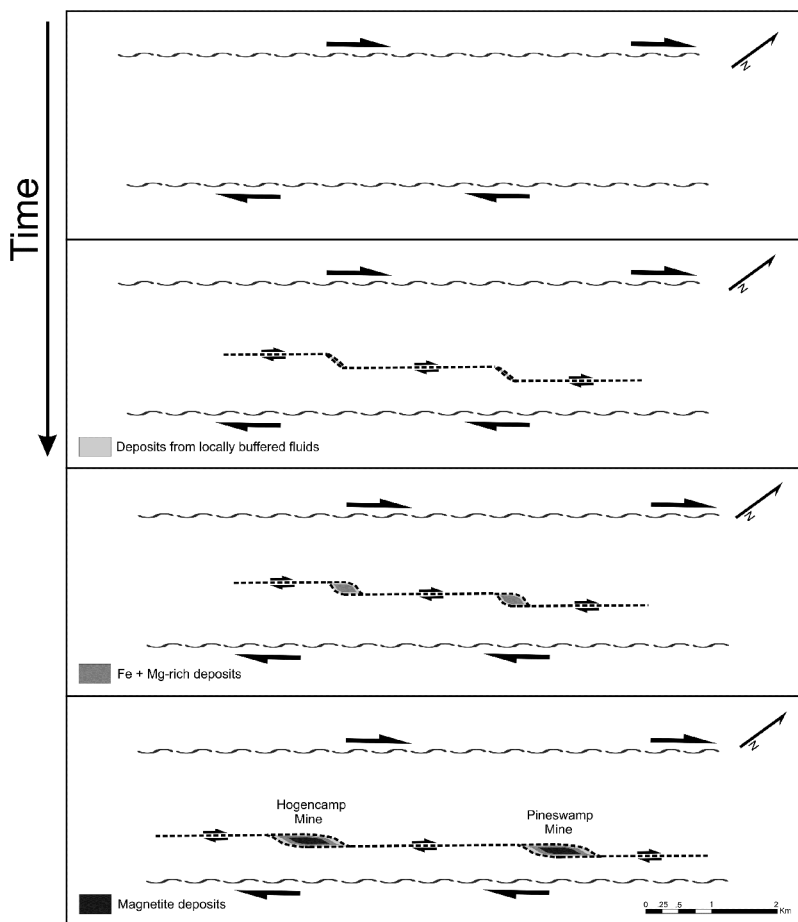


Figure 9. Schematic model of the successive stages of the vein deposits formation.

CONCLUSION

New geochemical and structural analysis of the mineralized faults leads to the interpretation of a mode of formation for one type of massive magnetite ore bodies in Harriman State Park (Figure 9). During the latest stages of dextral shearing, rapid temperature decreases caused the shear zones to cross the brittle-ductile transition and dilational structures formed a variably closed and open fracture system. Metamorphic fluids flushed through fractures and reacted with wall rocks. Changes in chemistries into the altered bleached zone, at the wall of the vein, relative to unaltered country rock reflect the exchange of various chemical species from the buffered fluids. Buffered to the composition of local country rocks, these fluids were transported along strike, eventually encountering favorable physical and/or chemical conditions for precipitation.

As deformation continued, the dilational structures continued to pull-apart, and fluids turned to ferromagnesian-rich solutions, interacted with locally buffered fluids, and deposited early layered deposits, primarily pyroxenes and amphiboles. With continued dilation, the dominant mode of precipitation switched from chemical to pressure driven, favoring the deposition of massive over layered deposits. The zones continuously opened, as fluids turned extremely iron-rich. As the fluids flushed through the dilational zones, they encountered lower pressures, less acidic conditions, and oxidized dissolved Fe, deposited as the massive mineralized magnetite bodies. Late interstitial minerals were precipitated from residual fluids that were locally buffered. Latest pegmatites intruded along some of the same pathways as the vein forming fluids.

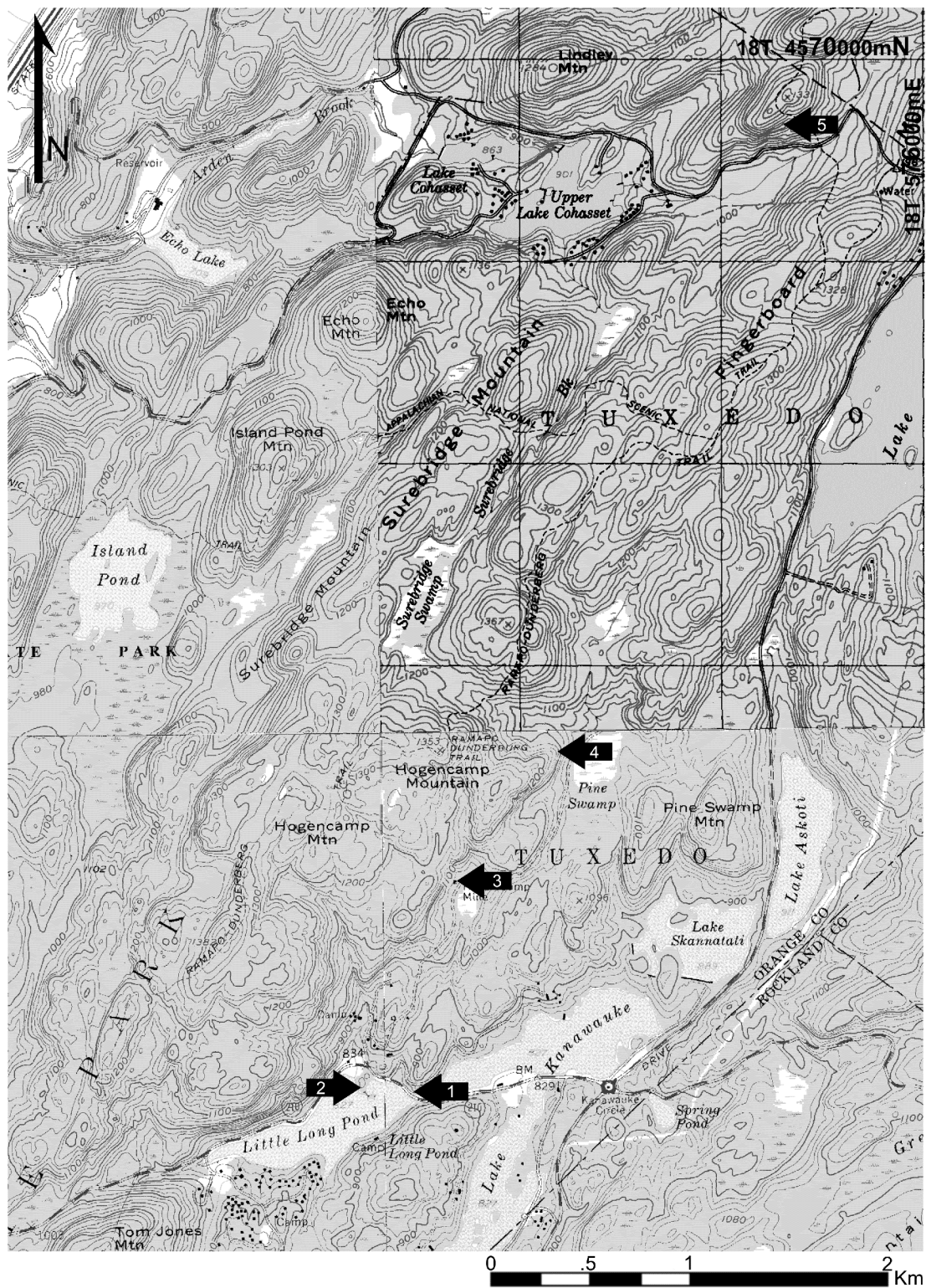


Figure 10. Map of field area showing locations of each step.

ROAD LOG

As we are in a State Park, we would ask that no hammer be used and only pictures taken. Be prepared for two and a half miles of trail hiking.

Mileage

- 0.0 Start at SUNY New Paltz, NY.
- 0.3 Turn right at County Route 17/Jansen Rd.
- 1.8 Turn right at NY-208/State Route 208.
- 3.1 Turn right at Main St/NY-299/NY-32.
- 34.5 Take the ramp onto I-87 South.
- 35.1 Take exit 16 for NY-17/US-6 toward Harriman.
- 35.4 Keep right at the fork; follow signs for Harriman/US-6/NY-17/West Point.
- 39.5 Turn left at NY-17/NY-32/State Route 32.
- 42.0 Turn left at Arden Valley Road.
- 44.6 Turn left to stay on Arden Valley Road.
- 47.9 At the Tiorati traffic circle, take the 1st exit onto Seven Lakes Parkway, heading south.
- 51.2 At the Kanawauke traffic circle, take the third exit onto County Route 106, heading west.
- 51.8 Continue on County Route 106 for 0.6 miles until reaching the bridge over the eastern-most extent of Little Long Pond. If needed, parking can be found a quarter mile east on route 106 at the picnic area (lavatory facilities). Walk 100 feet to the north-west on route 106 to the first road-cut (Stop 1, Figure 11).

Stop 1. Unsheared Metavolcanic Gneiss (30 MINUTES) (UTM: 18 T 0573484 4565104). Strongly interlayered intermediate and mafic gneisses with migmatitic bodies, just outside of the SE shear zone (Figure 11). Mafic layers are characterized by assemblages of clinopyroxene and amphibole with minor plagioclase, magnetite, sphene, and apatite. Intermediate layers are mainly plagioclase with minor quartz, apatite, amphibole, and biotite. The felsic leucosome is composed of coarse plagioclase, quartz, and K-feldspar, which form veins and clots including classic “net veining”. Minerals exhibit preferred orientations in the gneiss and appear granular in the leucosome. Late stage K-feldspar pegmatites can also be observed in this outcrop, containing mafic gneiss xenoliths.

Gneisses exhibit a strongly banded, intermediately dipping, foliation that strikes N-NE. Isoclinal intrafolial folds can also be observed, with axis following similar orientation. This deformation is indicative of main stage Grenville tec-



Figure 11. Stop 1, Road-cut of migmatitic metavolcanic country rock.

tonism, and was unaffected during the later tectonic event, associated with the formation of the ore deposits. Contrast these rocks with Stop 2.

- 51.9 Head west on route 106 for roughly 600 feet to the large peninsula protruding into Little Long Pond. Proceed 250 feet to the south to reach the tip of the peninsula (Stop 2, Figure 12).

Stop 2. Southeastern Shear Zone Boundary (30 MINUTES) (UTM: 18 T 0573272 4565124). Several meter scale lozenge and cigar shaped boudins of mafic gneiss are contained within mylonitic quartzofeldspathic gneiss, with folded biotite and local amphibole-rich layers (Figure 12). The layers appear contorted and wrap around the mafic bodies. The encased mafic gneiss is similar to that of Stop 1, however it also contains contorted folds and veins of magnetite. The long axis of the bodies and fold axis appear sub-parallel, and shallowly plunge to the northeast.

Both Stops 1 and 2 are similar in composition and therefore grouped within the same metavolcanic sequence. Long axis and fold axes roughly parallel shear zone boundaries and fabrics within. This location is at the edge of the southeastern dextral, strike-slip shear zone. Deformation steadily increases to the northwest, into the central shear zone, characterized by a steepening of planar fabric and increase in intensity of linear fabric. The contrast of the features at Stop 1 with Stop 2 shows the difference between the first main Grenville and second strike-slip events.

- 51.9 Locate the gated path directly on the opposite side of route 106. Walk this path heading north for a quarter mile, to the intersection with Dunning Trail (E-W trending). Head roughly northeast on Dunning Trail for about a half mile until you reach several large holes in the ground, and the path crosses a small stream. Hike upstream for roughly 100 feet until you are almost cliff side. A linear open pit mine should be visible to the southwest and a large mine shaft to the northeast into the cliff, beneath Cape Horn (Stop 3, Figure 13).

Stop 3. Hogencamp Mine (50 MINUTES) (UTM: 18 T 0573790 4566280). Hogencamp Mine lies in the southern part of the SE mineralized zone. Here the sheared wall rock is dominated by quartzofeldspathic and amphibole-pyroxene (metavolcanic) gneisses with interlayered calc-silicate gneiss and marble locally. Hogencamp Mine was active from the earliest to latest 18th century. It is characterized by a series of meter to several meter scale horizontal and vertical mine shafts and open pit mines. The mines can be traced along strike for up to a kilometer. The mineralized zone that hosts Hogencamp Mine is roughly 6 kilometers long and extends into Pine Swamp Mine (Stop 4). The vein ranges in thickness from 3-15 meters at the mine locations to as little as one meter in the narrow zones connecting the deposits. The Hogencamp Mines can be followed to the southwest from this location for up to a kilometer.

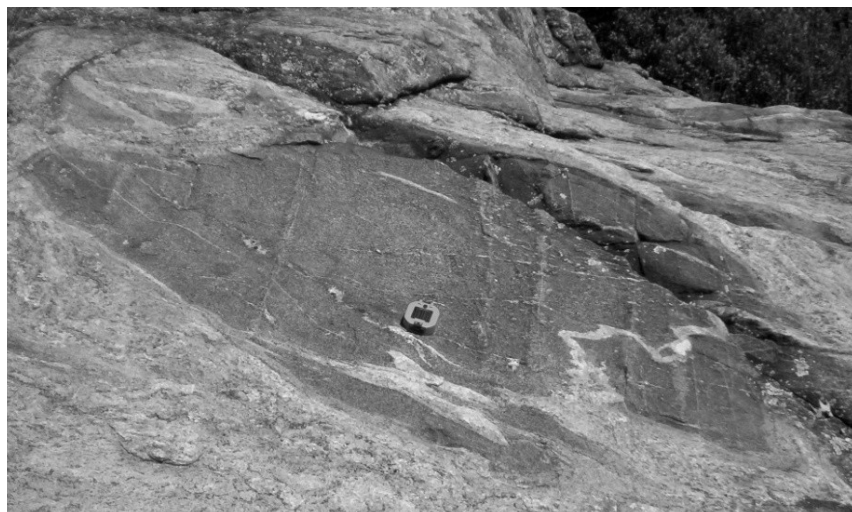


Figure 12. Stop 2, Outcrop on tip of peninsula, northeastern Little Long Pond.



Figure 13. Hogencamp (Stop 3), Pine Swamp (Stop 4), and Bradley Mines (Stop 5). From left to right.

The vein-wall rock contact is sharp and semi-concordant to mylonitic foliation. On the small-scale it appears slightly discordant, crosses foliation and erodes into the wall rock. This is best observed in the open pit mine, directly in front of the northeastern most mine shaft, below Cape Horn (Figure 13). Here, the bleached zone is characterized by the deposition of calcite and scapolite, and retrogression pyroxene to amphibole, also containing phlogopite, calcite, and minor apatite locally. Earliest vein deposit is characterized by layered amphibole, orthopyroxene, and clinopyroxene, later by massive clinopyroxene and magnetite, cemented by late stage, interstitial calcite in the ore zone. Clinopyroxene and localized magnetite are euhedral, forming doubly terminated crystals, thought to have crystallized in cavities. The veins are also intruded by very coarse grained pegmatites which contain xenoliths of the mafic vein material.

51.9 Back on Dunning Trail, walk east than north-northeast, another half mile until you reach a large swamp to the east, and a large, steep hillside to the west, littered with dark-colored mine tailings. Locate the makeshift path on the hillside. Take this path uphill for about two hundred feet, until your reach the entrance to the mine (Stop 4, Figure 13).

Stop 4. Pine Swamp Mine (50 MINUTES) (UTM: 18 T 0574253 4566795). Pine Swamp Mine lies in the northern part of the SE shear zone within the same mineralized vein which hosts Hogencamp Mine, roughly one kilometer along strike (NE). Pine Swamp is also characterized by a 3-12 meter horizontal mine openings and several, meter scale open pit mines (Figure 13). The mines that compromise the Pine Swamp deposit can be traced along strike for several hundred meters, with minor semi-concordant offshoots in the northern extent of the deposit. This portion of the hydrothermal vein lies dominantly within sulfide-bearing quartzofeldspathic gneiss country rock, which contains interlayered metavolcanic gneisses. The bleached zone is primarily defined by retrogression of pyroxene to amphibole, also containing minor amounts of scapolite, and apatite locally. Orthopyroxene and amphibole dominate the narrow layered vein and thick massive sequences followed by the magnetite deposits. Massive minerals are locally cemented by late stage sulfide minerals, mainly pyrite and pyrrhotite. The flat wall adjacent to the mine proper is yellow to rust colored, due to the weathering of the sulfide-rich country rock. Massive minerals orthopyroxene, amphibole, and substantial magnetite can also be observed.

51.9 Take Dunning Trail south and west back to the first intersection of paths. Head south on the first path, to exit back at the gated entrance.

52.7 Drive east on route 106 to Kanawauke Circle, and take the third exit to head north on Seven Lakes Drive.

56.0 Head north on Seven Lakes Drive until reaching Tiorati Circle (lavatory facilities and parking if needed). Take the third exit to head west on Arden Valley Road. ***Lunch & Lavatory Stop***

- 56.8 Drive west on Arden Valley Road for 0.8 miles until a large, cliff-face is visible to the north, roughly thirty feet off of the road. Walk uphill, along the path directly adjacent to the cliff (west-side) for about 200 feet (Stop 5, Figure 13).

Stop 5. Bradley Mine (50 MINUTES) (UTM: 18 T 0575371 4569883). Mine is the northern deposit in the north-western shear zone, located a few hundred meters north of Arden Valley Road (Figure 13). Bradley Mine was active through the latest 19th century, when it closed permanently in 1874. Wall rock is primarily calc-silicate gneiss, containing salite, K-feldspar, apatite, sphene, scapolite, and amphibole. Bradley Mine also hosts a several meter thick discontinuous body of diopside marble, composed of medium-coarse grained calcite, fine-medium diopside, and minor amounts of garnet locally. Diopside is generally dispersed, though locally forms cm-scale aggregates.

The vein deposit at this location appears only a few hundred meters in length and varies in thickness from 2-10 meters. The bleached zone here is characterized by scapolite, diopside, and sericite, followed by narrow zone of layered clinopyroxene, calcite and very fine grained micas. Massive intergrowths of clinopyroxene cemented by calcite, than magnetite, appear in the center of the deposit. Both types of pegmatites can also be observed here. The earliest, white-colored pegmatite appears slightly deformed; containing minor internal fabrics, whereas the latest stage, pink pegmatites appear very coarse grained, and shows no deformational fabrics. A much more recent brittle fault also cross cuts the entrance to the mine proper.

- 53.7 Head west on Arden Valley Road.
56.2 Turn right to stay on Arden Valley Road.
60.1 Turn right at NY-17/Rte-17, heading north.
60.5 Turn right onto the ramp.
61.2 Take the I-87 exit toward Albany.
92.0 Continue toward and merge onto I-87 North.
92.8 Take exit 18 toward New Paltz.
94.2 Turn left at Main St/NY-299.
95.6 Turn left at S Chestnut St/NY-208.
95.9 Turn left at County Rte-17/Jansen Rd.

END OF TRIP

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Stratigraphic and Structural Relationships of the Ordovician Flysch and Molasse along the Western Boundary of the Taconic Allochthon near Kingston NY

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Overview

This trip will visit recently described exposures of the allochthonous Ordovician Normanskill Group juxtaposed against autochthonous Quassaic and Martinsburg sedimentary rocks. Exposures in the Kingston and Esopus Townships demonstrate through stratigraphic position, sedimentary structures and fossils, the collapse of a foredeep basin and subsequent down warping of a foreland basin during the latest stages of the Taconic Orogeny. The trip includes several stops at outcroppings of the allochthon strata, Taconic Unconformity and the later arenites of the Quassaic, which contain an unusual molasse facies.

Introduction

Sandstone petrology of compositions of the strata in the field trip area indicate Ordovician formations originate from a volcanic terrain and are classified as recycled orogen blocks directly or indirectly. It is likely that these formations were formed proximal to one another. A structural inlier consisting of fossiliferous thin shale and siltstone is identified and delineated within unfossiliferous massive arenites. Structural geometry, biostratigraphy and sedimentology were analyzed to constrain the inlier stratigraphic boundaries to the Martinsburg Formation. Massive Ordovician arenites in the study area are uncharacteristic of those of the Normanskill Group, lacking fauna and allocyclic characteristic. Bedding is massive, exceeding 5 meters containing laminate sets and contained greater amounts of quartz and lesser amounts of calcite and lithic fragments as well. Sedimentology of the massive arenites suggests this formation is an alluvial or olistostrome deposit and its structural position would place it above the Normanskill Group. However, analysis distinguishing structural domains potentially places the arenites within the Normanskill Group. Thus, the arenite strata are proposed as a new formation for the uppermost position of the Normanskill Group. The stratigraphic sequence of the Ordovician strata within the study area represents a transgressive progression from an older flysch to a younger molasse facies. This is interpreted as a tectonic depositional loading to a flexural extension of underlying continental blocks.

Geologic Setting

The study area topography is the result of multiple orogenic events and created several notable landmarks in the region. The largest and most prominent landmark is Hussey Hill, which is a northern extension of the Marlboro Mountains and overlooks the low-lying village of Port Ewen to the east. This range of 'mountains' originally named "Marlborough Mountain" (Mather 1843) extends due south approximately 40 kilometers from Connelly to the city of Newburg in Southeastern New York and attains its highest altitude of just over 300 meters at Illinois Mountain. Hussey Hill is formed by a large north plunging syncline consisting of the Quassaic Group, a late-Ordovician arenite sequence (Waines 1986).

Another prominent and more important landmark is Hasbrouck Park, this lies immediately north of the Rondout Creek within the Rondout district of Kingston. It is the southernmost point of a well-exposed, eastward facing steep ridge called North Hill that separates Rondout and points west from the low-lying Ponck Hockie fronting the Hudson River, and extends north approximately 1 kilometer to the Terry survey station of the USGS and attains an average height of 76 meters. The ridge is capped with the folded and faulted late Silurian and Early Devonian Rocks. Mining of this ridge during the latter half of the 19th and early 20th centuries for dolostones and limestones of the late Silurian Rondout Formation, of which the Rosendale Member was the most sought after, have in large part contributed to the ridge's steep and exposed face. Underlying the dolostone and limestone formations are Ordovician silt-

stones and shales of the Austin Glen the Normanskill Group's upper member. The Austin Glen is overturned and thrust between repeating layers of the later Silurian and Devonian limestones. At the base of Hasbrouck Park exposed on the east side are massive quartz arenites. Unlike the Austin Glen, sandstones the arenites consist of massive coarse-grained sandstone light grey-bluish to buff in color and overturned. Also out of character are lenticular conglomerates less than one foot thick of semi to sub angular large to small pebbles of possible Normanskill origin are set into the quartz matrix. These characteristics place the arenites potentially above the Normanskill Group. This carbonate ridge extends south and west crossing the Rondout Creek at Connelly where it meets and overlies the Quassaic Group in angular unconformity to the west of Connelly. The contact continues southwest towards the town of Rosendale and there is buried under postglacial alluvial sand deposits. To the east of North Hill and Hussey Hill are low-lying sand plains that underlie Connelly, Ponck Hockie, and Port Ewen and extend east to the Hudson River. These plains are underlain by highly faulted and folded thrust slices of a medial to Late Ordovician phyllarenite sequence belonging to the highest sequence of the Normanskill Group. The Normanskill Group of rocks extend contiguously south to city of Newburgh and north to the city of Albany and as far east as Pine Plains New York. Locally, the Normanskill sequence lies in angular unconformity with the overlying younger Silurian carbonate sequence and thrust upon and over younger Ordovician Quassaic Group. The angular unconformity and the fault suggest at least two orogenic events took place, the first described by the Normanskill- Quassaic contact, a named fault known as the Esopus Thrust (Dames and Moore 1973) where two Ordovician layers are tilted, overturned and thrust upon themselves. The second is an unconformity located in Hasbrouck Park identified by a Normanskill-Silurian contact. The folded and tilted Silurian beds overly already deformed Ordovician beds, this is typically referred to as the Taconic Unconformity. Another structure of interest lies within the largely overturned thrust sheets of Ordovician arenite beds. Upright beds of fossiliferous thin-bedded sandstones and siltstones are bounded above and below by the massive arenites. The low-lying shales contrast against the ridges of the arenites. This positioning of the inlier also implies it is of Ordovician age. Fossils identified from the shales would indicate an age later than the Normanskill and fauna in kind with ones found within the Martinsburg Group. The Martinsburg inlier is not completely exposed within the Ordovician layers, it is partially submerged under younger Silurian and Devonian strata to the North.

Description of Strata

Normanskill Group

Austin Glen Formation

This formation is the only formation of the Normanskill Group present in the study area and its thickness is estimated at less than 300 meters, based on observed outcrops. Estimates of its thickness are difficult due to the ubiquitous faults, lack of marker beds, and incomplete stratigraphic sequence within the study area.

Beds consists of alternating sequences of laminated and cross-stratified thin grey phyllarenites, calcareous grey to dark-grey siltstones interspersed with thin argillaceous dark-grey siltstones and blue-grey and black shales. Arenite and siltstone beds are typically less than 1 meter thick but can range from several centimeters to two meters. Shale is often lenticular and varies in thickness from a few centimeters to less than a meter. Erosive contacts of the hummocky cross-stratified beds with underlying dark gray shales are common. Occasionally shale and siltstones occur in shallow troughs less than a meter in depth. Much of the beds are bioturbated virtually obliterating nearly all fossil evidence in most of the outcrops observed. Limited shelly fauna was found in thin siltstones and shales in detrital assemblages at Hasbrouck Park at the contact with Silurian beds. Additionally shelly and planktonic fossils are reported to be found at Kingston Point (Howell 1942). They include *Sowerbella sericca* and *Dalmanella testudinaria* brachiopods as well as graptolites; *Corynoides gracilis perungulatus*, *Climactograptus bicornis*, *C. modestus* and *Diplograptus (Orthograptus) acutus*.

Clastic sedimentation and presence of fossils indicates marine depositional facies. Sandstone-shale contacts show abundant ripples, load casts, tool and sole marks indicating active margin deposition. A number of these sedimentary structures were determined to indicate a paleocurrent direction of northeast-southwest with a provenance likely to the east. Furthermore the sequence of interbedded sandstones siltstones and shales describe a synorogenic origin.

Modal analysis of siltstone thin section samples show abundant amounts of quartz (>90%) and lithic fragments (<6%) followed by lesser amounts of plagioclase (<3%), calcite (<1%), mica (<1%) and traces of potassium feldspar as dominant detrital mineral phases in a calcareous matrix. Quartz grains display morphological variation in the form of recrystallization and suturing. The Austin Glen deposits originate from clastic flows resulting from the uplift of older formations. Turbidite sequences are evident in a number of outcrops, sandstone grading into shales that form laminate couplets a few millimeters to a centimeter in thickness. Locally the Austin Glen was restricted to a narrow exposure along the western banks of the Hudson River.

Thrust faults are numerous and no continuous sections larger than 20 meters were found. Orientation of thrust faults would indicate westerly beds are generally younger than those to the east. Beds lie unconformably against younger Ordovician sandstone beds of a proposed formation (Ulster Park) at the western boundaries. To the north, the Austin Glen also lies against the Silurian limestones of the Rondout Formation in angular unconformity. Bedding is upright and occasionally overturned, sometimes bedding is overturned and upright within a single outcrop preventing a definite correlation of beds. Erosive contacts of the hummocky cross-stratified beds with underlying dark-grey shales are common. Regionally the Austin Glen formation is part of the Normanskill Group, which the Austin Glen is the highest stratigraphically (Rudemann 1908). To a greater extent, the Normanskill Group is part of the Taconic Sequence. The overall extent of the Austin Glen member of the Taconic Sequence is contiguous from Southwestern Vermont to the lower Hudson Valley of New York (Zen 1967, Rowely and Kidd 1981).

Dominant bedding is moderately to steeply east dipping, homoclinal and generally strikes 000-030 degrees. Much evidence of tectonism is found in the Austin Glen Formation. Recumbent to moderately inclined folds and bedding parallel reverse faults are common. Reverse faults often cut through the recumbent fold axis and then are also subsequently faulted again into smaller segmented thrusts or slices. Brittle shear zones occur in reverse faulted shale and siltstone layers. En echelon tension fractures or gashes often forming sigmoidal vein filled fractures are common. Shear zone rock fragments exhibit brecciation and crushing. Slickensides evident on sandstone bedding planes and are often in-filled with quartz and calcite. Small thrust segments can be less than 25 feet in length though most are greater in length but could not be determined in outcrops due to limited exposure. The reverse faulting occurs so frequently that discerning a stratigraphic sequence and relative thickness is difficult. Adding to the already difficult sorting of the slice segments the younging direction occasionally flip flop on account of the cut off recumbent folds. This was evident in outcrops which beds would be overturned in the footwall of a fault and upright in hanging wall.

Occasional high angle normal faults were found and often at high angle and oblique to bedding planes. Multiple shale and sandstone cleavage planes often oblique to structural trend are common. Shale and siltstone cleavage is common on two planes and the shale exhibits penciling in a few outcrops. Sandstone layers demonstrate fracture cleavage limited to one plane predominately trending in a northeast southwest direction, limited to Hasbrouck Park. Orthogonal jointing and plumose structures are common in sandstone beds. Cleavage planes exist in shales and more competent siltstones orientations are NNE in shales and NE in siltstones. They are similar to those indicative with the foliation structures in the Taconic allochthon seen elsewhere and agree with Zen (1967), Rowely and Kidd (1981), Stanley and Ratcliff (1985), and others.

Ulster Park (Proposed)

The stratum named the Ulster Park is proposed as the upper formation of the Normanskill Group. Within the Kingston and Rondout areas, it lies juxtaposed as a fault contact against the Austin Glen, the contact is exposed at the base of Hasbrouck Park extending west to the base of the Quassaic where it rests in angular unconformity. The formation continues south of the study area and outcrop in the towns of Ulster Park and Lloyd (Cunningham 1990). The formation lies completely to the west of the Austin Glen. Thus, based on the assumption its placement follows the presently accepted tectonic model of thrust slice emplacement, where the leading edge of younger slices were emplaced prior to later aged slices which still mobile, over rode the trailing edge of the younger aged slice, (Stevens 1970) thus the proposed formation is presumed younger than the Austin Glen. Homoclinic with bedding largely overturned and moderately to steeply dipping, total strata thickness is reported to range from 1000 to 3400 meters (Cunningham 1990). A far less thickness was found in the study area, likely represents a partial stratigraphic representation of the formation. The formation as reported is described as having brown shales as its lowest member, overlain by increasing intermittent laminated sandstones and siltstones and lesser shale layers (Cunningham 1990). Sand-

stones increase up-section in frequency and thickness, often exceeding 5 meters and weather brown, black and rust yellow. Sandstones include lenticular conglomerates of limestone and clays. In some instances, the clays weather out of sandstones creating pocks in jointing planes. Fossils reported include several genera of trilobites, brachiopods, asterozoa, bryozoa, ostrozoa and crinozoa of the Middle Ordovician (Kirkfieldian) from two locations. Fossils of shelly fauna exist sporadically in shale layers. Initial petrographic analyses seem to indicate a significant sodic-rich plagioclase and K-bentonite content. The feldspar and other mineral grains are detrital and subangular, thus may represent an eastern provenance and orogenic terrane. Numerous shallow turbidites less than 0.5 meter deep, ripple marks, and lode casts exist throughout the sequence. The entire sequence represents a transgressive sequence. The lower beds contain calcareous muddy silt, an offshore relatively low sediment influx rate, which was the favorable environment to sustain the abundant shelly fauna. The silt and muds rapidly gave way to thick coarse sands from a upslope high sediment influx environment represented by the numerous turbidity structures. A type section within the township of Ulster Park was established (Cunningham 1990).

Outcrops within the study area absent of the fossiliferous shales and are predominately massive quartz arenite in most exposures, thickness of beds varies from <1 to 5 meters. Minor dark shale and siltstone beds vary in thickness from several centimeters to two meters. Color ranges from light to buff in fresh exposures to orange brown to black in weathered exposures. Erosive contacts of the hummocky crossstratified beds with underlying dark gray shales are common. Beds are often graded or cross graded, faint laminae sets are also present and vary from 5 mm to a centimeter in thickness. Laminae sets vary little in exposure, often absent of cross set pairs or truncated pairs make it difficult to determine younging direction. Found in outcrops located east of Esopus Fault and west of Normanskill beds bordering the Hudson River. A large outcrop near the base of the Wurt Avenue bridge location yielded 250 meters of exposed section; with a representative dip of 50 degrees, this gives a thickness of about 100 meters. Several other poor outcrops serve to identify the extent of the beds laterally but without any additional discernable thickness. Total thickness 300-400 meters, however, may be more representative of the unit within the study area. The arenite beds become thicker and more abundant to the east presumably higher in section. Ulster Park is similar in structural complexity to the Austin Glen but varies slightly in foliation, generally striking 010-050 degrees, beds dip east moderately (30 degrees) to steeply (>45 degrees) in outcrop. Faulting is ubiquitous; numerous reverse faults that follow structural trend of beds are common. As found in the Austin Glen, reverse faults often cut through fold axes or are subsequently faulted again into smaller thrusts or slices. Normal faulting occurs occasionally in larger outcrops, strike of these faults are approximately 060 and often at high dip. Faulting bounds the upper and lower contacts of the formation so that discerning a stratigraphic sequence and thickness is difficult. Outcrops near the contacts with Silurian and younger Ordovician beds are nearly devoid of shales as opposed to more frequent layers away from contacts. It is likely that there is an increase in shear stress as beds near the contact thus during deformation and less competent shales were squeezed out leaving behind arenites. Where observed cleavage planes orientations in shales is NNE. Generally this structural trend coincides with the Normanskill Group. A synclinal fold with axial plane striking 075 degrees was observed at one location along Connelly Rd. The orientation of this fold is not typical to those generally associated with Taconic orogeny, which folding and faulting occur generally north south. Folds are symmetric and fold axis is oblique to strike and plunge north. The fold orientation and intact structure suggest these folds are post Taconic in age. The fold orientation follows similar trends to those folds in Silurian strata and is more likely that these folds are associated with the later Acadian orogen. The Acadian effects are often seen to the south west of the Hudson Valley demonstrated in the folding of the Silurian Shawangunk Conglomerate and rise of the Catskill Mountains. Therefore, while finding these folds was surprising they were not completely unexpected.

Martinsburg Formation

The Martinsburg Formation is a mapped bedrock unit with extents in, New York, Pennsylvania, New Jersey, Maryland, Virginia and West Virginia. It is named for the town of Martinsburg, West Virginia for which it was first described. The Upper Ordovician Martinsburg Formation of eastern Pennsylvania consists of mudstone, siltstone, and sandstone turbidites, which are also seen along the western edge of the study area. The beds are gray to dark gray, and infrequently tan and purple shale and slate. Localized sandstone, thin, argillaceous limestone or phyllitic shales are present. The mudstone-rich Bushkill Member, the stratigraphically lowest unit of the Martinsburg exposed to the south and west of the study area in the townships of New Paltz and Rosendale, grades upward into approximately equal proportions of mudstone, siltstone, and sandstone of the Ramseyburg Member, which closely coincides with

observations made at a number of locations locally. Outcrops are generally upright though some overturned beds were found. Disconformities of the hummocky sandy beds with underlying dark grey shales are common, and ripple marks are infrequently developed on these beds. Individual cross beds reach a thickness of 30 to 50 cm. Several Pebble to cobble sized (10 cm), rounded mudstone clasts are found at locations with similar lithofacies. Bound structurally top and bottom by Ulster Park Formation by faulting; thus the stratigraphy is incomplete. The Bushkill-Ramseyburg sequence is reported to be approximately 2000 m thick, though its total section in the study area is approximately 40 m. Thus represents a partial section of the formation. The Martinsburg lies above and below erosional unconformities regionally. The Shawangunk Formation (New York), lie unconformably atop the Martinsburg. Below it, the Middle Ordovician Chambersburg and Myerstown Formations (Pennsylvania) lie in unconformity and it is unknown what lies below the Martinsburg Formation locally. The following descriptions represent a partial sequence of the formation. Bedding is generally homoclinic, striking north to northeast and dipping east low to moderately, 020 to 040 degrees. Undulating bedding and sedimentary structures are common as well as shallow turbidites typically fining-upward. Hummocky character of beds sedimentary structures are interpreted as reflecting the influence of external or allocyclic controls such as variations in the intensity of tectonic activity in shelf/slope areas, thus exhibit a flysch facies. In the study area, the Martinsburg exists as an inlier within the Ulster Park Formation of slightly older if not similar age. It is likely the turbidites are caused by underwater landslides stirring up sediments that accumulated on a slope along the periphery of a tectonically active foreland basin. The variation of argillaceous and limestone beds within calcareous matrix, these variations reflect the local rate of sea-level rise and/or variations in the intensity of tectonic activity in shelf/near shore or hinterland areas from a deepening trough or formed upon peripheral bulges due to the closing of a sea rather than more commonly cited allocyclic mechanisms related to terrestrial events as subsidence and deltaic flooding (Lash 1988).

Quassaic Group

The Quassaic Quartzite was first named on the Geologic Map of New York State (Fisher 1971) and later described as a massive pink and green quartzite, sandstone with occasional conglomerates, quartzites, conglomerates having arkosic, peletic, occasional red argillaceous matrix. Beds grade upward into green-grey sandstones with tabular cross-lamination common and few green-grey shale interbeds. The red matrix is debris from erosional shoals of older material is transported to an upper slope environment. This matrix is best described as a molasse (Fisher 1977). Later work by Waines (1986) raised the Quassaic to Group status. The Quassaic Group is a late Medial to medial Late Ordovician marine arenite, 3000 meters in thickness. It is subdivided (oldest to youngest) into five formations: Creek Locks, Rifton, Shaupeneak, Slab Sides, and Chodikee. General lack of shales, and scarcity of fossils indicate that the average depositional environment of the formations was near the base of the slope of a sedimentary apron or delta. The Quassaic Group is at a stratigraphic high point or erosional shoal thus no known unit overlies it. It is similar or equivalent in age to the Bushkill Member of the Martinsburg Formation. Thus likely represents a transition point in the closure of the foredeep basin.

Stratigraphy

Folded siliclastic quartzite, mostly upright and occasionally overturned. All outcrops generally occur in the western portion of study area along a ridge line of hills noted as the Marlborough Mountains. Beds on western side of hills grade conformably into younger late Ordovician Martinsburg Formation shales (Waines 1986). Beds on eastern side of the Marlboro Mountains are truncated by a fault and lie in angular unconformity with older sandstones and shales of Ulster Park and Normanskill Group and the younger Devonian limestones of the Helderberg Group.

Structure

The Quassaic Group bedding generally strikes 010-030 degrees and forms the most intact structure within the study area. The Quassaic strata make up a large asymmetric syncline (Waines, 1983) which is continuous along strike for at least 40 km (Cunningham 1991). Eastern limb beds dip east moderately to steeply 40 to 90 degrees or are occasionally overturned, western limb beds dip west steeply moderately (>45 degrees) in outcrops. The eastern boundary of the Quassaic is bounded by The Esopus Thrust fault (Dames and Moore 1973) which was traced from Newburgh north into the township of Esopus but stopped south of the study area. Within the study area this fault is mostly covered except for a few outcrops along the railroad cut. The Ulster Park Formation abuts this fault on its western

extents. Foliation of strata indicates the syncline plunges north and under younger Devonian beds thus indicating at least two tectonic events. Rarely, faulting occurs as normal or strike slip faults oblique to strike often at high angle. Slumping of bedding at the Stop 12 location indicates a molasse facies. This outcrop, located at the most northern exposure of the Quassaic, consists of a 200m ridge that decreases in elevation from north to south 30m. The ridge is mostly covered in thin soil with little exposed bedrock outcrop. The northern most outcrop of the Formation was quarried historically, at this area it appears there is a significant slumping of beds in which thin 30 degrees west dipping sandstones to the east side of the outcrop are truncated by near vertical red conglomerate and thick bedded sands. An upslope outcrop to the south and west exposes bedding as thin sands oriented 30 degrees west dipping, similar to that found on the east side of the slump. While no west contact of the block was found, it is inferred that this is evidence of a molasse facies. Further study of this outcrop site would likely produce interesting findings.

Field Trip Stops

Note: There is significant rock scrambling and hiking so sturdy hiking shoes are recommended. Trails are typically overgrown and long pants and shirts are also recommended.

Stop 1. Rotary Park at Kinston Point. Within the park are several outcrops of Austin Glen lithology, which generally dip 55 degrees east. Thin to thick beds of alternating sequences of laminated and cross-stratified thin grey phyl-larenites, calcareous grey to dark-grey siltstones interspersed with thin argillaceous dark-grey siltstones and blue-grey and black shales. Arenite and siltstone beds are typically less than 1 meter thick but can range from several centimeters to two meters. Shale is often lenticular and varies in thickness from a few centimeters to less than a meter. Erosive contacts of the hummocky cross-stratified beds with underlying dark gray shales are common. Shale and siltstone cleavage is common on two planes orientations are NNE in shales exhibiting penciling and NE in siltstones in a few outcrops at the park. Orthogonal jointing and plumose structures are common in arenite beds. Limited detrital fauna deposits are found in the shale and siltstone layers due to bioturbation.

Stop 2. Hasbrouck Park. A path off the parking area continues along the base of an escarpment of Helderberg limestones. The path continues along the southwest side of a large vertical adit runs northeast southwest. The escarpment was extensively mined in the past for the dolostones of the Rondout Formation to make cement. Carefully continue past the adit and scramble down off the path to a cut in the slope. Here the Austin Glen is juxtaposed against the remnants of the Wilbur Limestone member of the Rondout Formation in angular unconformity.

Stop 3. Trolley Museum. Take a few minutes and peruse the museum. Follow the railroad cut west along base of Hasbrouck Park, Outcrop is exposed along its entire length. Massive sandstone layers include some lenticular conglomerates of chert and limestone. Clasts of semi-angular large and small pebble-sized black and green chert, grey shale and weathered laminated light to dark grey limestone. Approximately 100' along the railroad bed from the museum the rock face gives way to a slope. Scramble up the slope to an abandoned road. Walk east along the road to the end of the chain link fence approximately 75'. Scramble up the slope toward a large rock fall. A section is cut out of the slope perpendicular to the mine adit. Here the upper Normanskill is in contact with Silurian or Devonian Limestone. The orientation of the unconformity has changed from a NNE strike to NE. This sight is also accessible from Hasbrouck Park. Beginning at the pavilion in the park, walk north along a sidewalk path toward an abandoned block building, continue through a break in the fence behind the building and scramble down a steep path through the talus. The ground levels and the contact is on the left approximately 50 feet from the base of the slope.

Stop 4. West Strand Street. Extensive exposure approximately 250 long at base of Wurts Ave Bridge exhibits large amounts of deformation and a few folds plunging northeast. Beds are both upright and overturned, generally striking N40W and dipping 50E. A number of low to high angle faults strike NE throughout the section. Orthogonal joints exhibiting plumose structures are common. Turbidite structures are used to determine overturned or upright sections of the outcrop. Pock marks eroded from joint faces generally run along bedding planes and are thought to be of soft mudstone pockets within the otherwise massive sandstone.

Stop 5. Slaughterhouse. Just West of Stop 4 this outcrop was recently uncovered after demolition of a large abandoned structure once used as a slaughterhouse. Massive quartz arenite in most of the exposure, thickness of beds exceeds 5 meters. Minor dark shale and siltstone beds vary in thickness from several centimeters to two meters. Cleav-

age planes orientations in shales is NNE. Color of arenites range from light to buff in fresh exposures to orange brown to black in weathered exposures. Beds are often graded or cross-stratified, faint laminae sets are also present and vary from 5 mm to a centimeter in thickness. A high angle normal fault occurs in the exposure, strike of this fault is approximately 060. Faulting bounds the upper and lower contacts of Street borders the upper part of the exposure on the South side. Bordering the north side of Abeel Street is an outcrop of Devonian limestone outcrop in apparent unconformity. Arenite beds near contact are nearly devoid of shales as opposed to more frequent layers away from contacts seen to the east. It is likely that there is an increase in shear stress as beds near the contact thus during deformation and less competent shales were squeezed out leaving behind arenites.

Stop 6. Apparent Taconic Unconformity. Top side of the Stop 5 outcrop, this part of the exposure demonstrates an apparent contact with the Devonian beds of the Coeyman limestone formation directly across the street.

Stop 7. Old 9W Roadcut. The proposed Ulster Park formation in the extensive road cut is exposed. Massive quartz arenites dominate the cut. Bedding displays graded and cross bedding and is overturned. The few thin shales exposed are cleaved in at least two directions. At least two major faults cut through the outcrop and is traceable across the highway. A high angle fault dominates the outcrop, its displacement effectively dissects the exposure into two distinguishable and different parts. Some shale and limestone pebble conglomerates are evident. Several lode casts are evident on bedding planes.

Stop 8. Connelly Road residence. Large outcrop behind residence exhibits plunging fold oriented N75W. This orientation is not typical of Taconian deformations that follow a NNE trend.

Stop 9. Intersection of County Route 25 and Millbrook Road. Locally fault-bounded above and below by upper massive quartz arenites and composed of low-to moderately-eastward dipping, 020 to 040 degrees, faulted and folded, medium to dark-grey sandstone and interbedded shale. Total section represented in this exposure is approximately 40 m. Dominate dark-grey to black mudstone with a few lenticular fine-grained, medium grey, sandstone turbidites, calcareous with mudstone clasts. Many of the turbidites observed are arranged in small-scale (1-2 m) fining-upward sequences demonstrating bedding is upright in position. Calcareous shelly fossils preserved primarily as molds and casts. Shelly fossils occur throughout the shale and sandstone with locally dense concentrations. There were several fossil localities along Millbrook Rd. Fossil localities chiefly contained disarticulated brachiopods, ostracods, crinoids, bryozoans, and trilobites. Genera and limited species of brachiopods found were identified as *Dalmanella sp.*, *Paucicrura rogata sp.*, *Sowerbyella spp.* and *Rhynchonella sp.* articulate brachiopods, *Dilobella spp.* ostracod, trepostomate bryozoa zoaria, and a pygidium belonging to a trilobite tentatively identified as *Decoroproetus spp.* was found.

Stop 10. This outcrop contains calcareous thin-bedded siltstones and shales. Bedding is dark grey to black, and is moderately dipping east 50 degrees. These strata are similar in composition to Stop 9, though the fossils are absent.

Stop 11. East side of railroad cut is massive sandstone bedding overlooking extensive sand cover to the southeast.

Stop12. Old quarry. This quarry lies the opposite side of the railroad cut west of Stop11. Here the Esopus Fault is presumed to lie within the railroad cut and plunges under the Silurian and Devonian limestone immediately north of the stop. The fault extends to the south with rocks of the upper Ordovician Quassaic Group sitting to the west of the railroad. Thin bedded sandstones are truncated by massive conglomerates of red clay clasts of pebble to cobble size. The bedding generally strikes N30 degrees dip east moderately 40 degrees. Slumping of bedding at the location indicates a molasse facies. An upslope outcrop to the south and west exposes bedding as thin sands oriented 30 degrees west dipping, similar to that found on the east side of the slump indicating a defined block of red clasts slumped into the sands.

Stop13. Railroad cut at County Route 25 and West Shore Railroad line. Here again lies the presumed exposure of the Esopus fault. Rocks on the west side of the railroad cut are massive steeply dipping and medium to thick quartz arenites and conglomerates. Conglomerates contain of rounded or subangular pink or grey (rarely banded) limestone and black or green chert and shale pebbles. In the railroad cut, the basal Shaupeneak Formation of the Quassaic group is exposed. To the east of this extend the proposed Ulster Park rocks of the upper Normanskill Group. They

are seen in numerous outcrops along the backstreets of Port Ewen and form low-lying ridges extending north to south.

Stop 14. Hussey Hill. A moderate climb up a steep slope is required to access this stop. Along the slope path are exposures of red clasts of the Shaupeneak. At the top are exposures of the Slabsides Formation of the Quassaic Group are found. They mostly occur as vertical beds of thin to medium quartz arenites. The Quassaic Group makes up the Hussey Hill at its northern point and extends south from the north to the Marlboro Mountains.

Road Log

Total Miles: Miles from last Stop

0 0 Start: Kingston Point, Battery Park Kingston N.Y. Parking lot. Looking North you will see North Hill a ridge running North South. It consists of deformed Helderburg Limestones resting on the Ordovician Normanskill strata.

Stop1. Enter Park through iron gate. Outcrop immediately along dirt path on right just after entering park. Cleavage face of Austin Glen formation. Continue south along dirt path to outcrop. Crosssection of angular beds of Austin Glen formation consisting of alternating shales and siltstones. Siltstones substantially bioturbated. Continue along path to southern peninsula of park. Large outcrop of sandstones, siltstones and shales.

- 0 0.01 Turn right out of parking lot onto Delaware Ave. Ascend North Hill and take note of , then turn left onto Hasbrouck Park Rd.
- 0.7 0.7 Take immediate left into parking area. Hike 1/4 mile into **Stop2.**
- 0.71 0.01 Take right out of parking area to end of Hasbrouck Park Rd. Turn right on Delaware Ave.
- 0.72 0.2 Turn right at Abruyn St. though the hamlet of Ponck Hockie.
- 0.92 0.3 Turn right at East Strand St. A good view of the North Hill ridge to the right and Rondout Creek on the left.
- 1.22 0.6 Turn right into the trolley museum.

Stop 3. The Trolley Museum is part of and adjacent to an active railroad switchyard and permission is needed to access this location. The outcrop is located behind the museum buildings along the base of North Hill. This is the northern extents of North Hill and is truncated by Rondout Creek. The outcrop consists of massive arenites and thin lenticular conglomerated of the proposed upper formation of the Normanskill formation.

- 1.82 Turn right out of Trolley Museum at East Strand St. to Broadway
- 1.82 0.1 Turn right at Broadway
- 1.92 0.05 Turn left at West Strand St., parking area will be on left. Across the street is the road cut **Stop 4.**
- 1.97 0.05 Walk west down West Strand St. to **Stop 5.** Outcrop set back on north side of road. Walk back to car and turn left (West) out of parking lot on West Strand St.
- 2.02 0.1 Turn right at Abeel St.
- 2.12 0.05 Park in lot on north side of Abeel St. **Stop 6** outcrop is at back of lot. When leaving turn left from parking lot.
- 2.17 0.4 Turn right at North Broadway over bridge. Pull into industrial pipe supply company and park. Permission is needed. Note, the bridge has a pedestrian walkway and provides an excellent vantage point to observe several field trip stops are visible (2, 3, 4, 5, 12) . Several other outcrops and structures of interest not covered in the field trip are also prominent.
- 2.57 Walk north along road. Caution must be used when walking along roadway. **Stop 7** outcrop is both sides of the road. Return to cars and turn left (north) out of lot.
- 2.57 0.1 Turn left at Connelly Rd
- 2.67 0.2 Turn left at Connelly Rd residence. **Stop 8.** Large outcrop in back yard. Permission is needed to enter location. Return to Car, turn left out of driveway.
- 2.87 0.4 Turn left at Millbrook Dr.
- 3.27 0.6 Turn left into gravel lot.

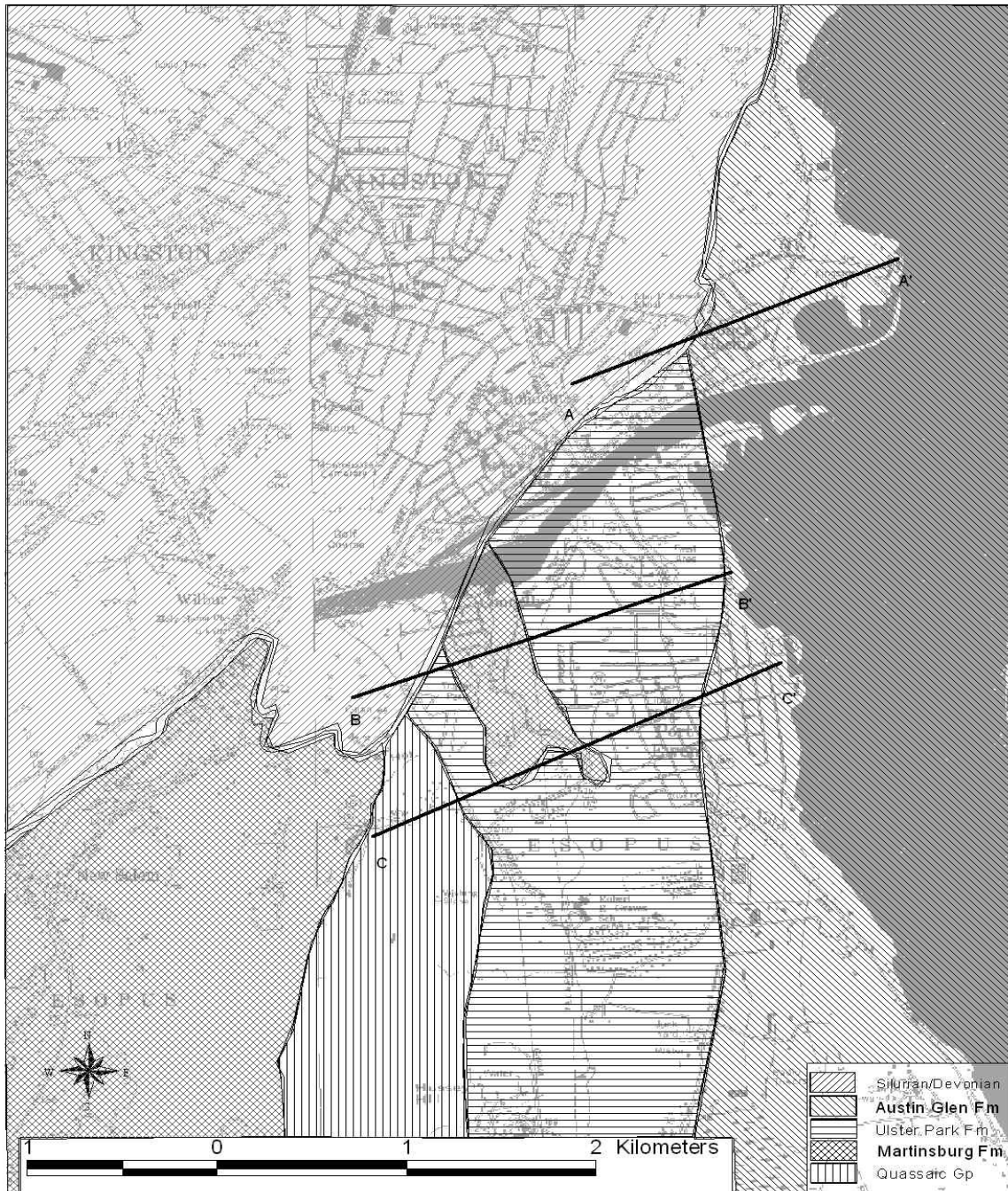


Figure 1.

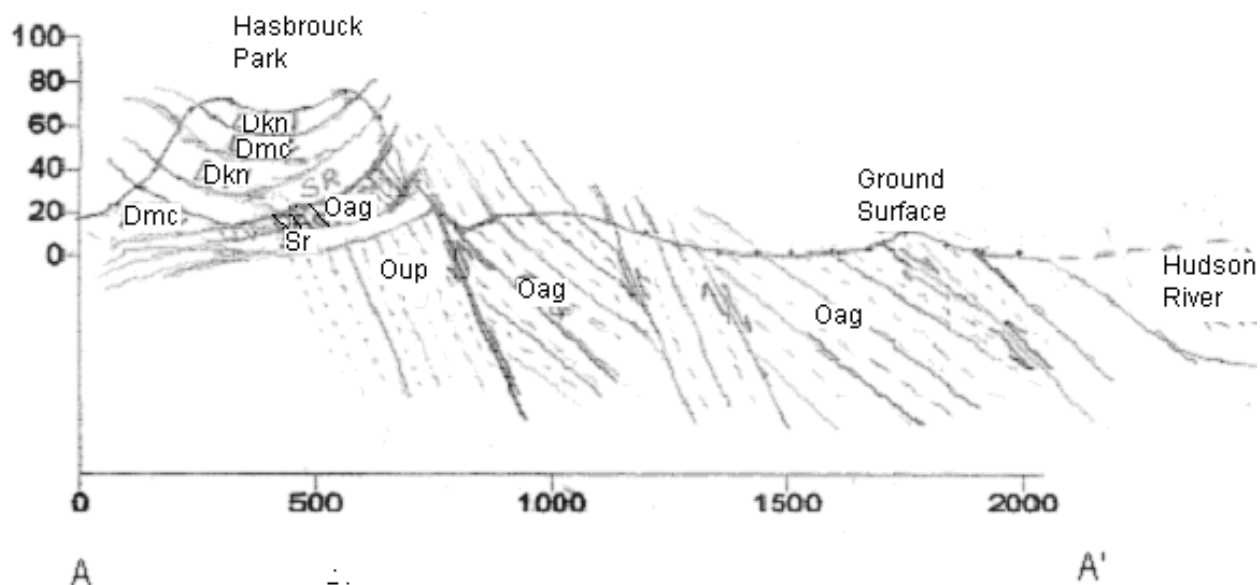


Figure 2. Units = meters. Cross sections x4 vertical exaggeration.

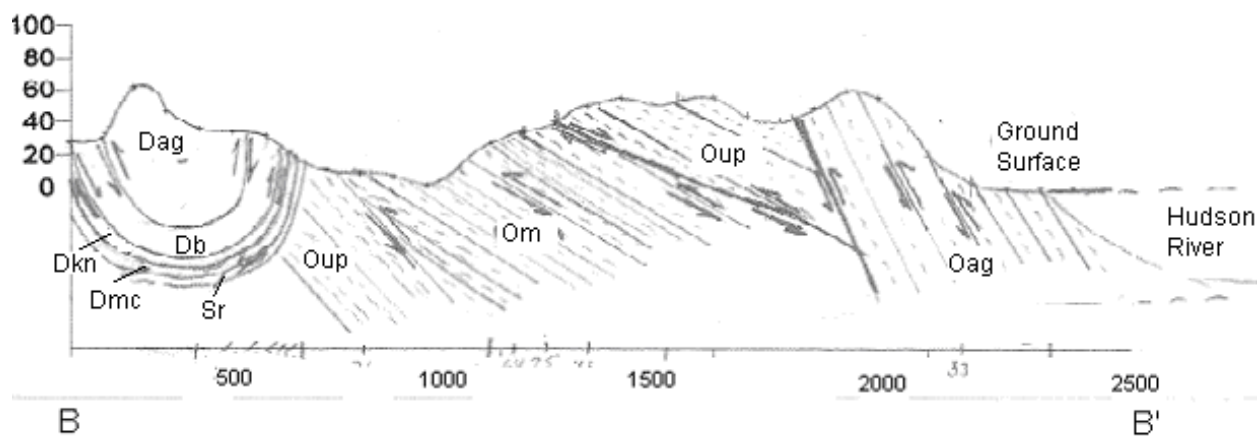


Figure 3. Units = meters. Cross sections x4 vertical exaggeration.

Stop 9 outcrop at back of lot. Facing east and up gradient along Salem St. a contact of Martinsburg with upper Normanskill strata is evident. Facing north, the outcrop continues for 0.76 miles along Millbrook Rd and within the stream gully that runs beside the road. The Martinsburg strata continues until it is truncated by the Rondout Creek. Facing west across the stream gully a steep slope rises, where outcrop of upper Normanskill is found. Leave lot north on Millbrook Dr.

- 3.87 0.3 North on Millbrook to right at Andorn Rd
- 4.17 0.1 Pull off road along shoulder at 1st culvert crossing. **Stop 10.** outcrop is northeast stream bank. Turn around.
- 4.27 0.1 Turn left at Millbrook Dr.
- 4.37 0.3 Turn left at Marys Ave.
- 4.67 0.1 Turn left at James St.

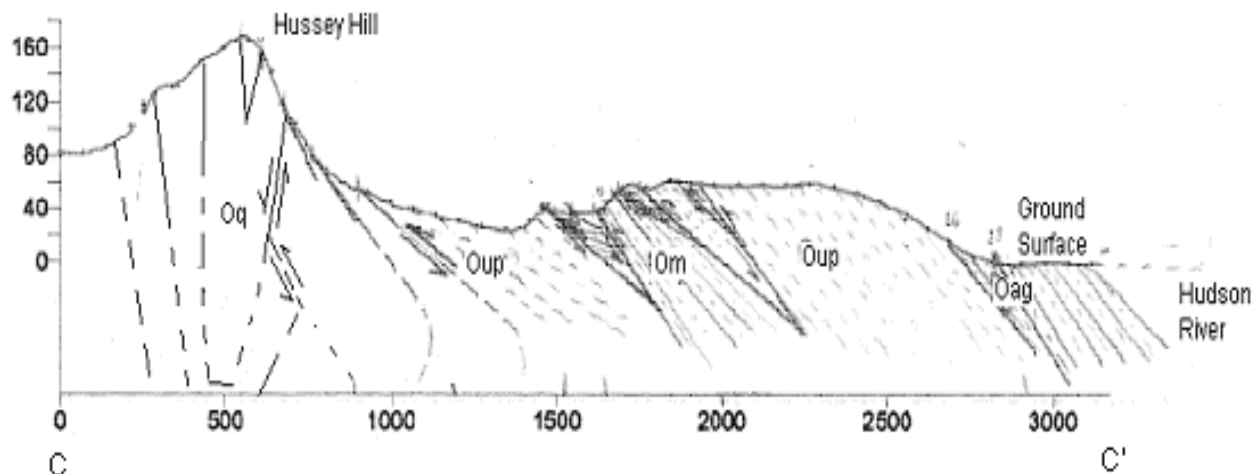


Figure 4. Units = meters. Cross sections x4 vertical exaggeration.

- | | | |
|------|------|--|
| 4.77 | 0.1 | Turn left at Florida St. |
| 4.87 | 0.3 | Arrive at gravel lot on left between mobile homes. Walk behind mobile home to the right along ATV path about 200 feet, Stop 11 . outcrop is low lying and may be difficult to locate in high vegetation. Continue along ATV path 300 feet toward railroad bed. Cross railroad bed and walk north west (left of path) 300 feet. Enter wooded area to right and down embankment. Stop 12 , outcrop is on south side of large gully. Return to cars and back down hill. |
| 5.17 | 0.3 | Turn right at James St |
| 5.47 | 0.1 | Turn right at Marys Ave. |
| 5.57 | 0.1 | Turn right at Millbrook Dr. |
| 5.67 | 0.6 | Turn right at Salem St. |
| 6.27 | 0.4 | Turn left at Station Rd. |
| 6.67 | 0.05 | Turn into Town Highway Maintenance area. Park in an area clear of any ongoing work or town vehicles. Walk to railroad cut. Stop 13 . outcrop both sides of railroad. |
| 6.72 | 0.25 | Walk back toward town maintenance area. Cross street opposite garage, to ATV path ascending slope of hill. Along ascent of path, note several outcrops of red clasts along path. Stop 14 outcrop along north south ridge of hill. Return to car. |
| 6.97 | 0.05 | Turn right at Salem St. |
| 7.02 | 0.9 | Turn right at Broadway/ 9W, through the town of Port Ewen. Return to SUNY New Paltz |

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Deglaciation in the Southeastern Laurentide Sector and the Hudson Valley – 15,000 Years of Vegetational and Climate History

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Introduction

In this field trip, we provide a review of the significant controversy concerning the timing of deglaciation in the Hudson and Wallkill Valleys. We outline the differences in methodology and chronology with a circular route throughout the Hudson and Wallkill valleys. We begin the trip at Lake Mohonk near New Paltz led by Kirsten Menking and Dorothy Peteet, then continue to the “black dirt” region of the Wallkill Valley where John Rayburn has contributed a new GIS model of deglaciation in the Wallkill Valley and Guy Robinson will review the history of fossil mammals, including mammoths. From this point we travel southeast to a rare exposure of glaciolacustrine beds on the west side of the Hudson River, described by Byron Stone and John Rayburn, and on to Croton Marsh at Croton Point, New York where Dorothy Peteet will review the marsh histories of the region.

A recent review of literature relating to the last glacial recession in the Hudson Valley indicates that the timing of deglaciation is very controversial (Peteet et al., 2006; Peteet, in review; Balco et al., 2006; Balco et al., 2009; Schaefer, 2007). Some questions to consider:

- 1) How does timing of new lake basal dates at the margin of the ice (Staten Island) compare with sites to the north and inland (ie. Mohonk)?
- 2) What is the vegetational history of the region and how does it compare with Deevey’s classical southern New England stratigraphy?
- 3) What is the latest model of the deglaciation of the Wallkill Valley?
- 4) What have the Hudson marshes added to our understanding of the vegetation and landscape history, particularly in the last few millennia?

I. Review of Hudson Valley regional glacial history – a continuing controversy

Bulk C-14 Ages

Muller and Calkin (1993) summarized the deglaciation history of southern New York, using bulk chronologies to indicate deglaciation by 18,750 ¹⁴C (23 -24 kyr). However, these authors remarked on the weak chronostratigraphic framework for the region. From the New Hampton Bog near Wallkill, NY, Connally and Sirkin (1986) argued for ice retreat at 17.2 ¹⁴C (20 kyr) based on an extrapolated age for basal inorganic sediments. These earliest glacial studies in the region (Sirken, 1982; Connally and Sirkin, 1986) argued for general deglaciation approximately 18 ¹⁴C (21-23 kyr), again calculated from extrapolated sedimentation rates below bulk ¹⁴C dates. In contrast, from bulk-dated sites near the glacial margin in Pennsylvania, Crowl (1986) argued for a somewhat younger glaciation (15 ¹⁴C (18-18.5 kyr) but Connally and Sirken (1986) continued to argue for the older timing for ice recession based on Cotter’s (1983) 18.6 ¹⁴C (23.3 kyr) date from Francis Lake and other bulk dates. The detailed surficial geologic maps of adjacent northern New Jersey (Stone, Stanford, and Witte, 2002) and Connecticut and Long Island Sound basin (Stone and others, 2005) incorporated these older postglacial dates as constraining ages on the early deglaciation of the region. The duration of the early glacial recession in New Jersey and Connecticut further was correlated with detailed stratigraphic frameworks that encompass recessional moraines, voluminous deposits of numerous glacial lakes in all of the valleys, and successive ice-marginal deltaic morphosequences that trace the recession of the ice margin in each glacial lake. These maps also utilize the thickness and known/inferred number of glacial varves in the major

glacial lakes, and the dated postglacial deposits in the valleys to estimate the date of ice recession. This older deglaciation chronology has been the accepted version of deglaciation that glacial stratigraphers, and varve and cosmogenic investigators have utilized.

AMS C-14 Ages

In contrast to this relatively early age for deglaciation, more recent AMS dating of basal sediments from lake/bog sites suggest a substantially younger age for initial accumulation of postglacial sediments, and, perhaps, deglaciation throughout at 12.5 ^{14}C (15 kyr) (Peteet et al., 1990; 1993; 2006; Peteet et al., in review). A suite of 10 lakes ranging from Linsley Pond, Conn. to Uttertown Bog in western NJ indicate a remarkably consistent pattern of tundra and/or spruce colonization that followed ice retreat throughout the region. Uncertainties in using basal-sediment dates and earliest pollen-succession chronologies for deglaciation include whether ice blocks remained in lakes and/or whether a significant migration lag could have taken place.

Cosmogenic Ages

The oldest ^{10}Be ages from Martha's Vineyard indicate a maximum glacial extent age of about 23.2 kyr, and formation of Cape Cod recessional moraine complex forming 18.8 kyr (Balco et al., 2002). The initiation of the deglaciation represented by the eastern Connecticut moraines occurred about the same time as Cape Cod, at 18.5-19 kyr. ^{10}Be dating on glacial erratics and bedrock from NYC is interpreted to indicate deglaciation about 18 kyr (Schaefer et al., 2007). However, more recently, Balco et al. 2009 utilized varve chronologies from New England to recalculate the deglacial age for the terminal moraine in New Jersey at about 25 kyr. Uncertainty in the ^{10}Be ages lies in production rate assumptions, geomorphic stabilities of boulders and outcrops, and the potential for inheritance in the erratics dated.

Varve Ages

The confirmation and rebirth of the New England varve chronology by Ridge et al. (2004) has resulted in a deglaciation scenario with an early chronology of deglaciation defined as 20-25 ^{14}C (23.7-28 kyr). Ridge has added floating varve chronologies from the New Jersey-New York-Connecticut region to the existing dated chronology further north in Vermont and derives this relatively old age which is more in agreement with bulk ^{14}C chronology but is 8-12 kyr older than the AMS chronology.

II. Mohonk region site descriptions and vegetation and climate history

Site 1 – Lakes Mohonk and Minnewaska

Lakes Minnewaska and Mohonk lie atop the Shawangunk Mountains in southeastern New York, parallel ridges of Silurian-age quartz pebble conglomerates and quartz sandstones (Shawangunk Formation) that are part of the southeasternmost ridge of the Appalachian Mountains (Bernet et al., 2007; Macchiaroli,

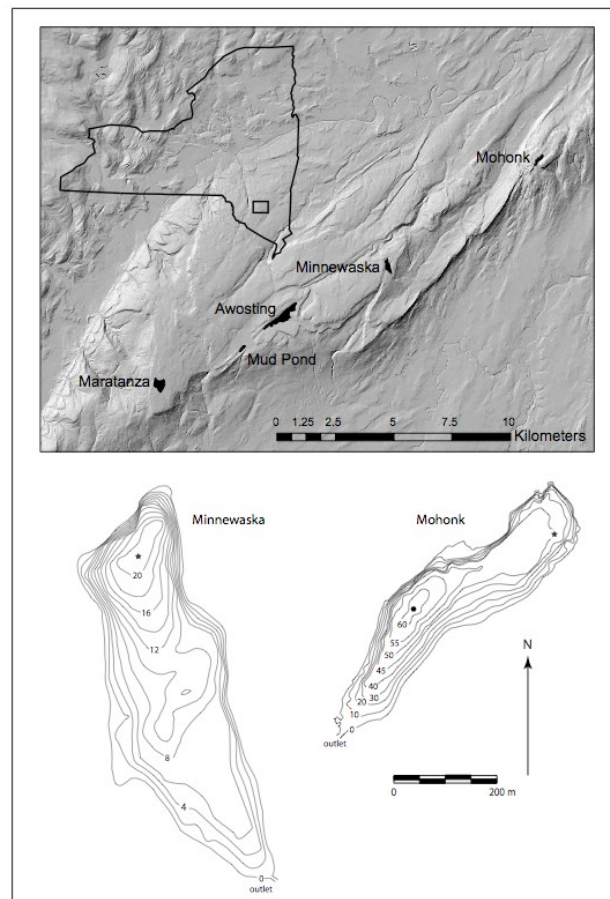


Figure 1. Map of Shawangunk ridge and bathymetry of Mohonk and Minnewaska.

1995). During the late Pleistocene, ice from the Hudson-Champlain lobe of the Laurentide ice sheet covered the mountains to a depth of ~600 m, and many glacially polished and striated surfaces are found throughout the range. Five lakes (collectively called the Sky Lakes), produced by glacial plucking and deepening of tectonically fractured bedrock, lie in a NE-SW trending line along the ridge crest (Figure 1).

None of the lakes have inflow streams, thus all are fed primarily by precipitation falling on the range, with very minor groundwater inflow (Coates and others, 1994). The lack of inflow streams to each lake means that sedimentation rates are low and that deposition is dominated by the annual cycle of deciduous trees and in situ aquatic production. Modern vegetation in the Shawangunks includes numerous pitch pine barrens, dwarf pine plains, and a variety of deciduous hardwood and mixed deciduous/conifer communities (Kiviat, 1988). Vegetation type is controlled largely by local variations in glacial till thickness, wind stress, and slope.

Methods

Sediment cores were extracted from Lakes Minnewaska (Feb. 1999) and Mohonk (Feb. 2004) in winter using a 7-cm diameter, 1.5-m-long Livingston corer dropped through a hole cut into the frozen lake surface. Cores were split and described the day after extraction followed by sampling. One cubic centimeter samples for carbon content and isotopic analysis were taken every 3-5 cm along each core. Samples were treated with weak HCl to remove any carbonate, rinsed in deionized water, dried at 60 °C, ground with a mortar and pestle, and stored in a desiccator prior to analysis. Organic carbon contents were measured by coulometry. Carbon and nitrogen isotopic values were determined using a Carlo Erba NA 1500 Series II NC elemental analyzer connected to a GV Instruments Optima stable isotope mass spectrometer at the University at Albany in Albany, NY. An additional 1 cm³ sample was taken for pollen and charcoal analysis. Half of this sample was dried and weighed to determine dry bulk density. The remaining half

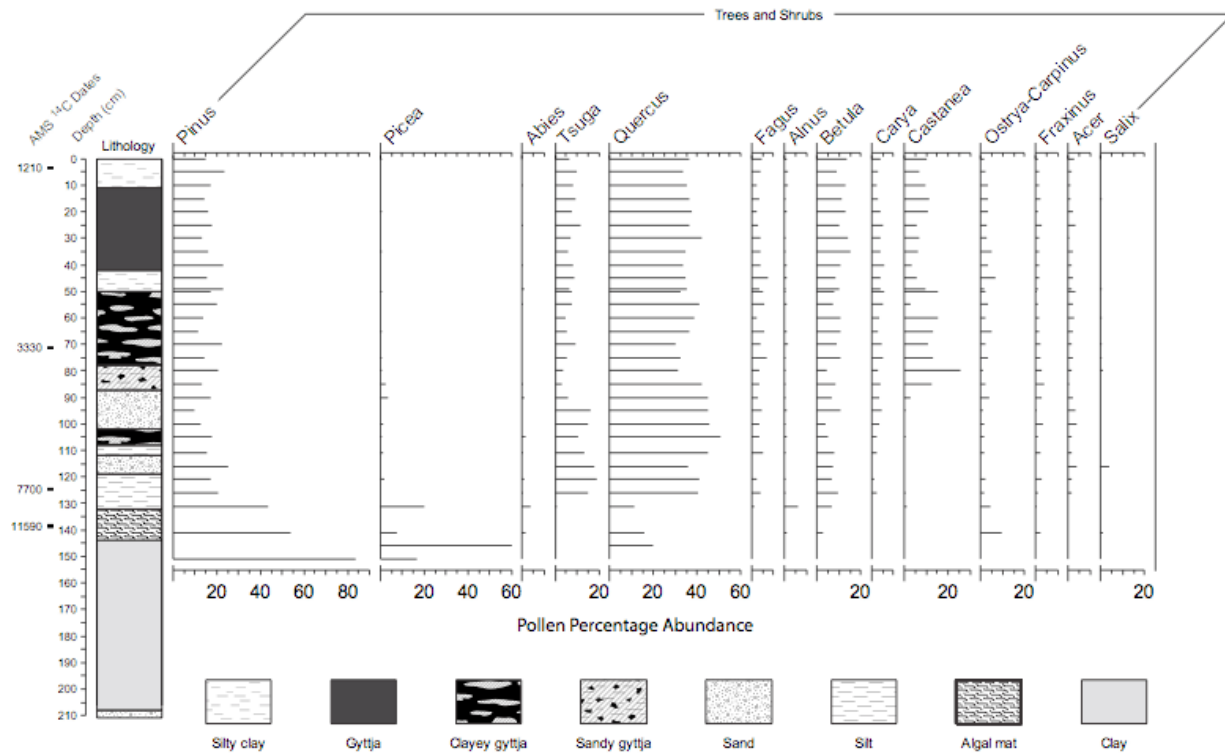


Figure 2. Lithology, pollen and spore percentages from Lake Mohonk. See discussion below (Kirsten Menking, analyst).

was washed through a 125-micron sieve to separate out plant fragments and macroscopic charcoal. Standard palynological and macrofossil techniques were utilized (Faegri and Iverson, 1975; Watts and Winter, 1966).

Lake Mohonk

The Mohonk Lake core (Figure 2) measures 2.1 m in length and was taken from a water depth of ~14 m. The core bottoms in sand, above which lies 0.6 m of gray organic-poor clay. An 0.1-m-thick mat of an unidentified aquatic plant caps the clay at 138 cm depth in the core and dates to 11,590 ¹⁴C yr b.p. Above this plant mat lies a mixture of sand, silty clay, clayey gyttja, sandy gyttja, and gyttja layers varying in thickness from approximately 0.05 to 0.3 m. Organic carbon content in the core varies from nearly 0% in the basal sand and clay units to a maximum of 20% in the algal mat. Above the algal mat organic carbon declines to values between 0 and 16%, with sand units showing low values and gyttja-rich layers showing higher values. Carbon isotopes on organic matter show little variation with depth with the exception of the plant mat, which is nearly 10‰ more negative than the rest of the organic matter in the core. Core sediments are nearly devoid of pollen below 1.55 m depth, so the pollen diagram shown in figure 2 is terminated at that depth.

Lake Minnewaska

The Lake Minnewaska (Figure 1) core measures 2.4 m in length and was taken in a depth of ~20 m of water. The bottom ~0.5 m of the core consists of layered silt, sand and clay, with individual layers a few mm thick. Above this unit lies ~0.5 m of uniform gray clay. Both of these units are devoid of organic matter, and neither contains much pollen. The top 1.4 m of the core is organic rich, with carbon contents measuring between 8 and 32%. The transition from sediments devoid of organic matter to organic-rich sediments occurs abruptly at a depth of 140 cm in the core and has been dated at 14.5 ¹⁴C kyrs b.p. Organic carbon content climbed from <1% to >20% of sediment mass by 10.1 ¹⁴C kyrs b.p. coincident with a 7‰ shift to more negative values of $\delta^{13}C$, indicating that this oligotrophic, rain-water fed lake required 4500 years to acquire a fully developed aquatic ecosystem. Within the organic rich zone occur several large oscillations in carbon content and $\delta^{13}C$, possibly reflecting changes in the relative importance of terrestrial versus aquatic vegetation. As in the Mohonk core, the Minnewaska sediments are devoid of pollen below 140cm.

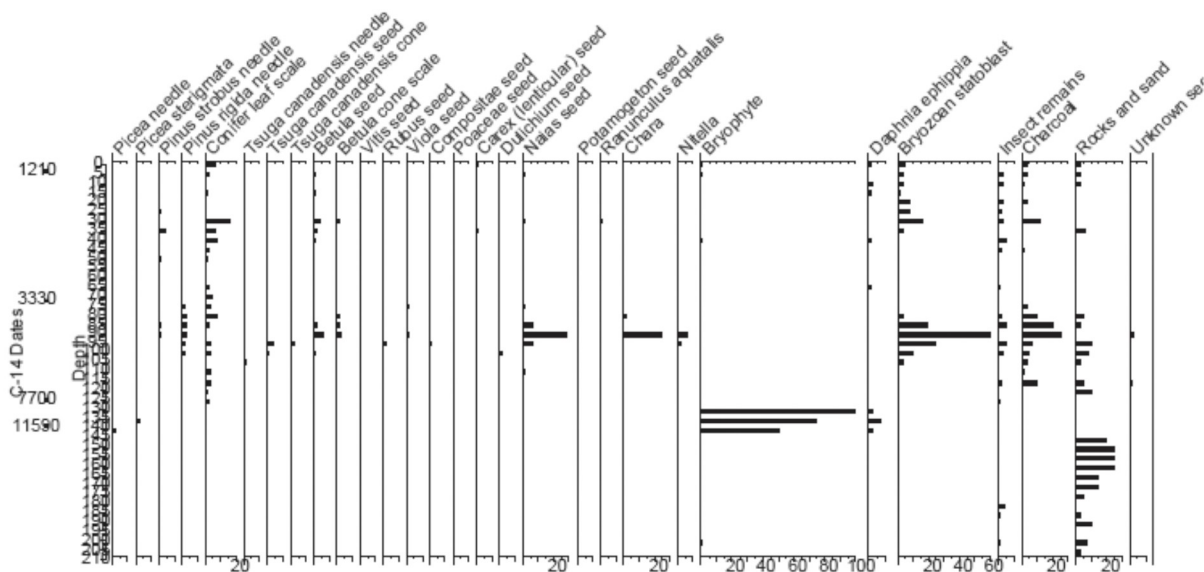


Figure 3. Macrofossil and charcoal diagram from Lake Mohonk (D. Peteet, analyst).

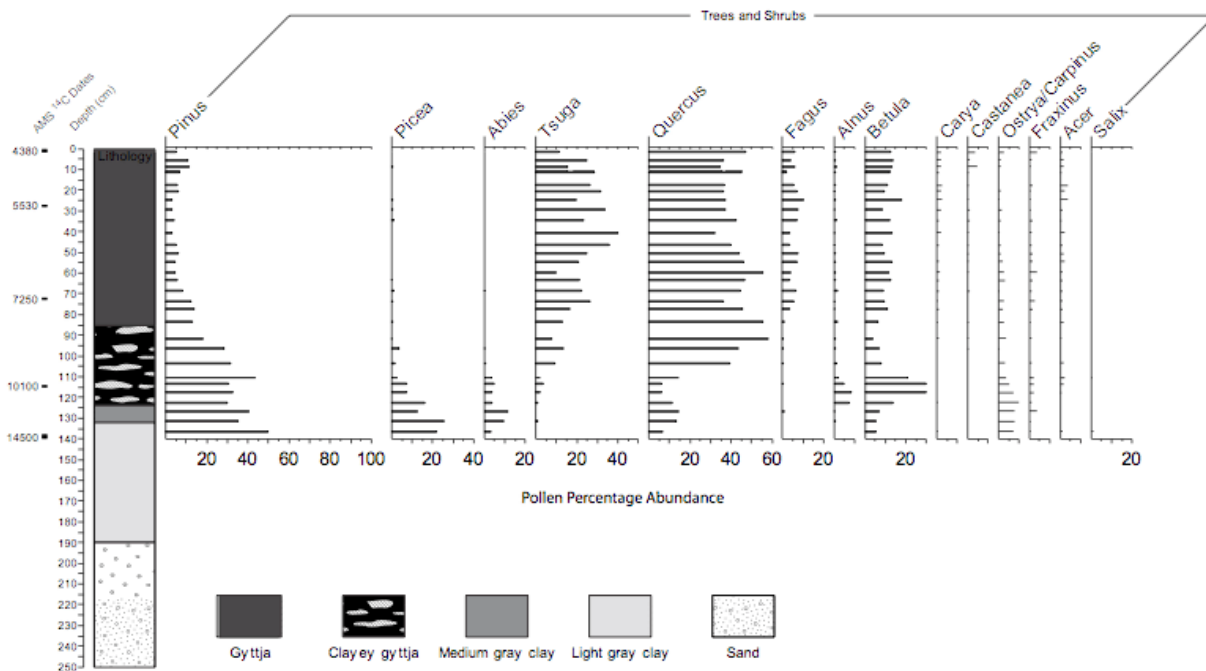


Figure 4. Minnewaska lithology, pollen and spore percentage diagram (Kirsten Menking, analyst).

Mohonk and Minnewaska pollen and plant macrofossil stratigraphies

The layered sands and silts at the base of the Minnewaska core resemble seasonal varves, and this combined with the lack of organic matter and pollen in the basal sediments of both cores suggests that both lakes had inflow streams during deglaciation. Deposition of clay thereafter probably indicates loss of the inflow streams and very local wind and rain erosion that mobilized silt and clay sized particles in the unstable landscape surrounding the lakes. This style of sedimentation ended with the arrival of forest vegetation, when gyttja deposition commenced.

Both the Minnewaska and Mohonk cores show the classical northeastern pollen sequence made famous by workers such as Deevey (1939) and Davis (1969). Spruce (*Picea*) and fir (*Abies*) pollen at the base of each core declined abruptly shortly after 10 ¹⁴C kyr b.p. and were replaced by pine (*Pinus*), oak (*Quercus*) and hemlock (*Tsuga*) as climate warmed following deglaciation. Relatively high amounts of alder (*Alnus*) and birch (*Betula*) between 110 and 125 cm depth in the Minnewaska core likely reflect the Younger Dryas cooling event (Peteet et al., 1990; Mayle et al., 1993). This event is not as readily apparent in the Mohonk core, though one sample at ~130 cm depth shows elevated alder pollen.

The top of the Minnewaska core dates to 4380 ¹⁴C yr b.p., just after the hemlock decline noted in many northeastern pollen records and attributed by Davis (1981) to an arboreal pathogen outbreak around 4800 ¹⁴C yr b.p. The beginning of the hemlock decline is evident in the Minnewaska core and still more convincing in the Mohonk core where hemlock pollen drops by a half to two thirds of its previous abundance at ~95 cm depth. Though we do not have a radiocarbon date on the timing of hemlock decline in the Mohonk core, simple interpolation based on a linear sedimentation rate between the 3330 and 7700 ¹⁴C yr b.p. dates places an age of roughly 5000 ¹⁴C yr b.p. at 95 cm, showing agreement with other records in the region.

Plant macrofossil analysis of the Mohonk core shows the presence of spruce needles in the basal gyttja. Pitch pine (*Pinus rigida*) needles, a pine adapted to fire, appear between 105 and 75 cm along with shallow aquatics such as water nymph (*Najas*), resting spores of green algae stoneworts (*Chara*, *Nitella*) and abundant macroscopic charcoal.

These indicators all suggest drier conditions and fire. .

Foster et al. (2006) have recently questioned the pathogen explanation for hemlock decline in southern New England, calling instead on mid-Holocene drought as the primary driver for changes in forest composition. The charcoal and plant macrofossil data from the Mohonk core support this idea, which is given further credence by the sedimentology, which shows an interruption of clayey gyttja sedimentation in favor of poorly sorted sand deposition in the middle Holocene. We interpret this stratigraphy to represent destabilization of the landscape in the presence of frequent fires along with mass movements.

Site 2 - Rhododendron Swamp

Rhododendron Swamp measure 2.4 hectares and lies 275 m above sea level at the base of a 20 m conglomerate cliff southwest of Lake Mohonk. A 4-meter core was taken in 2002, using both a Livingstone piston corer and a Hiller corer for the top sediments. Rhododendron Swamp records a basal age of 12.5 ¹⁴C (14.6 kyr) at 2.82 m depth, the base of the organic clay. Pollen and spore stratigraphy down to 118 cm (8000 ¹⁴C years) is provided by a masters student at LDEO, Sage Markgraf, and ongoing pollen, spore, and macrofossil stratigraphy to the base of the core is in progress.

The Holocene stratigraphy provides some interesting comparisons and contrasts with the Mohonk, Minnewaska, and Otisville records. Birch (*Betula*) is better represented at this site, probably due to the local nature of the vegetation dominating the pollen record from the swamp. A decline in hemlock and birch along with ground pine (*Lycopodium*) signals the mid-Holocene drought, AMS-dated here at 4975 ¹⁴C. The European impact is evident with the weedy ragweed (*Ambrosia*), plantain (*Plantago*) and grass (Graminae) rise in the top 30 cm. Relatively low hemlock percentages (10-15%) are more similar to Mohonk than to Minnewaska, while chestnut (*Castanea*) percentages were similar to Minnewaska and significantly lower (8%) than seen at Mohonk (up to 20%)

III. Lake Walkkill GIS Model

Methods

This paleo-topographic GIS model was constructed by first compiling a regional Digital Elevation Model (DEM) with a 23 m grid cell size. The model was then adjusted for isostatic rebound by assuming regional E-W isobase and total rebound slope of 0.70 m/km. Although this oversimplifies actual regional rebound, this slope is the same observed in the Champlain Valley and upper Hudson Valley (Rayburn et al., 2005; DeSimone et al., 2008), and similar to that illustrated in Stone, Stanford, and Witte (2002) and Stanford (in press).

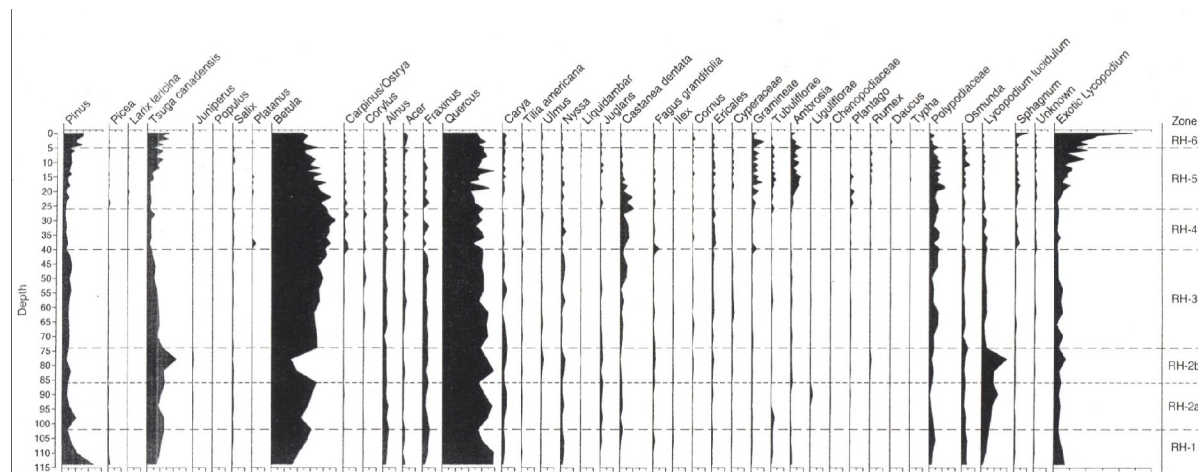


Figure 5. Pollen and spore stratigraphy from Rhododendron Swamp (Sage Markgraf & D. Peteet, analysts); remaining lower section is in progress.

Ice margins were digitized by modifying the Sussex, Pellets Island, Wallkill, and Rosendale ice margins published by Connally and Sirkin (1973) to fill the region, and estimating two intermediate ice margins at about New Paltz and Kingston. The Rondout and Wallkill Valleys were then flooded until they reached the Delaware River drainage divide. This was assumed then to be the highest lake level in each valley. The Hudson River basin was flooded to a level similar to that reported by Stanford (in press) to be the highest level of glacial Lake Albany.

The ice sheet was then made to recede from the oldest digitized margin (Sussex) to the youngest (Kingston) and as lower thresholds were uncovered lake levels in the valleys were adjusted to them. Drop in lake level could then be calculated.

Results

The descriptions below refer to Figure 6.

- A) The model predicts the Wallkill/Delaware drainage divide threshold (yellow circle) at roughly the same location and elevation published by Connally and Sirkin (1973) at a modern elevation of 151 m (495 ft). This is the bed-rock-floored spillway for the Augusta stage of glacial Lake Wallkill (Stone, Stanford, and Witte, 2002; previously the “500 ft level” of Lake Wallkill of Connally and Sirkin, 1973). The ice margin shown in the Wallkill Valley is Connally and Sirkin’s (1973) “Sussex” margin, and at this time Lake Wallkill drained to the North Atlantic via the Delaware River.
- B) The next threshold published by previous studies is the one shown in Figure C. This model, however, identifies another threshold between A and C. The ice margin here is slightly modified from Connally and Sirkin’s (1973) “Pellets Island” margin. A drainage divide is uncovered (yellow circle) that is 8 m above the threshold in A) given modern elevations, however when compensating for isostatic rebound, this threshold becomes 7 m below the threshold in A). If this is correct, then when the ice reached this position Lake Wallkill drained southward to the Atlantic Ocean via the Ramapo and Passaic Rivers, and the lake level dropped slightly (~7 meters). This is only a very minor change in lake level, and if this did happen it would probably be difficult to distinguish shoreline features between this level and the previous one. The Hudson Highlands separate the Wallkill from the Hudson drainage basin.
- C) At Connally and Sirkin’s (1973) “Wallkill” ice margin a significantly lower threshold across the Hudson Highlands is exposed (yellow circle). This threshold is at a modern elevation of 108 m (354 ft). The model shows this level to be 70 m below the threshold in A), which completely agrees with Stanford’s (in press) estimate. This threshold would allow Lake Wallkill to discharge into Lake Albany via the Otter Kill and Moodna Creeks. Stanford (in press) calculates that this 70 m drop (not having recognized the potential lake level change in B) would send about 25 km³ of meltwater into Lake Albany. He suggests that this flood overwhelmed the Lake Albany threshold at Hell Gate and caused the lake to overtop and breach the Narrows dam. This in turn would have lowered Lake Albany to a series of unstable levels (Lake Albany level changes are not depicted in this model). In the Wallkill Valley the most interesting observation from the model is that, given this ice margin position, Lake Wallkill appears to be nearly completely drained. Also, at this ice margin position the Rondout/Delaware drainage divide becomes exposed in the Rondout Valley, and a proglacial lake begins to form on the west side of the Shawangunk Mountains.
- D) This figure shows a hypothetical ice margin at about New Paltz. A lower level of Lake Wallkill has become well developed east of the Shawangunk Mountains, and the lake in the Rondout Valley continues to extend northwards. This level of Lake Wallkill, referred to as the “400 ft level” by Connally and Sirkin (1973), is still controlled at the 108 m (354 ft) Otter Kill threshold.
- E) The ice margin in this figure is shown at Connally and Sirkin’s (1973) “Rosendale” margin, which they considered a re-advance position. They recognized a 320 ft. (98 m) level and a 220 ft. (67 m) level in the Wallkill Valley, but thought that perhaps they may relate to early stages of Lake Albany. At the Rosendale ice position, however, a threshold is exposed across the north end of the Marlboro Mountains (yellow circle) just east of New Paltz at the Swarte Kill/Black Creek drainage divide. This would direct discharge through the gap that Rt. 299 currently follows between New Paltz and Highland. Drainage would then have followed the Black Creek route

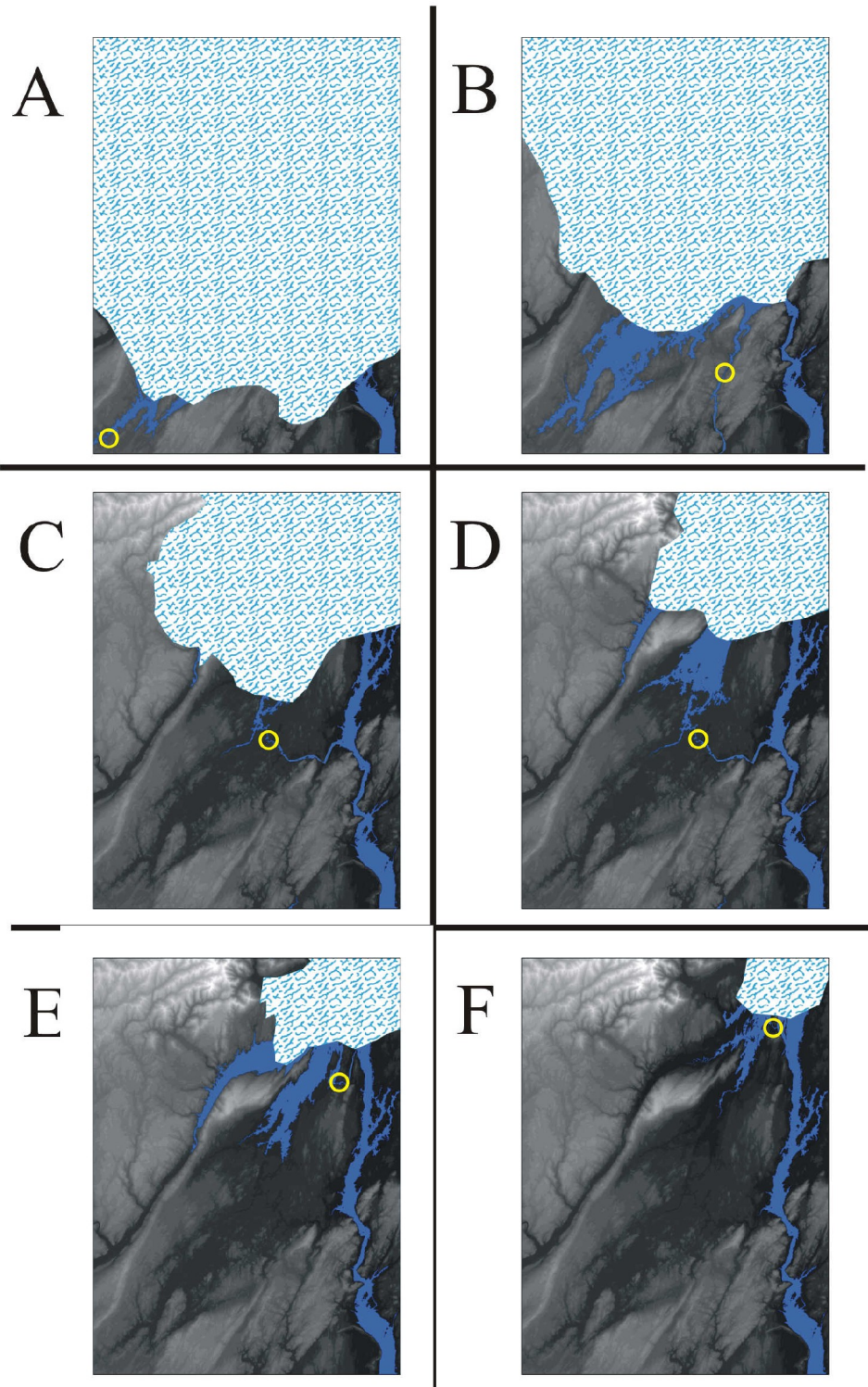


Figure 6. (see text for description)

north towards Esopus. Although currently 109 m (1 m higher than the Otter Kill threshold in D), this threshold would have been 21 m lower than the previous threshold given the estimated rebound. This would be a modest drop in the level of Lower Lake Wallkill. This ice position is pinned against the north end of the Shawangunk Mountains, and this figure shows the glacial lake in the Rondout Valley at its maximum extent. When the ice retreats from the Shawangunks Lake Roundout will drop about 65 m to become confluent with Lower Lake Wallkill given the threshold depicted here. That should have caused a significant discharge through the Wallkill basin and into Lake Albany.

- F) This figure shows a hypothetical ice margin position at about Kingston. At this point Lower Lake Wallkill extends into the Rondout Valley, and another threshold has become exposed (yellow circle). This threshold is at a modern elevation of 73 m (240 ft) and leads to the Hudson Valley through a gap called “The Hell” west of Ulster Park. This threshold, although not previously recognized is important for two reasons. First, the model shows that it was 46 m below the previous threshold, which would probably have caused a significant flood into Lake Albany. Secondly this elevation corresponds exactly with the deltaic deposit at Tillson and Rosendale. This is the best expressed lacustrine strandline in the lower Wallkill Valley, until discovery of this threshold, was difficult to account for. When the ice retreated to Kingston and exposed the entire Rondout drainage route, Lake Wallkill would have completely drained away into Lake Albany. According to Stanford’s (in press) estimates for Lake Albany levels, the final stage of Lower Lake Wallkill was only about 15 m above Lake Albany.

IV. Wallkill Valley Regional Sites

Robinson et al. (2005) investigated pollen stratigraphies, including from mastodon and stag moose sites in the Wallkill Valley (Figure 7), as part of an ongoing project to reconstruct the climate, vegetation, fire history and large animal densities from the Pleistocene to present. Megafaunal populations collapsed throughout the region at the end of the Pleistocene, as ice sheets retreated and the earliest humans arrived. The study has included microscopic charcoal analysis to follow the fire history at the beginning of the Paleoindian period in the northeast. Distinctive fungal spores of the dung fungus *Sporormiella* were used as a proxy for megafaunal biomass. The results from several sites show reflect a rapid decline in spore values, closely followed by a stratigraphic charcoal rise, reflecting changes that unfolded at least 1000 years before the end Pleistocene climatic reversal of the Younger Dryas (YD). The YD is identified as pollen zone III from Binnewater Pond shown in Figs. 8, 9. Megafaunal fossil sites from the Black Dirt display a broadly similar microfossil stratigraphy, and pollen zones are readily correlated.

Although most direct bone dates of extinct megafauna suggest that these animals lasted until at least the beginning of the Younger Dryas. Robinson et al., 2005 suggest that a regional collapse of large herbivore populations was followed by landscape transformation by humans. Elevated stratigraphic charcoal could result from reduced herbivory or human caused fires, or both. The findings are consistent with the proposal that human activities rather than climate were the key drivers of the extinction event. And although these data in themselves give no indication as to actly how this may have happened, the fluted points of Dutchess Quarry Cave and other nearby sites are suggestive.

The rise in spruce in zone II (Figure 8) shows the shift toward the Bolling/Allerod warming; the overlying zone III is interpreted as the Younger Dryas cooling. When compared with the *Sporormiella* and the charcoal analysis in Figure 9 below, a drop in *Sporormiella* in zone II is seen, concurrent with the spruce increase and warming, while charcoal appears from the beginning of deposition and peaks in zone II.

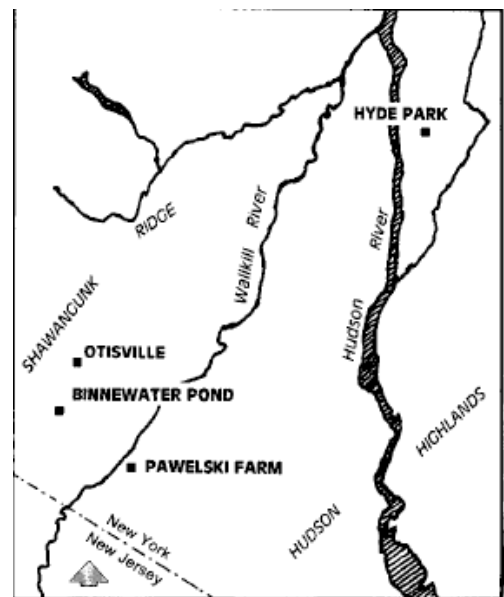


Figure 7. Sites investigated by Robinson et al (2005) in the Wallkill Valley.

Sporormiella spores are at least 3% throughout Zone I but decline below 1% in lower Zone II, whereas charcoal concentrations increase by 10 fold.

V. Hudson Marsh Paleocology and Croton Point Varves and Till

Hudson marshes provide a unique perspective of the Hudson Valley climate because they are very high sedimentation archives, and record not only the pollen of the upland and marshes, but archive the inorganic component of a watershed as well as the charcoal in the watershed. Through the last decade, Peteet et al. (2006) have focused on the collection and analysis of Hudson River marsh cores for understanding the paleoenvironment of the estuary (Table 1, Peteet et al. 2006).

The pollen and charcoal record from Piermont Marsh (Figure 10), for example, documents the most detailed, well-dated Medieval Warm Interval between 850 and 1350 AD with very high charcoal and pine and hickory (*Carya*) expansion (Pederson et al., 2005), while the upper sediments record invasive species due to human impact.

Ongoing research on the cores from this site reveal sequences of apparent droughts and wet intervals which probably are correlative with the upland sequences from Mohonk, Black Rock Forest, and the Black Dirt region.

The upper meter of the Croton Marsh core sampled at 2cm intervals provides a glimpse of what the typical Hudson marsh looks like today (Figure 11), dominated by reed grass (*Phragmites*) which invaded the cattail (*Typha*) marsh, but which originally was comprised of sedges (*Scirpus*, *Cladium*) prior to European impact.

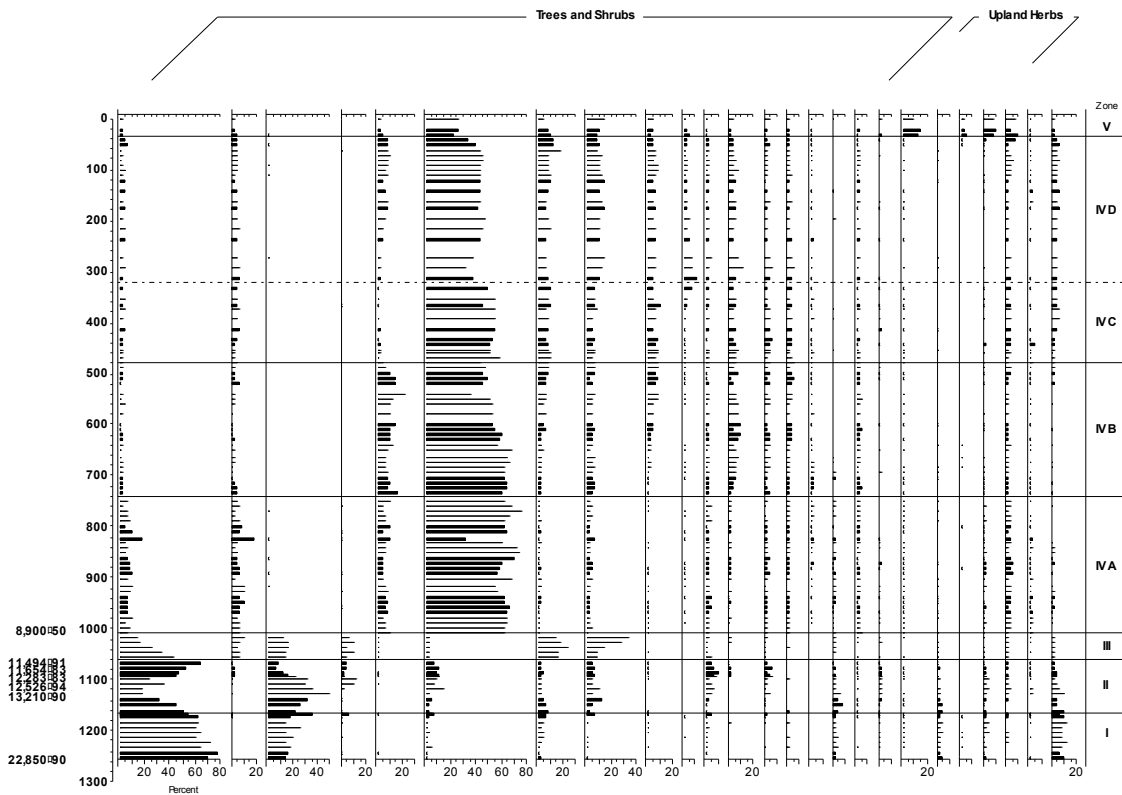


Figure 8. Binnewater Pond Pollen Stratigraphy(adapted from Robinson et al., 2005) .

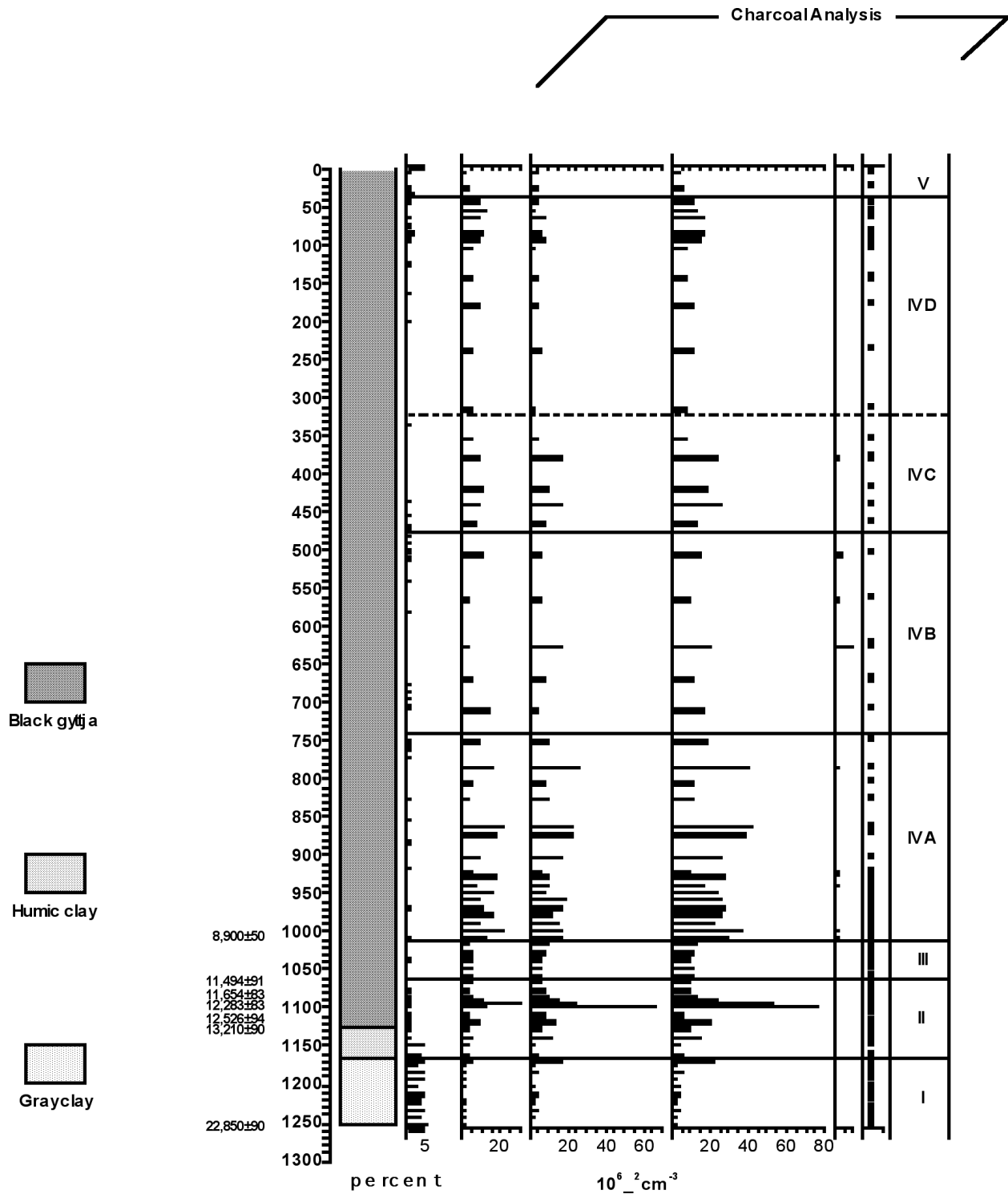


Figure 9. Binnewater Pond charcoal and Sporormiella (adapted from Robinson et al., 2005).

Table 1. Location, peat depth, and basal ¹⁴C age of Hudson marshes.

Tidal Marsh	Latitude/ Longitude	Peat Depth	Basal ¹⁴ C Age
JoCo	40° 37'N	2 m	>460 <2000
Jamaica Bay	73° 47'W		
Yellow Bar	40° 37'N	0.8 m	>450
Jamaica Bay	73° 50'W		
Arthur Kill	40° 36'N	8.0 m	11,100
Staten Island	74° 13'W		
Hackensack	40° 48'N	3.7 m	2,610
New Jersey	74° 04'W		
Piermont N	41° 00'N	13.7 m	5,700
Piermont	73° 55'W		
Croton Marsh	41° 14'N	10 m	4,630
	73° 50'W		
Iona Marsh	41° 18'N	10 m	5,500
	73° 58'W		

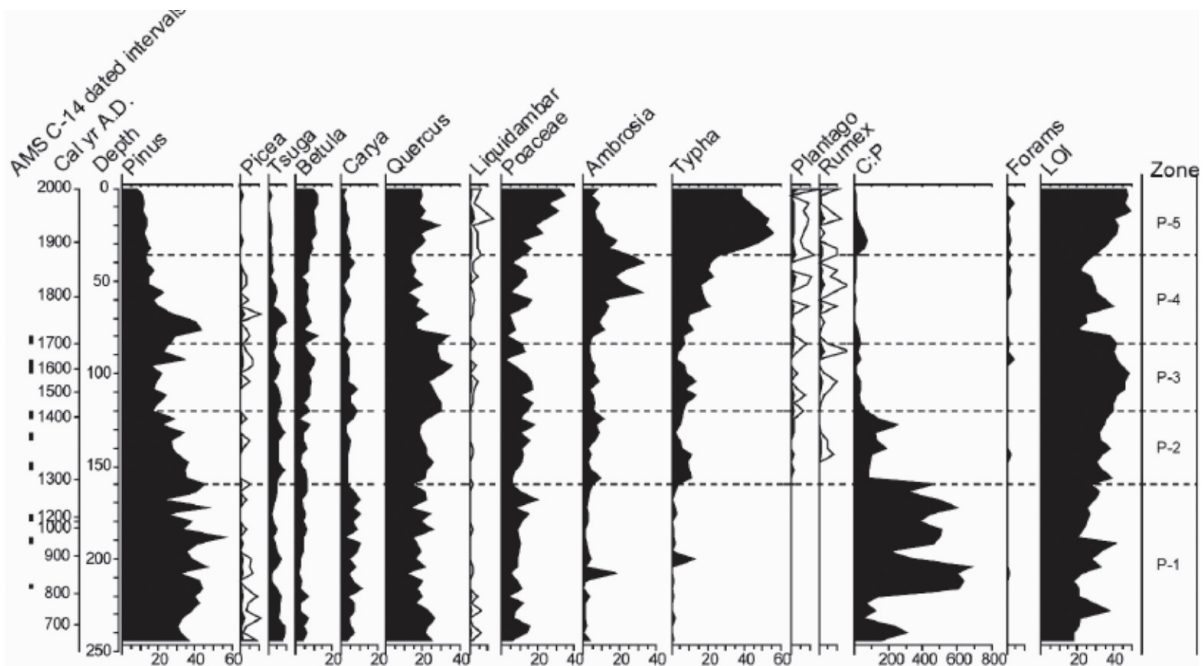


Figure 10. Pollen and charcoal/pollen stratigraphy from Piermont Marsh (Pederson et al., 2005).

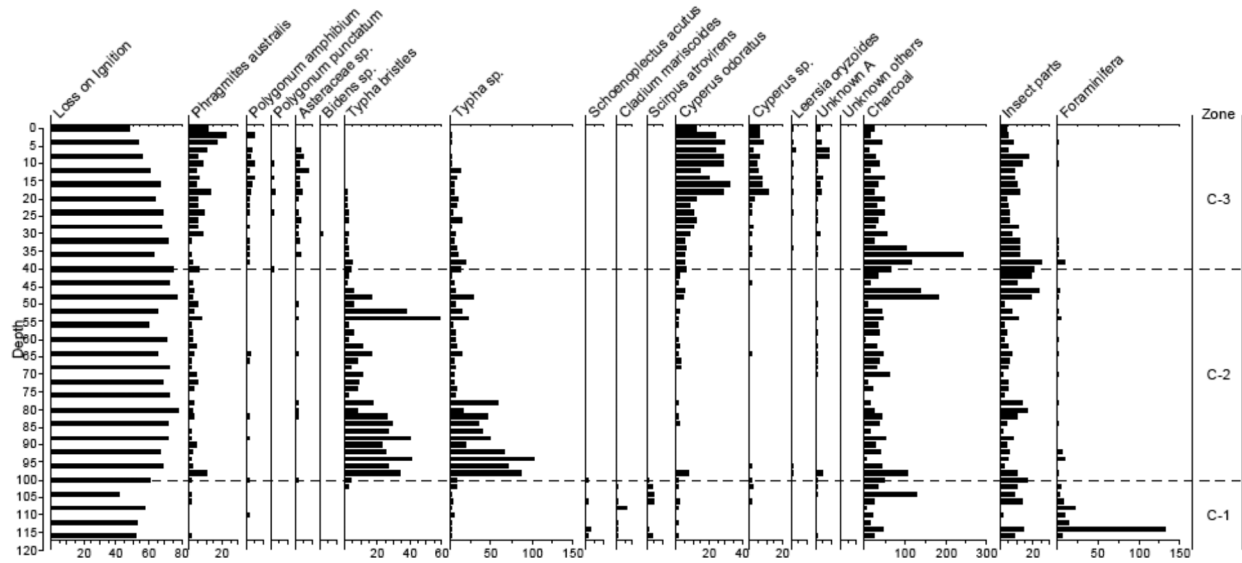


Figure 11. Croton Marsh loss-on-ignition and macrofossil stratigraphy in top 1.2 meter of the core (Caity Schumehl and Dorothy Peteet, analysts).

At Croton, as in much of the Hudson estuary, marshes have been destroyed by landfills atop them. But the Croton archive has much to tell us about the regional droughts, and nearby Croton Point provides a unique river site to examine varve stratigraphy atop tills (see details in Stop).

Road Log

Miles Between Points	Cumulative Mileage	Description
0.0	0.0	Assemble at the parking lot adjacent to the Wooster Science Building on the SUNY New Paltz campus
0.02	0.02	Drive southeastward out of the parking lot and immediately turn left onto an unnamed campus rd.
0.10	0.12	Turn left (northwest) onto Plattekill Rd.
0.20	0.32	Bear left (west) onto Hasbrouck Ave.
0.20	0.52	Turn right (north) onto S. Chestnut St.
0.11	0.63	Turn left (west) onto Rt. 299/Main Street and cross the Walkill River.
0.32	0.95	Turn right (north) onto Springtown Rd.
0.47	1.42	Bear left (west) onto Mountain Rest Rd.
3.40	4.82	Turn left (southwest) at the entrance of Mohonk Mountain House
0.9	5.72	Pass through the check point and continue along the 1-way road.
0.87	6.59	Continue along the 2-way road.
0.06	6.65	Bear right at the Y in the road
0.07	6.72	Bear right again
0.13	6.85	Turn right onto Cedar Dr. and descend the hill
0.03	6.88	Turn right and park in front of Elms Cottage

Stop 1. Mohonk Lake. We begin our trip with a walk around the shores of Mohonk Lake, guided by Paul Huth, director of research for Mohonk Preserve. Starting in front of Mohonk Mountain House, a Victorian era hotel built in 1869 and operated continuously by several generations of the Smiley family, we will climb onto the quartz pebble conglomerate cliffs of the Silurian Shawangunk formation that surrounds the lake. Glacial striations and crescentic gouges are evident at several points along the trail, and Mohonk Lake itself is thought to have been formed by gla-

cial plucking of tectonically fractured bedrock. The sediment core described earlier in this guide was taken at the northern end of the lake near the Mountain House and spans most of the Holocene (see Figures 1-3). The first appearance of organic matter in the core dates to approximately 11.6 ¹⁴C kyr b.p., in keeping with young AMS dates found at other sites in southern New York and New Jersey. In addition, the core appears to record a mid-Holocene drought episode that might have caused the decline of hemlocks previously attributed to a pathogen.

Continuing along the trail affords a view of Rhododendron Swamp from above. A paleoindian rock shelter adjacent to the swamp contains evidence of human habitation as early as 10 kyrs b.p. and possibly as early as 11.5 kyrs b.p. (Eisenberg, 1991). Ongoing paleoecological stratigraphy attempts to link the Swamp record to the archeological record..

End Stop 1. Retrace steps to depart Mohonk Mountain House Resort.

From Mohonk Lake to Dutchess Quarry Caves, Goshen, NY (as given by Google Maps).

1. Head south on Garden Rd 13 ft
2. Turn left to stay on Garden Rd 1.1 mi
3. Turn left at Terrace Rd 0.9 mi
4. Continue on Garden Rd 236 ft
5. Turn right at County Rte-6/Mountain Rest Rd 3.4 mi
6. Slight right at County Rte-7/Springtown Rd 0.5 mi
7. Turn left at County Rte-7/New Paltz Plaza/NY-299 W/State Route 299 W
Continue to follow NY-299 W/State Route 299 W 1.6 mi
8. Take the ramp onto I-87 S Toll road 16.6 mi
9. Take exit 17 for NY-17K/I-84 toward Newburgh 0.5 mi
10. Merge onto Auto Park Pl 0.3 mi
11. Take the ramp onto NY-300/Rte-300/State Route 300 0.9 mi
12. Make a U-turn at Union Ave 69 ft
13. Take the ramp onto I-84 W 17.4 mi
14. Take exit 4E to merge onto NY-17 E toward New York 4.5 mi
15. Continue on US-6 E 0.3 mi
16. Take exit 124 for NY-207/NY-17A toward Florida/Goshen 0.3 mi
17. Turn left at NY-17A/Rte-17A/State Route 17A 2.1 mi
18. Turn right at Pulaski Hwy 0.5 mi
19. Turn left at Quarry Rd 0.3 mi

Destination will be on the left. We park along Quarry Road and walk up to the caves on Mt Lookout.

Stop 2. Dutchess Quarry Caves: Paleoindian culture and Pleistocene Megafauna. Perhaps the most significant early human site in the northeast is at Dutchess Quarry, in a group of small caves formed in Paleozoic limestone on the northwest side of Mount Lookout in southern Orange County, NY. One complete and four partial Paleoindian fluted projectile points have been recovered from cave numbers 1 and 8 (Funk et al., 1969; Funk et al., 1970; Kopper et al., 1980; Funk and Steadman 1994) Bones of caribou, (*Rangifer tarandus*), extinct flat-headed peccary (*Platygonus compressus*) and the extinct giant beaver (*Castoriodes ohioensis*) have been among the 71 species of vertebrates discovered, although there is no clear association between any of the Pleistocene fauna and the cultural material (Steadman, Stafford and Funk 1997).

At an elevation of 177m Mount Lookout offers a view from 80m above the Black Dirt agricultural region, itself lying over the largest accumulation of terrestrial peat in eastern United States after the Florida Everglades. A succession of proglacial lakes occupying the Wallkill River Valley left a large, poorly drained area that was to become an immense peat deposit continuing to build throughout the Holocene. By the early 20th Century, these mucklands were being artificially drained for agriculture. Occasionally, maintenance of drainage ditches exposed the remains of the extinct Pleistocene megafauna. In this way, numerous mastodons (*Mammuth americanum*) and at least three stag moose skeletons (*Cervalces scotti*) have been discovered in and around this vast wetland. The two most recent of the

stag moose have been AMS dated to 12,180±60 and 11,040±110 ¹⁴C yrs before present. The latter individual is the latest known occurrence of this species in North America.

Depart Mt. Lookout.

Start at: Quarry Rd Goshen, NY 10924

1. Head northeast on Quarry Rd toward Florida Rd/Florida Green Dr/NY-17A/Rte-17A/State Route 17A - 0.1 mi
2. Turn right at Florida Rd/Florida Green Dr/NY-17A/Rte-17A/State Route 17A - 0.6 mi
3. Turn left at Durland Rd - 0.7 mi
4. Turn left at NY-94/Rte-94/State Route 94 Continue to follow NY-94/Rte-94 - 2.7 mi
5. Turn left at West Ave - 0.5 mi
6. Turn right at Brookside Ave - 0.2 mi
7. Turn left at Academy Ave - 0.2 mi
8. Turn left at Main St - 0.3 mi
9. Continue on High St/NY-94 Continue to follow NY-94 - 14.4 mi
10. Turn right at Forge Hill Rd - 1.4 mi
11. Continue on Sloop Hill Rd - 486 ft

Arrive at: Sloop Hill Rd New Windsor, NY 12553

Stop 3. Newburgh Terrace.

The only available exposure of glaciolacustrine deposits on the west side of the Hudson River is in this pit at the mouth of Moodna Creek on the south side of Newburgh. The pit is in the south side of the excavated terrace that has a surface altitude of over 49 m (160 ft). A small inset plain at the top of the exposure had a surface altitude of just above 33 m (100 ft). The pit exposes: 1) cobble gravel at the surface, 2) glaciolacustrine beds that extend to the bottom of the cut. The gravel, 1-2 m thick, underlies the inset terrace surfacecut into the southern part of the original higher glacial terrace landform. The gravel rests on a sharp, horizontal disconformity with the underlying sand. The surface of the gravel plain is correlated with nonspecific levels of glacial Lake Albany, projected from deltas surfaces and topset-fore-set contacts from deposits east and north of this site. The gravel in this exposure is a thin, postglacial fluvial terrace deposit, probably graded to lowering lake levels in the valley.

The glaciolacustrine sand deposit consists of alternating, laterally continuous beds of fine to coarse sand, pebbly sand, and very fine sand and silt. Basal bedding contacts are sharp; bedding forms are chiefly flat thin beds and laminations, with few ripples and hummucky forms. Color of beds is related to their composition and, therefore, their particle size: coarser beds are lighter colored and contain quartz, some feldspar, minor carbonates, and scattered pebbles of carbonate and sandstone. Darker beds are finer grained and contain platy rock frag-



Figure 12. Fluvial cobble gravel disconformably overlying glaciodeltaic sand deposits in the 49 m (160 ft) terrace at the mouth of Moodna Creek south of Newburgh, New York.

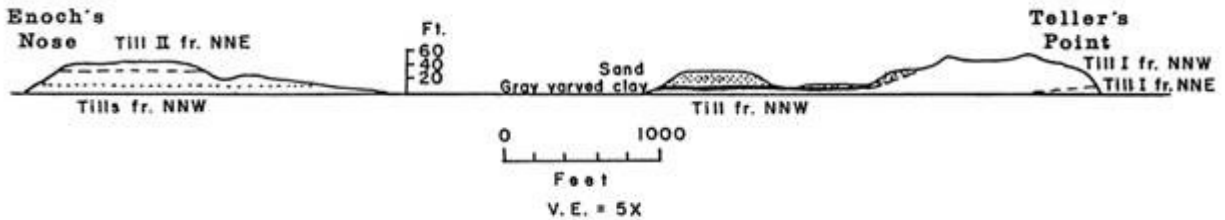


Figure 13.

ments of shale and some carbonate. The lake beds have a gentle southerly slope of $<1^\circ$ reflecting their subaqueous origin as bottomset beds of an ice-marginal delta deposited in glacial Lake Albany.

The excavation exposed on this trip affords us an opportunity to determine the local vertical successions of delta bottomset underflow deposits, their cross-cutting relationships, and paleocurrent flow directions.

1. Drive north on Sloop Hill Rd. toward Shore Rd. - 0.2 mi
2. Turn left onto US-9W South. - 13.4 mi
3. Enter next roundabout and take 3rd exit onto US-202 North/US-6 East. - 0.2 mi
4. US-202 North/US-6 East becomes US-6 East/US-202 East (Portions toll). - 4.1 mi
5. Enter next roundabout and take 1st exit onto US-202 East/US-6 East/US-9 South/Lower South St. - 0.2 mi
6. Turn right onto US-202/US-6/US-9. Continue to follow US-9 South. - 8.2 mi
7. Take the Croton Point Ave ramp. - 0.2 mi
8. Turn right onto Croton Point Ave. - 0.4 mi
9. Arrive Croton Point Park

Stop 4. Croton Marsh. Archeological evidence indicates that the Croton region was populated by the Kitchawanc tribe as early as 5000 BC. The marsh was called Senasqua, Croton itself is believed to be named for the local Indian Chief, meaning “wild wind”. A 1718 census of the area counted 91 inhabitants in the Manor estate of Cortland, including Dutch settlers and English Quakers (<http://www.crotononhudson-ny.gov>). An overview of the Croton Marsh remaining from the landfill road to the south of the entrance to Croton Park. Today, one can only see the invasive *Phragmites communis* covering the site, but prior to European impact it was comprised of a diverse sedge mixture. Peteet et al. extracted a 10m marsh core and pollen/macrofossil work on the core is in progress.

Park at parking lot closest to Nature Center, and walk south to river exposure.

Stop 5. Croton Moraine and Varves. A summary of Croton Point Pleistocene glacial history is provided by Sanders and Mergurian (1994). It is as follows:

- 1) Several tills starting with gray-tan, then red-brown, then, after red outwash was deposited, another red-brown till, followed by the yellow-brown till. The red-brown tills contain erratics from the west side of the



Figure 13. Varves atop till at Croton Point. Photo by D. Peteet.

Hudson River whereas the youngest and oldest tills contain only rocks found on the east side of the Hudson River.

- 2) After the youngest of the tills had been deposited and the glacier responsible for it had melted away, the region was flooded. All the drainage from the Great Lakes flowed eastward through the Mohawk Valley and down the Hudson. Proglacial Lake Albany was backed up behind the natural dam of till at the Narrows. Deltaic sediments from the ancestral Croton River and possibly drainages to the north were deposited along the east shore of this lake. The water plane presumably stood at about elevation +70 feet (level of the flat terrace underlain by topset beds of the delta that coincide with the uppermost water level). To the west, the depth of water where the clay was deposited away from influence of the delta was 70 feet. The coarse browner clays probably represent the dark suspended load of the river(s). The light clays are winter deposits when river(s) experienced low-flow conditions and/or were shut down altogether because their waters froze solid.



Figure 15. Varves at Croton Point. Photo by D. Peteet.

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The Role of Imagery in Popular Geoscience Writing: The Ice Age at North Lake

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Introduction

Scientists, and science itself, have not always communicated well with the general public. This is especially true at the local or grassroots level. Scientific prose is laden with heavy terminology and scientists are not comfortable in shedding this writing mindset when addressing the public. It makes reading our prose burdensome and that, in turn, makes it difficult to establish two-way communications. We should, all of us, eschew the obfuscation that comes from arcane and sesquipedalian jargon, but good popular science writing is a great deal more than just finding short words. It needs to be understood that popular science writing is a form of literature, or at least it needs to try to be. That involves a very different approach.

The purpose of this field trip is to illustrate and explore one literary device used in popular geoscience literature: imagery. Our science has an almost unlimited potential to generate fascinating, even captivating images of the Earth's deep past. Such images, when presented in a literate prose, can capture the attention and imagination of the general reading public. Exploiting this properly, we geologists can be ambassadors of science to a very wide audience.

It must be obvious how much this is needed in a culture where too many have succumbed to the lures of pseudoscience. We need to be proactive in presenting our various sciences in order to counterbalance the efforts of those such as psychics, astrologers, faith healers, and, most of all, creationists. These "schools of thought," are, or should be, the artifacts of an earlier stage in our culture's development. We have been lax in confronting their challenges at the local level. Too often we have ceded the battlefield to them, assuming that we would win the fight for hearts and minds on the basis of our own history of success and logic. We have been in our labs while they have been on our school boards!

Geology is interesting. Every branch of this varied field offers fascinating topics which compel the broadest appeal to the general public. This field trip will explore the ice age history of one of New York State's most scenic realms: North-South Lake State Park (referred to hereafter as "North Lake"). It is an ideal location to introduce the general public to the Ice Age itself. North Lake has a venerable heritage, occupying a site that has attracted visitors since the early 19th Century. For nearly 200 years, generations of people have been drawn to its scenery: its view of the Hudson Valley, its picturesque mountainous landscape, and the two beautiful lakes. North Lake has been a very influential locale; it is the birthplace of the Hudson River school of art, America's first and very serious contribution to the world of landscape painting.

Scenic as it might be today, North Lake has a very complex ice age history, recorded in the landscape and on the bedrock. These evidences of the Ice Age bring to the mind's eye vivid images of a chapter of glacial history that would have fascinated any of the Hudson Valley school landscape artists who worked here. Nobody painted those scenes from the past, but they can be "painted" in words.

Our journey begins at the top of the Catskill Front, the "Wall of Manitou," on the ledge which is the historic site of the famed, but long gone, Catskill Mountain House Hotel. This ledge, in some form, must have been there at the time when the glaciers first came down the Hudson Valley.

Recently I stood at the lip of the ledge and gazed into its past. I saw the Hudson Valley, before me, much as it had been 23,000 years ago. It was heavily forested then, and I found that the forest was not much different than you would expect nowadays; there were oaks, maples and birch. There were a lot of chestnuts back then too. Many of these trees were enormous; there had never been any axes in this forest. Trees had grown to great age and size.

As a geologist I was blessed with being able to watch this forest for centuries, a lot of them. I thought that the summers were getting shorter and grayer. They seemed to stop getting really hot; in fact they were downright cool and cloudy. There seemed to be fewer warm weather birds, and I thought there were fewer summer insects as well; I rarely heard katydids.

The winters were not all that cold though, but they too were overcast. It always seemed to be about 31 degrees out and snowing. I got to be a little weary of the snowfall, but it would not stop. As more centuries elapsed I noticed that the trees atop of nearby South Mountain seemed to sicken. Even in late August they seemed pale, even yellow. They were losing a fight (or perhaps a war) with the climate. With more centuries I saw that patches of them were dead.

Then, down below, I noticed the same affliction in the forest of the Catskill Front. Their leaves, even in August, were small and yellow. Summer, it seemed, was just not warm enough to allow healthy growth of these trees. What was going on? I was the only person in all of North America. There was nobody to go and ask.

Now I noticed that the snowfall was lasting into May and then June. The new snows were retuning in October and even late September. We are used to climatic cycles, but our droughts are always followed by rainy seasons; our heat waves are balanced by cold spells. But I was watching a one-way process. The Catskills were changing into a land without summer. Now the forests up on North Point were all dead and the trees below were dying as well. It was a somber sight.

There was something else. I saw weeks of weather when the skies were blue, clear, and cloudless, while cold, very dry winds blew steadily out of the northeast. I looked that way and wondered if there was something cold and dry in that direction. I could not even guess an answer. It was a relentless wind, and it got on my nerves after a while, but it would not stop.

Now the snows lay on the ground longer into the "summer" season. I saw dirty old snowdrifts in mid-July. One summer the ground never thawed out and that was the year all the trees finally died.

Trees don't make noise, but once they are dead all the noise makers soon disappear. I watched as all the singing birds, all the noise-making insects, and all the vocal animals vanished. The dead forest became unnervingly silent.

But real silence was rare; those howling dry winds continued and now they desiccated the dead trees. Soon dry and brittle twigs were blown down, then the branches fell, and finally even the great limbs crashed to the ground. This made the forest a very noisy place . . . until only the tree trunks were left.

I, the lone witness to all this, again gazed to the northeast. There I spied something I had not seen before. A large massive entity now filled that distant stretch of the Hudson Valley. In the early morning it was a dark blue which graded to a green and then a yellow as the mornings progressed. The object was a brilliant white at noon. In the afternoon the pattern of colors was reversed. But what was I looking at? As weeks and months progressed, it crept farther south, down the valley. With enough time its image sharpened and its identity was revealed; it was a great glacier, filling more and more of the valley and advancing to the south. Now, at last, I comprehended the very nature of all the mysteries that I had been watching. The Hudson Valley and the Catskills had been sinking into an Ice Age.

Events accelerated. Soon, the great stream of ice passed by below me. Ominously, it began to ascend the slopes of the Catskill Front. It rose closer to where I stood. Then I heard a distant snapping, falling and crashing sound, and then another. These sounds became routine and I wondered what it was. This mystery, too, would be solved. I was looking in just the right direction when I saw and heard the advancing glacier pushing over one of those old and dead tree trunks, and then another.

The full impact of all this was frightening; it was overwhelming. Now streams of ice were flowing across the top of the Catskill Front. More ice was flowing up Kaaterskill Clove. The ice was reaching towards the top of South Mountain. Then it even overtopped that mountain's crest (Figure 1).

NYSGA 2009 Trip 5 - Titus

I had been transfixed by the flows of the glaciers, now all around me, and was not paying much heed elsewhere. But then I did turn to the north and gazed up at North Point. There, in a spine chilling flash, I saw the image of a great ice sheet coming across that mountain. It signaled the main act of this drama. This was not some small valley glacier; this was the real thing. An enormous and thick ice sheet had been advancing out of the north and now it was about to override all of North Lake and all of the Catskills and much of North America! (Adapted from Titus, 2005 and 2007)

The preceding passage is hardly something to be called “scientific.” It is an *image* of the Ice Age, bordering on fantasy. Most of it lacks evidence so it is not science; it is art in the service of science. And that is the point of this field trip and this guide.

Most conventional scientific prose might well be compared to realistic art. The motives, both in science and such art, are generally the same. The scientist and the artist seek to portray the world as it is. But the artist is always given “license.” He is free to play with colors and shades of darkness to bring an “artistic” interpretation to his portrayal. He is always free to alter nature in order to capture Nature.

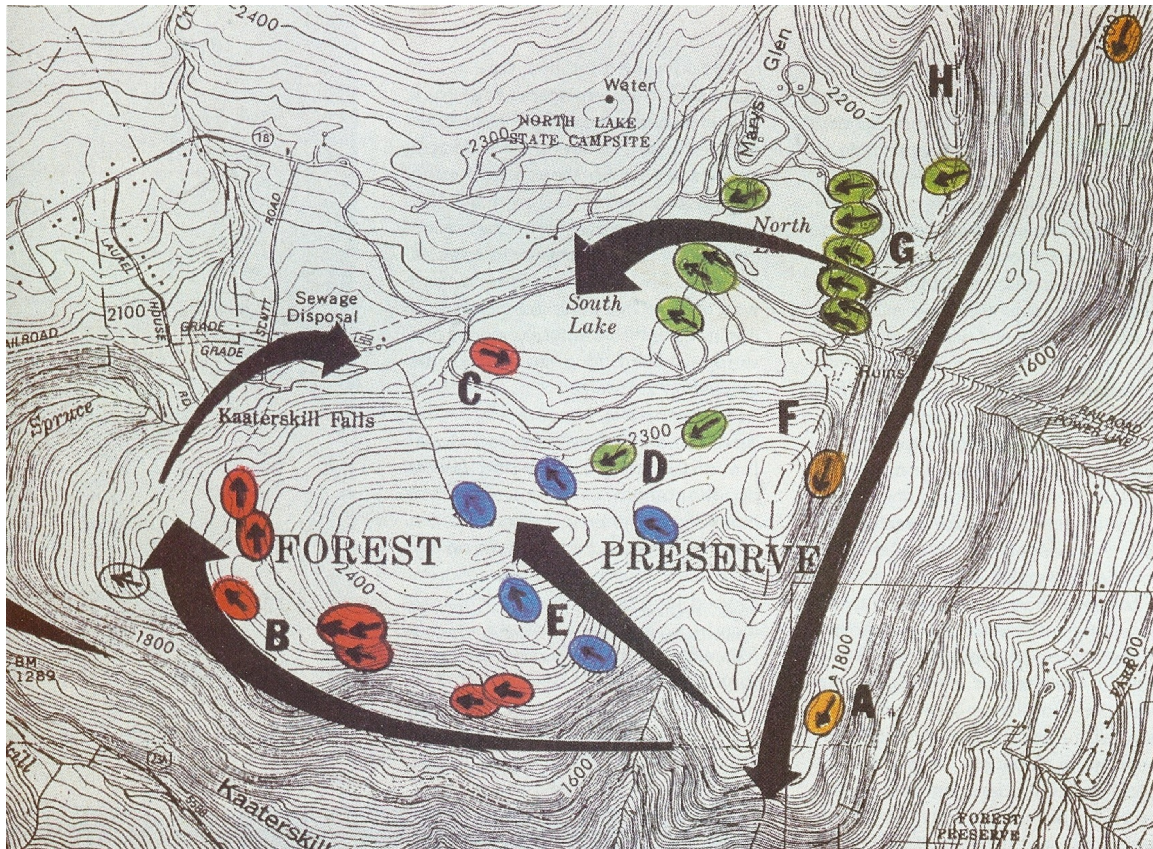


Figure 1.

Mind you, I am not arguing for abstract art, but impressionistic. It has been the case, since the later 19th Century that artists have employed impressionistic forms of painting. Here the effort is to seek to portray reality, but in a more “artistic” fashion. The purpose is to capture the eye as well as to portray, and to reveal more than to represent. The description of the advance of the glaciers, as presented above, is intended as an impression. Impressionistic literature is not new; it is a recognized genre of literature (Wikipedia, 2009).

What is the purpose of this? Should scientists even do such things? The answer is that the purpose and justification are one in the same. The purpose is to communicate, in this case, the *nature* of the Ice Age to a broad and general public.

There is, of course, a strict reality of the Ice Age and its history, and that is the largely the domain of the professional scientist. That reality is portrayed in dry, often technical prose. It can also be difficult to follow even for the experienced scientist but, in the normal context of professional science, that is acceptable. But that is *not* acceptable in popular writing. If it tends to be difficult, or even impossible to follow for the average reader, then nothing is being accomplished.

And there is so much more than the strictly realistic. There is a spirit of the Ice Age, even a romance of the Ice Age. Artists, such as Frederic Church, have attempted to capture all that in canvases such as “The Icebergs” (1861). Writers such as John Muir have sought to capture the adventure of glacial landscape (See his “Travels in Alaska,” 1915). We scientists have rarely followed such paths, and that is unfortunate.

We continue our field trip: Head south and up the Blue Trail, ascending from the Mountain House site. Continue on the Blue Trail where it forks to the left, until reaching Boulder Rock. Boulder Rock is a fine glacial erratic, brought here when the ice had covered all of South Mountain. Look also for numerous striations and a few rat tails in its close proximity. A chapter in time is found here:

There was a time when the very concept of the Hudson Valley did not mean very much and it was not all that long ago, at least not in the way that geologists think about time. Let's take a plane ride back 20,000 years ago. We are the mind's eye, the human imagination, and we can do such things. We are flying from what will someday be the village of Catskill to what will be the city of Kingston. It is a clear day and we can see all the way to the horizon in any direction that we care to look. There is, however, virtually nothing to see.

Looking straight down we see a flat expanse of white. We drop down and fly close to the “ground” (the mind's eye can do such things) and we benefit with just a little more detail. Now the whiteness is broken by a few dark fractures. We are close enough to tell that it is ice that we are looking at, and now we can also see some drifts of snow. But, for all practical purposes, this is a featureless and white Arctic landscape.

We rise up high into the sky, higher than before. Off to the east, the white extends to the horizon with absolutely no blemish. That horizon shows the Earth's curvature, but it is a white curve against a very pale blue sky. We look north and see exactly the same visage. Then we look west and there, at last, is the one blemish to the perfectly white landscape. The peak of Slide Mountain pokes above the ice; it is an island in a sea of ice.

The sight of Slide's peak is a welcome one, but this view quickly generates a rush of awe. Slide rises to more than 4,000 feet in elevation, but only a bit of its summit is showing. The conclusion is inescapable: there are thousands of feet of glacier beneath us. I used the word Arctic but I might better have called it Antarctic. There is nothing in the modern world to match what we are seeing except the vast whiteness of Antarctica. It is this notion that inspires such awe.

Millennia from this time, scientists will recognize this as one of the great glaciations of all history, and they will name this glacier the Laurentide Ice Sheet. More than half of the North American continent, on this day, is covered with ice. We are the mind's eye; we rise thousands of miles above the surface and gaze to the north. Even this high, there is nothing but white curved globe as far as we can see in that direction. Once again, a rush of awe overwhelms us.

To the south we do better. We are now high up enough to easily see the southernmost extent of the glaciation. The ice has reached into northern New Jersey and northeastern Pennsylvania. More ice has reached as far south as Long Island. Beyond the whiteness is a barren and desolate landscape. Someday scientists will called this bleak region a tundra or a “periglacial” zone.

Now we understand why, at this time, the very notion of the Hudson Valley is meaningless. All of this valley, along with the Catskill Front, is buried in the thickness of the ice we see. It gets worse: off to the east both the Taconic and the Berkshire ranges are similarly submerged in the ice.

We continue our ice age flight and drift back to the north. With our mind's eye we operate a form of radar that penetrates the ice below. We can see the Hudson Valley beneath the ice and we can see the Catskill Front and the Catskill Mountains. We are the mind's eye; we can do such things as this. (Adapted from Titus, 2008)

The “mind’s eye” is another literary device. It is freely used here along with the first person “I” and “we.” All are supernatural entities serving to transport the reader back into time. Scientists don’t use such things in professional literature, but, of course, that is the point of this effort.

There is a secondary function of the supernatural first person. Importantly, the mind’s eye serves also to emphasize that this is not professional literature. The reader understands that, and understands that this is largely fantasy, exempt from all the rules of professional literature. The reader should not be confusing any of this with professional science and that is critical.

We follow the Blue Trail west from Boulder Rock. Soon we reach the junction with the Red Trail and, just east of that junction, we find an inconspicuous notch that the Red Trail passes across. This is the very top of a substantial meltwater spillway (Figure 2). Once large volumes of water, impounded by a glacier to the north, passed this way and tumbled down an increasingly large canyon below. We can look at this feature as a modern bit of the landscape, or we can travel into the past.

This notch was once actively draining water out of an ice-bounded lake, extending off to the north. If you visit this site, look and, in your mind's eye, see that lake. All along the two shores, left and right, you are likely to see platforms of ice extending out into the lake. The middle is ice free, dark and deep. A slow current will be flowing toward you. As it approaches the notch, it is squeezed into the narrows between the rock cliffs here and it picks up speed. A silent but very powerful flow of water rushes through the notch. There is something akin to the hum of electricity here, but otherwise it is remarkably silent. (Adapted from Titus, 2009)

Off to the south, the flow quickly becomes a loud chaos. A raging, foaming cascade plummets downhill into Kaaterskill Clove. It disappears under the ice which still fills most of that deep clove. There is an enormous amount of power to all this; it comprises the very image of the end of the Ice Age. Glaciers are melting rapidly and vast amounts of water have to drain off somewhere and they too must do it quickly. (Adapted from Titus, 2001)

Turn west and continue on the Blue Trail and follow it toward the Hotel Kaaterskill site. Turn right and travel down the old unmarked hotel trail. Near the bottom of this trail we will again turn right and follow eastward on another unmarked path, keeping a sizable ledge immediately to our left. Back east, almost to the Catskill Front, are enormous boulders called the “Druid Rocks.” These were probably dislodged as the ice flowed northward, across them, and plucked them free.

We are beneath the ice. Above us, a sizable glacier is flowing northwards. This glacier has peeled off of the Hudson Valley ice and turned into Kaaterskill Clove and then curled around South Mountain (Figure 1). After rising up to that mountain's crest, it has overtopped it. The moving ice is plucking the rocks just beyond that crest and that is what we are experiencing now. We hear the low groaning sounds of advancing ice, punctuated by sporadic sharp cracking and popping sounds. The years pass by and we, the mind's eye, are able to stay here and experience all that is happening. With time, these sharp sounds abate, and eventually they cease. There has been, however, the steadily increasing gurgling and rushing sound of flowing water. The glaciers have passed their zenith, the climate has begun to warm, and the ice age will soon begin its end.



Figure 2.

Continue east until returning to the Blue Trail. Hike down to the North Lake parking lot. Towering above the lake is North Point. A great hollow up there appears to be a late ice age cirque. John Lyon Rich (1934) and George Halcott Chadwick (1944) quarreled over how many cirques were to be found in the Catskills. Rich recognized many of them; Chadwick was skeptical, believing that all or nearly all were simply the products of dendritic (he used the word arborescent) drainage. At the top of Ashby Falls in Mary's Glen two patterns of striations can be seen. One, from the east, represents ice overtopping the Catskill Front, but the other, is oriented and a manner consistent with an Alpine glaciers descending the southern slope of North Mountain, down from the cirque. We, the mind's eye, can see this Alpine glacier.

Mar. 13, 12,748 BC - *We gaze above North Lake to the North Point cirque. It has been an unusually rainy and warm early spring season, but today is the worst. A powerful nor'easter has been moving up the east coast and now intense rain pummels the North Lake vicinity. On this day there is still a glacier within the now old cirque, but its days are numbered. The climate has been warming and it has been quite some time since the ice has managed to advance down the valley of Mary's Glen. The rain has melted all snow off of the ice and what remains is a shiny beryl blue colored glacier. It is a beautiful sight, but we do not get to see it for very long. The warm rain, falling on the cold ice, is generating a ground fog. Our view disappears into that fog.*

Continue on to campsite 151. There, you will find a very fine scoured and polished ice age surface. Striations and very large crescent marks can be seen here. They display three compass orientations. The first is most poorly preserved and is oriented north to south. It appears to possibly represent the peak advance of the ice down the Hudson Valley. The second and third orientations are at west 25, degrees north, and west, 25 degrees south. These seem to represent two successive younger advances of the ice over the lip of the Catskill Front at North Lake. These can be hypothesized as being events *such as* the Wagon Wheel and the Grand Gorge re-advances, but no evidence in support of such notions can be mustered.

March 14, 12,748 BC – *The rain continues with no slackening, in fact it seems to have gotten worse. In the foreground water is running across the bare bedrock, sweeping it clear of sediment and gravel. The striations of the recent ice age epochs are clearly seen. The ground here will not “recover” from this storm; it will stay bare for many millennia and those striations will remain exposed for all of that time.*

The use of dates in these last two pieces is, of course, an utter fiction. But the intent is to add an out-of-time note of “drama” to the image. Descriptions of weather conditions, likewise, serve to add atmosphere to the imagery. The purpose is not to advance science but to advance art in its service to science. It an effort at literature.

Return to the Blue Trail and continue north toward Sunset Rock. Just short of the turnoff onto the Yellow Trail is a canyon. This is a deep jagged gorge, but today it contains virtually no water within it. It appears to be a late ice age paleoform. The floor of this gorge displays what appears to be bedrock with glacial striations oriented parallel to its direction. This would indicate that a tongue of ice split from the main Hudson Valley glacier and penetrated this canyon. The canyon itself seems to date back to just before the latest phases of the ice age. This steep vee-shaped canyon would appear to be a temporary spillway, dating back, likely, to the last melting of the Hudson Valley glacier (Figure 2). There is another similar meltwater canyon about a quarter mile to the north and, another at Rip Van Winkle ledge. More remarkably, there is a double canyon at Dutcher Pass, almost four miles farther to the north, along the Blue Trail.

July 16, 14,154 BC - *We stand in the center of the canyon in the face of an enormously powerful flow of icy meltwater. A cascade is washing by us, but we are the mind's eye and we can hold our ground. We rise up a hundred feet or so and look to the east. There, before us, the whole Hudson Valley is filled with a glacier. It extends off as far as we can see. This mass of ice is huge and it slopes upwards to the east for miles; we cannot see beyond its top. But, large as it might be, this glacier is condemned to melt away. The climate has warmed and this glacier is, right before us, turning into meltwater. That is the origin of the flow that we have just experienced.*

Return back down the Blue Trail. Follow it to the Artist Rock Ledge, overlooking the valley. From here, there must have once been quite a view at the very end of the Ice Age. Out in the distance was Glacial Lake Albany, nearer was Glacial Lake Kiskatom, and just below to the south was the Palenville alluvial fan, where Kaaterskill Creek emptied out onto the floor of the newly deglaciated Hudson Valley.

Dawn, April 3, 13,445 BC – *We stand on the edge of the Catskill Front and before us, as is so often the case, the valley is filled with fog, which reaches half way up the front to where we are. In the distance, off to our south, is the steady roar of a great flow of fog enshrouded water. What mystery is this? A few hours pass and the fog has been thinning. Now a breeze picks up, blowing along the Catskill Front. It starts to drive the mist away and we can look toward the noise and see the source of that roar.*

NYSGA 2009 Trip 5 - Titus

A great flow of whitewater is thundering out of Kaaterskill Clove and flowing through what will someday be Palenville. Fed by enormous amounts of meltwater, this is a very much larger Kaaterskill Creek than what is seen in modern times. The flow has already deposited a substantial alluvial fan. This ancestor of today's Kaaterskill Creek splits up as it passes onto the fan. There are four large distributary streams and a dozen or so small ones. Each is flowing down the gentle slope of the fan.

At the bottom of the fan, the various distributaries reassemble themselves, once again, into a single stream. This is a most remarkable manifestation of Kaaterskill Creek. On this day it is a raging flow of pounding, churning whitewater. It would be called a Class 6 stream in modern times, except that it is bigger than any seen today. We are watching at the exact day and hour that Kaaterskill Creek has more water in it than it ever had before or will ever have again. Up in the mountains the glaciers are melting rapidly. That is where the powerful flow comes from.

The breeze picks up and the fog continues to dispel. Now it can be seen that Kaaterskill Creek flows into a sizable lake, which in modern times, though now dry, is called Lake Kiskatom. Across, at the eastern side of the lake, Kiskatom's waters pass into a narrow bedrock gorge. Beyond, and out of sight, is a sizable waterfall, today called High Falls. Now the fog is gone entirely and that unveils, far out in the distance, a very large lake which fills much of the Hudson Valley. This is Glacial Lake Albany.

With the fog gone, the sun shines brightly. We now see rainbows, two of them. One rising above the upstream end of the Palenville fan, and the other, far away, rises above the location of High Falls. Powerful flows at both sites are producing large, fine sprays of mist. It is a beautiful panorama that we see on this day. In the future people will have to settle for imagining this sight, but we are geologists, and we are privileged to see it for ourselves.

Professional scientists may likely rebel against the use of exact dates for this and earlier passages, but it is, of course, another literary device. The dates reinforce the concept that there really was a past here, moments in time when events of this sort actually did happen. In most other respects this passage is probably close to being acceptable to professional science; the features described have been deduced from abundant clear evidence. The style of writing separates this passage from professional writing, again the whole point of this exercise.

We return to North Lake parking lot. Our trip is over. Our exercise today has been about popular writing much more than science. What is argued here is a doctrine of the separation of popular and professional writing. The pursuit of one should have no influence upon the other. The author disclaims that any of this composition is professional science. The author is a paleontologist, not a glacial geologist. The author stakes no "claim" to any of these subjects and encourages professionals to study at North Lake. But this composition is an appeal for other geologists to take up the writing of their sciences, in local periodicals, for the benefit of their own communities, and for the benefit of themselves. And for the benefit of science.

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Economic Geology of the Central Hudson Valley, New York

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INTRODUCTION

This trip provides an overview of the Portland cement and aggregate mining industry, past and present, in the central Hudson Valley. We will visit the Holcim US Catskill Quarry, as well as the Colarusso and Holcim US quarries on Becraft Mountain in the Town of Greenport, Columbia County. We will review the classic Tristates and Helderberg stratigraphy, see how geologic principles are applied, examine complex structural settings and see how various professions—e.g. geologist, mining engineer, chemist—work together to make viable products in as efficient, economical and environmentally friendly a manner as possible.

HISTORY OF PORTLAND CEMENT MANUFACTURE IN NEW YORK

Portland cement replaced natural cement starting in the late 1800's, primarily due to the commercial availability of rotary kilns capable of producing large amounts of cementitious material at a relatively low cost. A number of Portland cement manufacturing plants sprang up in the Hudson Valley due to: (1) the proximity to the large metropolitan New York City and other regional markets; (2) the presence of thick, fault-repeated sections of fairly high calcium carbonate limestones; (3) the abundance of clays, shales and argillaceous limestone; and (4) the presence of the Hudson River that served as an efficient transportation route to end users of the cement.

As late as the early 1950's there were at least nine cement plants in eastern New York, including: (1) at Catskill on the east side of U.S. Route 9W, operated variously by Lone Star, North American Cement and Marquette (the cement plant is currently operated by Holcim US—formerly St. Lawrence Cement and Independent Cement—and also contains an aggregate plant operated by Peckham Materials); (2) at Cementon on the east side of U.S. Route 9W, the “Alpha” Plant operated by the Lehigh Portland Cement Company (still used by Lehigh Northeast Cement to grind imported clinker and distribute cement); (3) at Cementon on the east side of U.S. Route 9W, the “Alsen” Plant operated by the Lehigh Portland Cement Company (closed in the early 1980's); (4) at East Kingston, the Hudson Cement Company plant (closed in 1982); (5) at Hudson, on the north side of N.Y.S. Route 23B, the Lone Star plant (closed in the mid-1960's); (6) at Greenport, on the west side of U.S. Route 9, the Universal Atlas Cement plant (the cement plant closed in 1976 but the quarry is still worked for aggregate); (7) at Ravena, on the east side of U.S. Route 9W, the Atlantic Cement plant (still active as a cement plant operated by Lafarge and an aggregate plant operated by Callanan Industries); (8) at Howes Cave on the west side of Sagendorf Corners Road, the North American Cement plant (the cement plant closed in 1976 but Lehigh Northeast Cement maintains cement storage silos and Cobleskill Stone operates a state-of-the-art aggregate plant); and (9) at Glens Falls on the south side of Warren Street, the Glens Falls Cement plant (still active and operated by Lehigh Northeast Cement).

Most of the cement plants once active in eastern New York have shut down, victims of outdated equipment, changing economic forces and environmental activism.

AGGREGATE MINING IN NEW YORK

There are approximately 2083 permitted and approximately 2384 reclaimed commercial mines in New York. More than 99 percent of these mines are or were aggregate mines supplying the sand, gravel, stone, clay and topsoil used to build, repair and maintain the infrastructure that is one of the foundations of the economy. In 2007, New York State's aggregate mining, blacktop production, concrete production and cement mining was responsible for:

- Direct sales of \$3.3 to \$3.5 billion

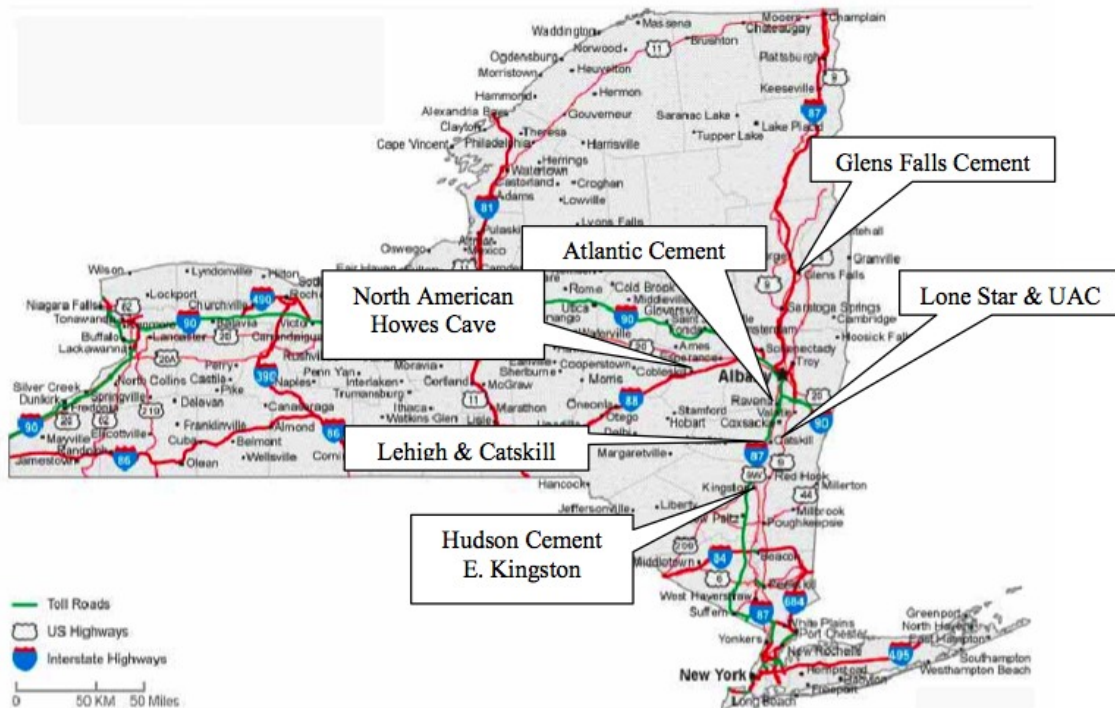


Figure 1. Locations of Cement Plants in Eastern New York in the 1950's.

- Employee wages of \$1.2 to \$1.3 billion
- Employment of 28,000 to 30,000 people in all parts of the state (average salary of all employees was approximately \$48,000, slightly above the median salary for New York industry)
- Direct payment of \$87 to \$101 million in sales tax, personal income tax, fuel tax, corporate franchise tax, and Mined Land Reclamation permitting fees.

The mining industry has a tremendous economic impact on New York State but is far from secure. Local governments frequently pass restrictive zoning, without proper consideration of the availability of mining resources, which prevents or severely limits mining. Environmental regulation is being enacted at an ever increasing rate - merely familiarizing oneself with the new regulations is a full-time job. Existing mines are inevitably being depleted and, alarmingly, not being replaced equally by new mines. Local shortages of construction materials have resulted in increased costs (hauling aggregate 20 miles roughly doubles the cost) to end users and increased transportation distances, with all the related impacts associated with heavy truck traffic. Since 2001, the number of mines in New York has decreased and the percentage of the mines' permitted area that has been worked has increased dramatically. Downstate New York is particularly hit by local shortages of concrete sand. This product typically sells for about \$8 per ton in upstate New York but sells for about \$25 per ton downstate. The increased costs are the result of shortages and increased transportation costs as the material must be hauled in from as far away as southern New Jersey, Canada, the Capital District, the Adirondacks and Central New York.

BACKGROUND

The geologic units being worked in the sites visited today are part of the Devonian Tristates and Helderberg Groups. These units represent deposition in a repeated sequence of increasingly deeper oceans. These transgressive se-

quences resulted in significantly different depositional environments that exhibit different physical and chemical characteristics that dictate the stone's use as cement or aggregate stone. Purer (higher calcium carbonate) limestones are better suited for use as cement stone and have limited uses as aggregate. Siliceous (lower calcium carbonate) limestones are better suited for use as aggregate, particularly the high friction aggregate used to provide traction on road surfaces (see Figure 7).

These geologic units have been subject to tectonic activity as part of the Acadian and Alleghenian Orogenies. At Catskill, the folding and related thrust faulting is severe. The eastern part of the quarry is characterized by repeat sections of Kalkberg-Coeymans-Manlius. The central part of the quarry is characterized by repeat sections of New Scotland-Kalkberg. The southern and western parts of the quarry are characterized by repeat sections of Port Ewen-Alsen-Becraft. Relatively undisturbed sequences of the Esopus Shale outcrops in the area west of the quarry.

The combination of physical and chemical variations in the various beds and the presence of folds and faults every 100 feet in any direction results in the Catskill Quarry's reputation as a geologist's dreamland and a mining engineer's nightmare. At the Holcim US Catskill Quarry, seven or eight imbricate sheets have been identified beginning with the outstanding work done by George Chadwick in the 1940's, and culminating with massive mapping and core drilling projects in 1985 and 2004 by Holcim and the authors.

The same geologic formations being worked at Catskill are present at the Colarusso and Holcim US's Greenport Quarries. These quarries are located on Becraft Mountain, an outlier of Lower Devonian limestones and shales that was thrust west and over the Taconic Orogeny jumbled up Ordovician shales, siltstones and sandstones.



Figure 2. Electric shovel loading rail cars on the floor of the Universal Atlas Greenport Quarry in 1936.

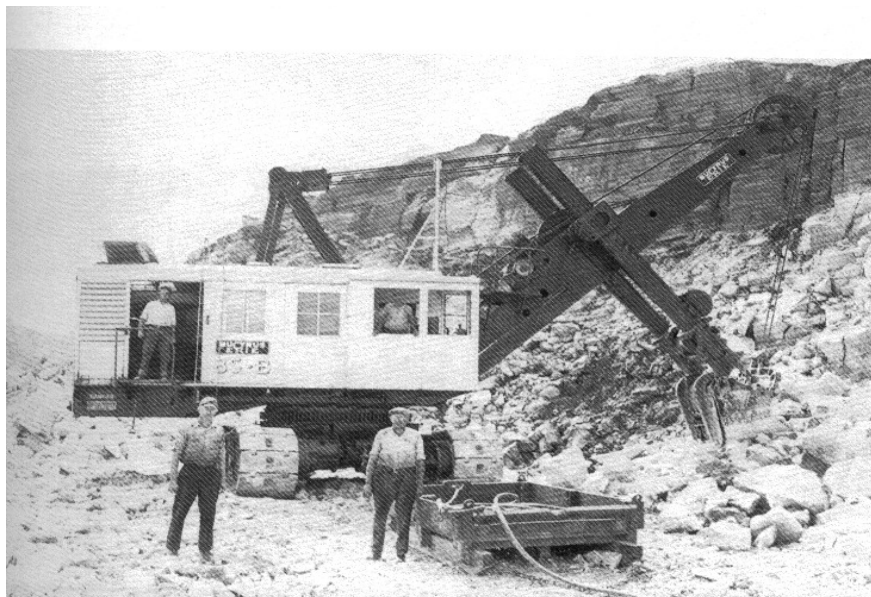


Figure 3. Electric shovel and operators at Universal Atlas Greenport Quarry in 1938.

Mount Ida, located a few miles to the northeast of Becraft Mountain, is the only other significant outlier of Devonian limestone in Columbia County.

Mining on Becraft Mountain has been ongoing since 1675. Cut stone for foundations and buildings gave way to small lime kiln operations which in turn gave way to Portland cement mining and manufacture. Cement mining was done in 1901 by Knickerbocker Cement and the Hudson Portland Cement Company. By 1930, these operations were being run by Lone Star Cement and Universal Atlas Cement. Lone Star operated a cement plant to the northeast and a quarry on the northeast side of Becraft Mountain until the mid-1960's. Universal Atlas was bought out by U.S. Steel but continued to operate a cement plant on U.S. Route 9 west of and a quarry on the northwest and center of Becraft Mountain.

Until the 1950's, Universal Atlas operated rail cars on an ever shifting series of rail lines to move stone from the quarry to the plant. The rail cars were loaded by electric shovels and hundreds of workers were employed handling electrical cable and moving the rail lines as the quarry faces advanced (see Figures 2 to 6).

ROAD LOG AND STOP DESCRIPTIONS

Meet at the parking lot of the Home Depot in Catskill near Exit 23 of the Thruway at 9:30 am. To get to the Home Depot, take the N.Y.S. Thruway to Exit 21 (Catskill/Cairo/N.Y.S. Route



Figure 4. Moving railroad track at Universal Atlas Greenport Quarry in 1937.

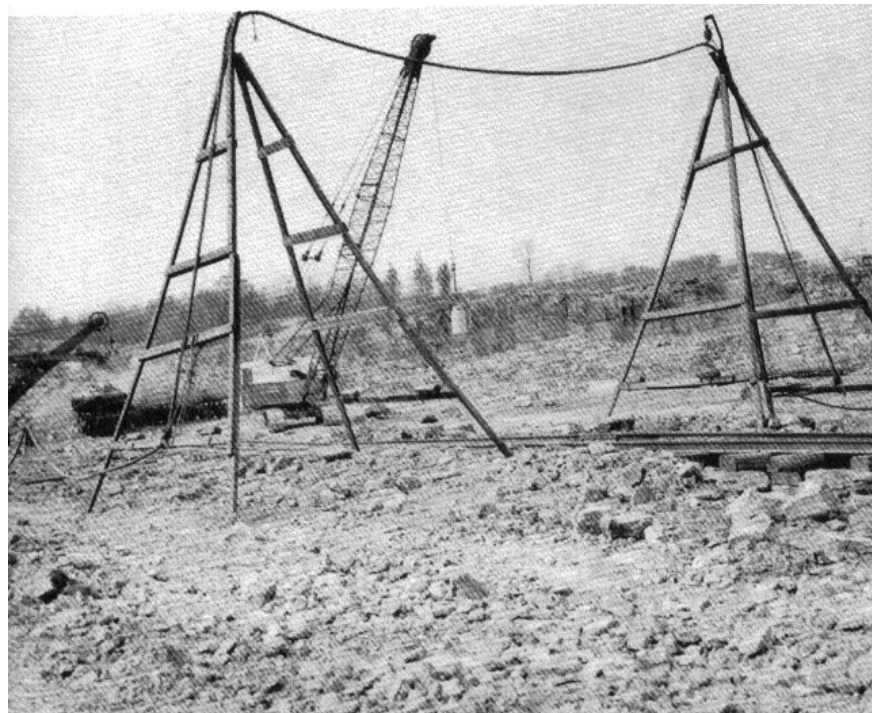


Figure 5. Electric cables and drop ball in Universal Atlas Greenport Quarry in 1953. The drop ball was state-of-the-art equipment at the time, replacing the “blast agent monkeys” whose responsibility it had been to hand drill holes in several pieces of shot rock too big to be put into the crusher, load these holes with sticks of dynamite, light the sticks of dynamite and run like crazy.

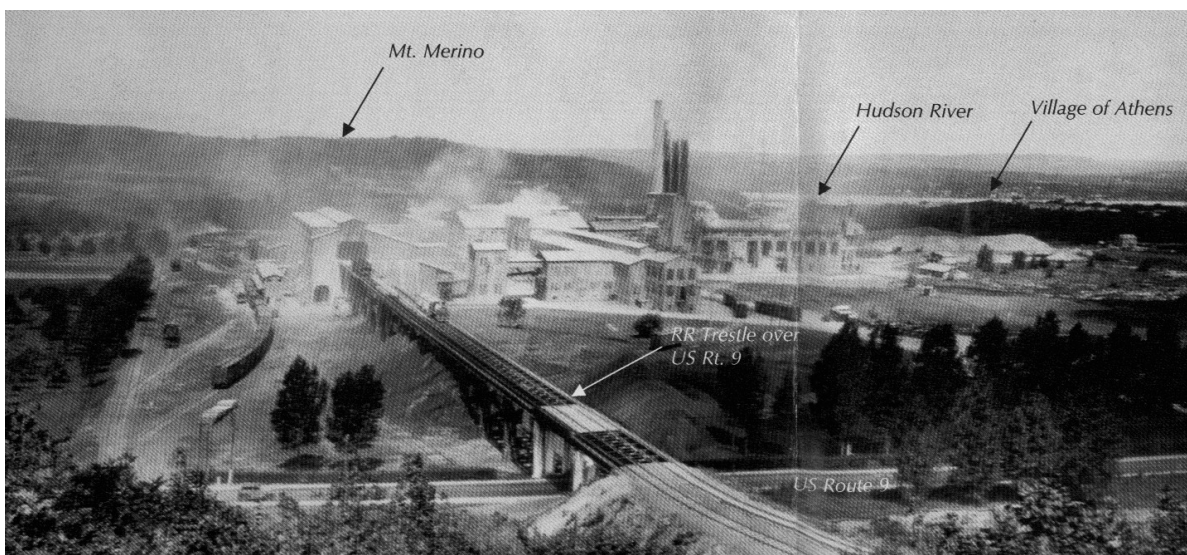


Figure 6. Cement plant on the west side of U.S. Route 9 in Greenport in 1925. Note the trestle bridge over the highway that brought stone to the plant by rail.

23). Note the benign looking outcrops of Becraft Limestone (every cement producer’s desire) on the west side of the off-ramp. Stop and pay toll. Go 0.1 miles beyond toll booths to intersection with N.Y.S. Route 23. Go straight across the highway to the Home Depot entrance road. We will assemble on the left (southeast side) of the parking lot. Note the outcrops of New Scotland Argillaceous Limestone on the southeast side of the parking lot.

After assembling at Home Depot, we will drive to our first stop, the Holcim US Catskill Quarry.

Mileage	Route Description
0.0 0.0	Return to N.Y.S. Route 23 and turn left, note rip rap drainage stone in roadside ditches and black-top in road, both produced at the Catskill Quarry.
0.3 0.3	Go straight and continue on County Route 23B. The classic Jefferson Heights road cut is on the left on the N.Y.S. Route 23 off ramp.
1.6 1.3	Turn right onto U.S. Route 9W south. Bridge crosses the Catskill Creek. Go straight up the hill underneath the aged railroad trestle.
2.4 0.8	Junction with N.Y.S. Route 385. Veer slightly to the right, staying on Routes 385 and 9W south.
2.7 0.3	Junction with N.Y.S. Route 23A, keep going straight.
3.3 0.6	Turn left, staying on U.S. Route 9W south.
6.6 3.3	Follow Route 9W south to the entrance to the Holcim US Catskill Quarry on the right. Do not go into the Peckham Materials entrance! We will park our cars in the unpaved lot on the left side of the entrance road and car pool for stops in the quarry.

The haul roads, location of quarry faces and various hazards frequently shift in an operating quarry. Therefore, we will drive by convoy to each stop. Please bring your lunches.

STOP 1. Tracey’s Landing Quarry. Historically, the quarry operators gave numerical names to their quarries. This practice became difficult at the Catskill Quarry as the property was an amalgamation of at least three different quarries and each company often had the same name on different quarries. Beginning in the early 1990’s quarry nomenclature reached an art form under the guidance of Richard Playle and Dave Blasko, mining engineers, Jim Sucke, Murray Craft and Floyd Mower, quarry managers and Paul Griggs, consulting geologist. We became frustrated at trying to use ever-changing landmarks in planning discussions so a systematic nomenclature system based on geologic structure was devised (see Figure 8).

The Tracey’s Landing Quarry was named after Bart Tracey, a mining engineer and quarry manager who worked at the quarry in the early 1990’s. He had done some of the original chip sampling exploration in this area and was one of the first people to suggest this area as a viable source of cement stone.

Overall, the geologic structure in the Tracey’s Landing Quarry is a rolling, doubly-plunging asymmetrical, thrust-faulted syncline. The east limb of the syncline is nearly vertical and continues east down the slope overlooking U.S. Route 9W. The center of the syncline was typically capped by siliceous limestones of the Glenerie and Port Ewen Formations overlying the cement-friendly Alsen Limestone and the cement mother lode, the Becraft Limestone. The west limb of the syncline contains the argillaceous limestones of the New Scotland Formation.

This quarry was worked by removing the argillaceous limestones from the underlying Alsen and Becraft. The “overburden” rock was used by Peckham Materials as high friction aggregate. This quarry served as the main source of cement stone for about 10 years.

STOP 2. Mower’s Lake. This quarry is part of the Apple Orchard and was named for Floyd Mower who worked his way up from an equipment operator in 1970 to Quarry Manager in 2003. The quarry serves as a sump for the Apple Orchard which has been a major source of cement stone since the early 1990’s.

A series of pumps discharge water from the quarry so that it can be worked dry. The water is used in the cement plant for thermal cooling. Overall, the structure in this area consists of a southward-plunging asymmetrical, heavily faulted syncline-anticline pair.

Unit	Depositional Environment	Thickness at Catskill Quarry	Thickness at Greenport	Primary Use
Onondaga Limestone	Varies	Up to 30'	Up to 50	Aggregate (not used yet)
Tristates Group	Schoharie Calcareous Grit	190	90	Fill
	Esopus Shale	200	190	Fill
	Glenerie/"Oniskany" Siliceous Limestone	15	10	Aggregate
Helderberg Group	Port Ewen Argillaceous Limestone	25-30	35	Aggregate/Cement
	Alsen Limestone	30		Cement
	Becraft/"Becraft Contact" Limestone	65	70	Cement
	New Scotland Siltstone-Argillaceous Limestone	75	70	Aggregate
	Kalkberg Argillaceous Limestone	75	75	Aggregate
	Coeymans Limestone	17	10	Cement
	Manlius Limestone	50	50	Cement
Rondout Shale/Sandstone/(Silurian)	Lagoonal/Varies	10-30	Up to 10'	Causing Miners' Curses

NOTE: Geologic nomenclature in quotations is based on quarry usage.

Figure 7. Stratigraphy of the Catskill and Greenport Quarries.

Holcim US Catskill Quarry Locator Map

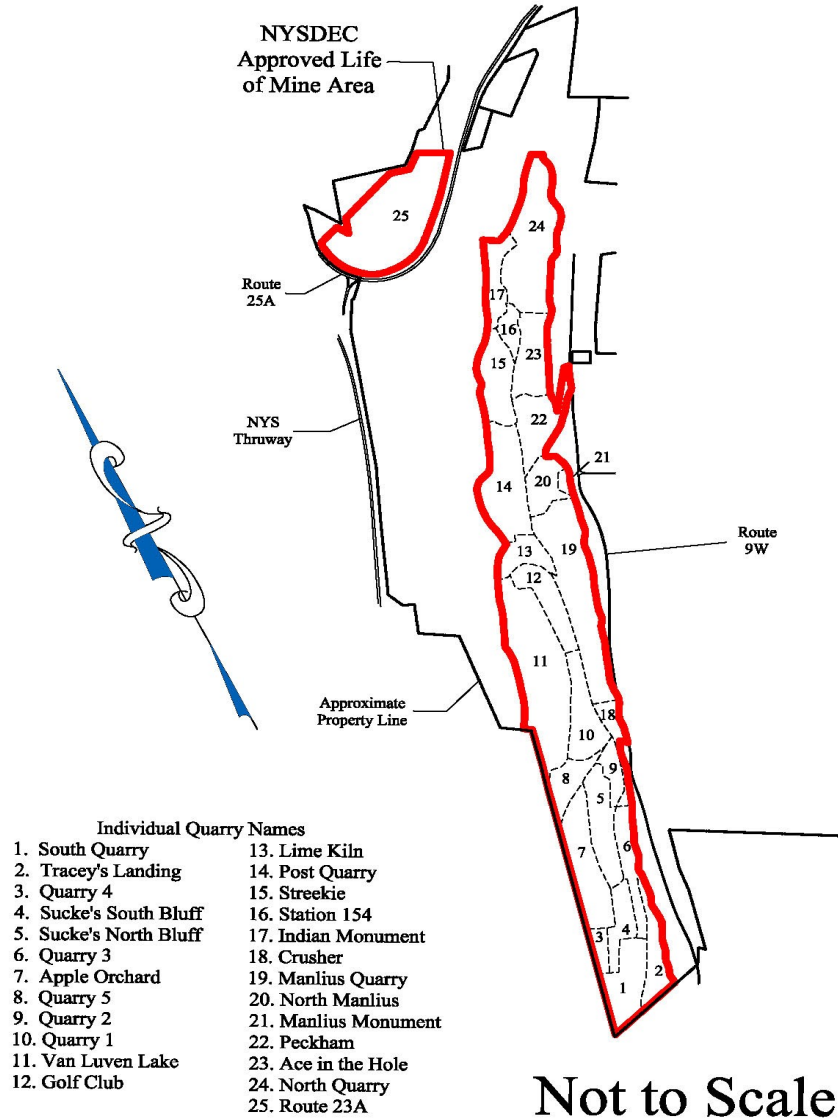


Figure 8. Holcim US Catskill Quarry Locator Map.

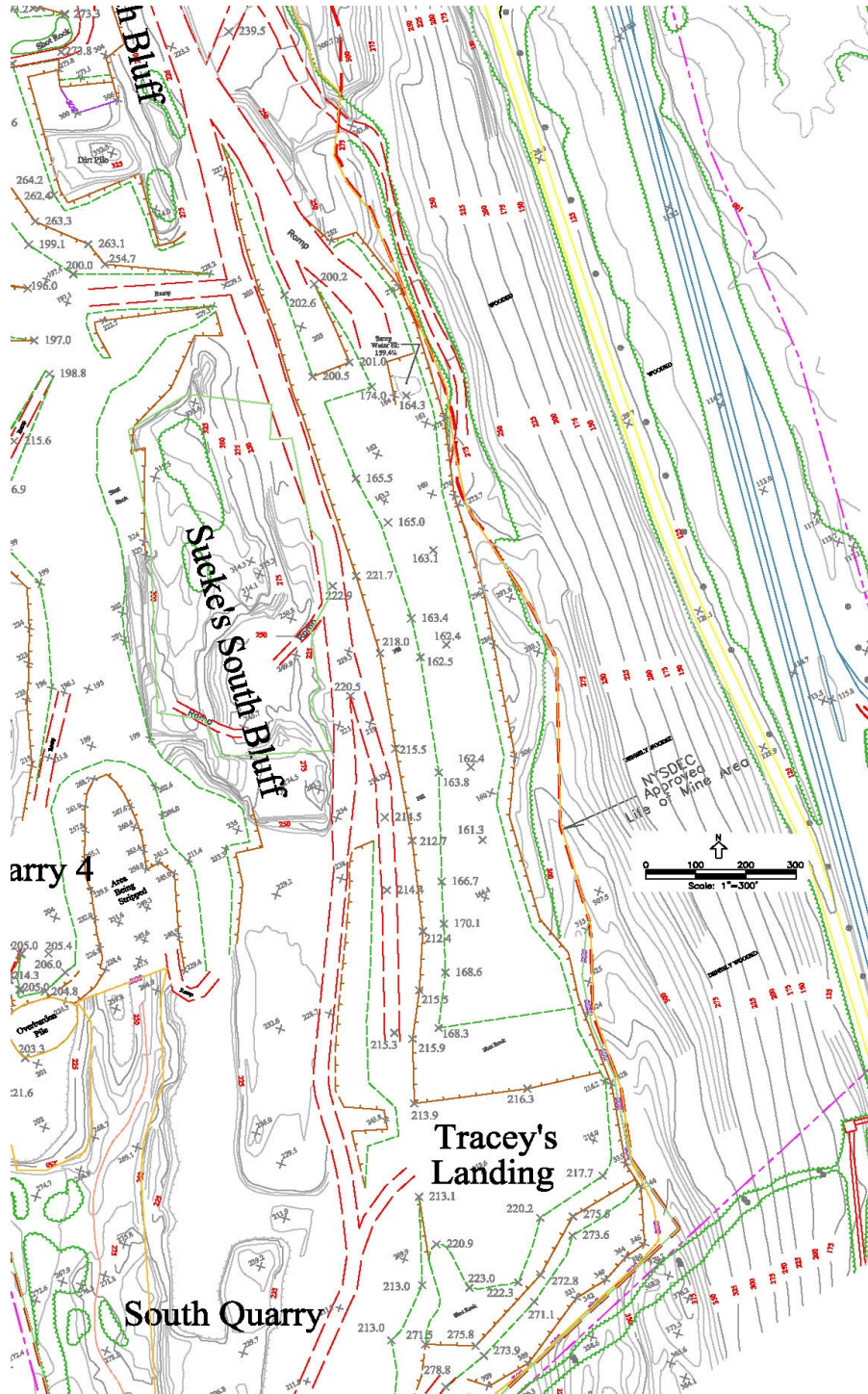


Figure 9. Map of the Tracey's Landing Quarry.

STOP 3. Floyd’s Folly. The aforementioned Floyd Mower contributed greatly to the economic viability of this quarry which is in the final developmental stages for full cement stone production. In the 1980’s the quarry manager was considering plans for filling this area with overburden rock (this was prior to the involvement of the aggregate producer, Peckham Materials). Floyd, who often operated the bulldozer and was a pretty good amateur geologist, pointed out the presence of near vertical beds of Becraft Limestone in the floor and convinced the powers that be to not cover this area.

STOP 4 Apple Orchard. The Apple Orchard Quarry was developed in the early 1990’s and has been a major supplier of cement stone since that time. The cement stone reserves were first evaluated by George Chadwick in the late-1940’s but were not mined due to the thickness of overburden rock overlying the fault-repeated thick section of cement stone.

In 1988, a geology student named Bill Reinhart worked as an intern at the quarry and researched the prior studies, did some field mapping and made some logical conclusions based on his work. There is an infamous memo in the quarry files from a bemused high level manager that questioned how a geology student

could find such a large reserve of cement stone so close to the primary crusher when everyone else had been searching all over the quarry. The answer is, like the quarry geology, fairly complex: (1) the other searchers did not have the requisite geologic expertise; (2) until the authors’ involvement at the quarry, geologic investigation had largely been confined to specific areas with little or no formal correlation between individual studies; (3) the task of map making was more time consuming in past days and there were few landmarks that could be used to accurately determine the locations of individual study areas; (4) the text and large scale plans in the geologic reports had gotten separated over the course of the numerous quarry sales and land swaps; and (5) the geology is insanely complex.

The development of the Apple Orchard was a turning point in the history of the quarry. This quarry led to the involvement of Peckham Materials, starting in 1991, to remove the aggregate stone and led to the ultimate geologic investigation and mine planning of the entire quarry.

Overall, the structure in this quarry consists of an asymmetrical, doubly plunging, rolling, highly faulted anticline-syncline pair. The west limb is nearly vertical and the east limb is bounded by the most prominent fault in the quarry, the Sucke’s Bluff (or “Sucke’s”) Fault. This fault has been traced the entire length of the quarry and onto the Lehigh Northeast Cement quarry to the south. The dark colored stone to the left (east) is New Scotland argillaceous limestone and the stone underlying the fault is mainly Alsen and Becraft limestone.

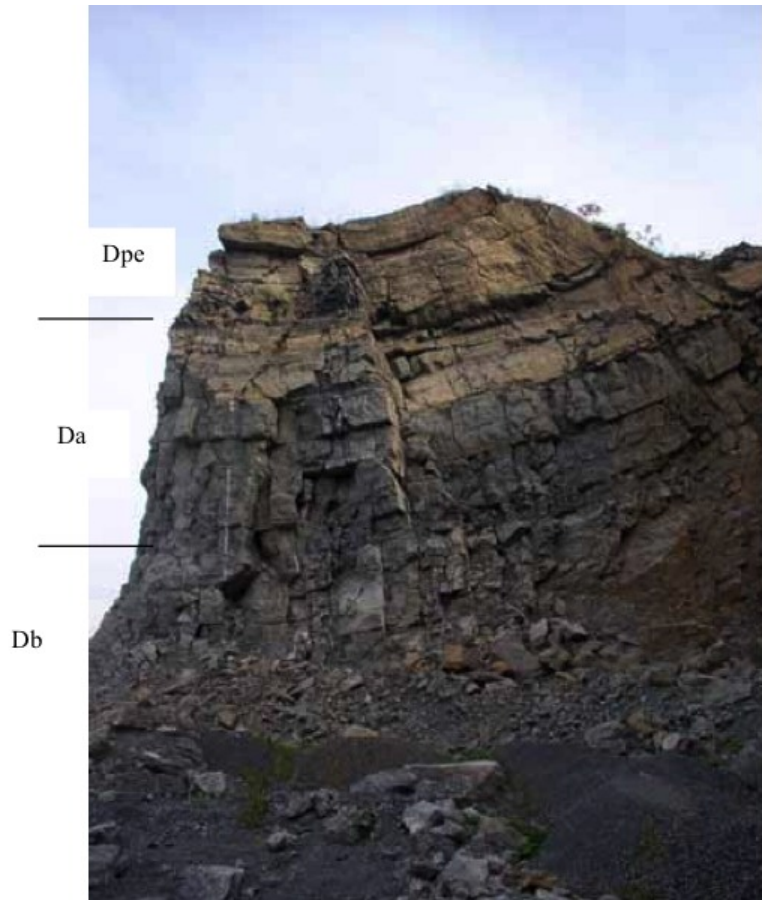


Figure 10. Photo of the north face of the Tracey’s Landing Quarry showing the center of the syncline. The dark colored argillaceous limestones of the Port Ewen Formation cap the face. The middle of the face contains the Alsen Limestone and the light-colored stone in the lower third of the face is the Becraft Formation.

Note in driving to the next stop that the main haul road parallels the path of Sucke's Fault and is mainly underlain by thick repeat sections of aggregate grade rock.

STOP 5. Golf Club. The Golf Club Quarry was last mined more than 50 years ago. At that time, a public road ran through the middle of the quarry. The northward advancing face containing high grade Becraft Limestone at the surface neared the road so the quarry operators started moving in a westerly direction, giving the quarry its distinctive shape. Using much more primitive blasting methods than employed today, they opened up large fractures in the rock that allowed a steady stream of water to flow into the quarry, effectively ending excavation operations at that time.

Overall, this quarry is part of the east limb of a faulted, fairly symmetrical syncline bounded on the east by Sucke's Fault.

STOP 6. Streekie Lake. The Streekie Lake Quarry is a current and future major source of cement stone. Named for a prior owner and its tendency to hold water during periods of wet weather, the quarry structure consists of the west limb of a faulted anticline and a complexly faulted anticline-syncline pair. In general, the top of the face contains Port Ewen siliceous limestone that is removed for use as aggregate overlying repeated sections of Alsen and Becraft limestone to great depth.

The hydrogeology in this area is interesting. The upland areas to the west are underlain by till-



Figure 11. Photo of the north face of Mower's Lake showing the complex folding and thrust-faulting typical of this quarry. The dark colored rock at the upper right of the face is from the lowest part of the New Scotland Formation. It occurs in a sheet that has been thrust over a complex mélangé of Port Ewen-Alsen-Becraft.



Figure 12. Photo looking south-southeast showing the general expression of Sucke's Fault in the east wall of the Apple Orchard.

covered Devonian shales and a large drainage basin feeds to a sinkhole just west of this quarry. During dry weather, the sinkhole may be dry or is surrounded by a small pond a few feet across. This pond grows dramatically during wet weather (see Figure 14).

STOP 7. Ace in the Hole Quarry. Most of the aggregate stone used by Peckham Materials is blasted, excavated and hauled to their surge pile by Holcim US. Peckham started the Ace in the Hole Quarry in the early 1990's to cover times when the cement company could not haul enough stone or the needed quality of stone to meet the aggregate demand.

This quarry consists of a highly faulted, westward dipping west limb of an anticline primarily containing repeated sections of New Scotland and Kalkberg argillaceous and siliceous limestone. Holcim US has removed Coeymans and Manlius Limestone from the east wall. The west side of this quarry is bounded by Sucke's Fault (see Figure 15).

The north-south trending Sucke's Fault typically dips to the west at about 60 to 80 degrees. However, the fault plane tends to wander at times and varies from slightly overturned to as low as 20 degrees. The two sides of the fault tend to widen in areas where the fault begins to rotate and larger fault slivers are common. The fault acts as a barrier to east-west flow of water and locally promotes north-south flow. Large doubly-terminated, internally flawed quartz crystals (nicknamed "Catskill" diamonds) are periodically found near this fault (see Figure 16).

We will stop for lunch at this location before returning to the parking lot and proceeding to the Greenport Quarry.

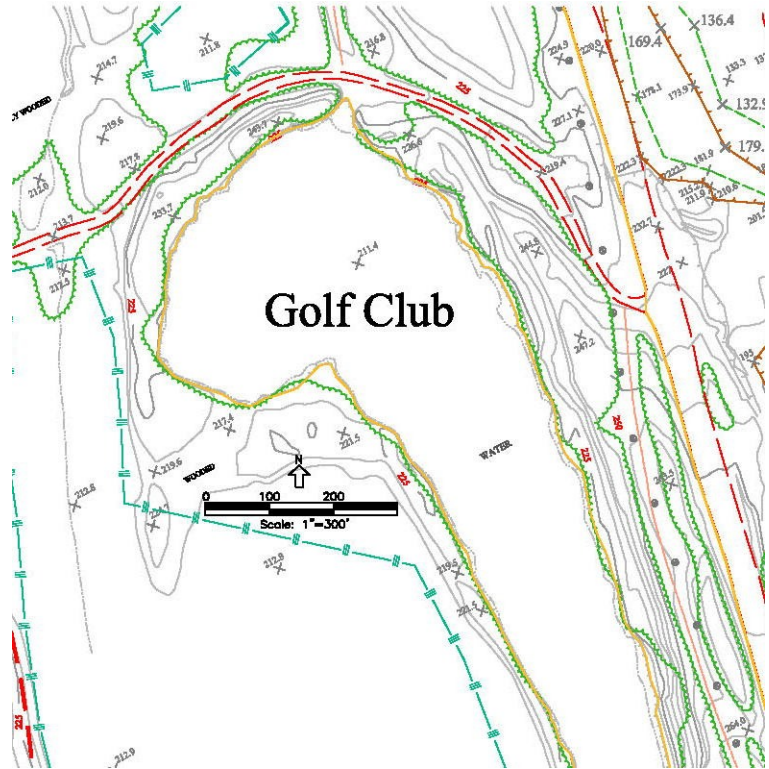


Figure 13. Map showing the northern end of the Golf Club Quarry.



Figure 14. Photo looking northwesterly of the pond backed up behind the Streekie Lake sinkhole which is located in the lower right portion of the photo.

Mileage	Route Description
0.0	0.0 From the Catskill Quarry entrance, turn left onto U.S. Route 9W.
3.4	3.4 Turn right onto U.S. Route 9W and N.Y.S. Route 23A.
4.0	0.6 Turn right onto N.Y.S. Route 385 heading towards the Village of Catskill.
4.9	0.9 Route 385 becomes Bridge Street, crosses the Catskill Creek and climbs up the hill out of the village. Turn left, staying on N.Y.S. Route 385 (Spring Street), following signs for the Rip Van Winkle Bridge.
5.5	0.6 Turn right onto N.Y.S. Route 23 east to the Rip Van Winkle Bridge.
5.9	0.4 Stop and pay toll.
10.1	4.2 Cross the Rip Van Winkle Bridge. Note that any of the islands in the Hudson River contain sand dredged for channel maintenance. The home of Frederic Church of the Hudson River School of artists), Olana, is visible on the hill directly to the east at the end of the bridge. This route passes over a mix of Ordovician age shales. Go past Columbia Greene Community College on the right and the intersection with U.S. Route 9. Turn left onto Fingar Road about 0.3 miles beyond the intersection with Route 9.
10.8	0.7 Drive along the south side of Becraft Mountain on Fingar Road. Turn left onto Newman Road.
12.9	2.1 Turn right into the parking lot and office of A. Colarusso and Son.



Figure 15. Photo looking northerly of chevron folding in Sucke’s Fault in the Ace in the Hole Quarry.



Figure 16. Photo showing doubly-terminated quartz crystal (“Catskill” diamond) found in Sucke’s Fault. The crystal is approximately 11 inches long.

The haul roads, location of quarry faces and various hazards frequently shift in an operating quarry. Therefore, we will drive by convoy to each stop.

STOP 8. Overview of Aggregate Processing and Blacktop Plants. Like the Catskill Quarry, mining activities at Becraft Mountain have included both cement stone mining and mining for the production of construction aggreg-

ates. The differing end products actually result in beneficial production logistics, as the rock formations suitable for construction aggregates are generally not suitable for cement manufacturing and vice versa.

Crushed stone aggregate [any of several hard, inert materials, such as sand, gravel, slag, or crushed stone, used for mixing with a cementing or bituminous material to form concrete, mortar, or plaster; or used alone, as in railroad ballast or graded fill] is a finite, non-renewable natural resource which is essential in the construction and maintenance of roads, industrial development, building structures, airports, railways and dams and should be recognized as an important component of any comprehensive land-use or resource management program.

The New York State Department of Environmental Conservation Division of Mineral Resources website states that each person in New York consumes about 50 pounds of mineral products per day [NYSDEC Division of Mineral Resources <http://www.dec.state.ny.us/website/dmn/rocktalk.htm>]. This amounts to approximately 175,000,000 tons of mineral products consumed per year in New York State. Most of this consumption comes in the form of construction materials such as those being produced at the A. Colarusso & Son, Inc. (Colarusso) quarry on Becraft Mountain.

Construction aggregates are mechanically processed by a combination of crushers (primary, secondary and, often, tertiary), screens and conveyor systems. These mechanisms serve to reduce, separate and stockpile rock materials that have previously been removed from the quarry by blasting.

A high percentage of blacktop and Portland cement concrete is composed of aggregate: approximately 94 percent of asphalt pavement is aggregate and 80 percent of concrete is aggregate. Due to the high percentage of aggregates in asphalt and concrete every mile of interstate highway contains 38,000 tons of aggregates and about 400 tons of aggregates are used in construction of the average

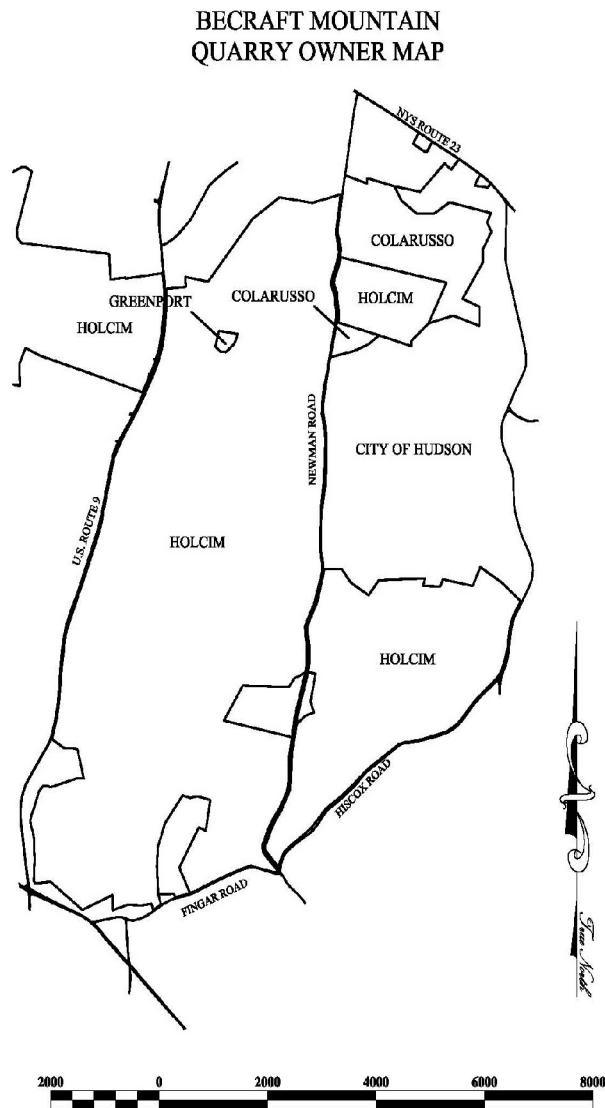


Figure 17. Becraft Mountain Quarry Owner Map.

home [National Stone, Sand and Gravel Association website: <http://www.nssga.org/>].

Note that approximately 25 percent of cement used on the east coast is imported.

The properties that make a particular rock unit attractive for use as a construction aggregate can make it undesirable for use in cement production. Cement stone manufacturing generally targets pure or relatively pure limestones - limestones with high calcium oxide (CaO) contents. Extensive mapping, core drilling, sampling and chemical analysis is performed to confirm a rock unit's suitability for use in cement manufacturing.

Although CaO is the most important component of limestone used in Portland cement, other oxides such as iron, silicon and aluminum among other components must be present in the correct proportions. The raw materials (limestone, shale, iron ore, flyash, bauxite, etc.) are crushed and ground to a fine powder and fed to a kiln that heats the powder to about 2700 degrees F, breaking down the chemical bonds. As the material cools, the resulting chemical reaction forms an intermediate product called "clinker". The clinker is finely ground along with gypsum to make the finished cement.

By contrast, construction aggregate producers in limestone terrains target rock units with high acid insoluble residues (AIR). The geologic formations on Becraft Mountain consist of interbedded high and low residue limestones. Colarusso preferentially removes some of the high residue limestones, leaving the lower residue limestones for Holcim. In general, Colarusso mines the geologic units (New Scotland and Kalkberg Formations) containing high acid insoluble residue (AIR, usually in the form of

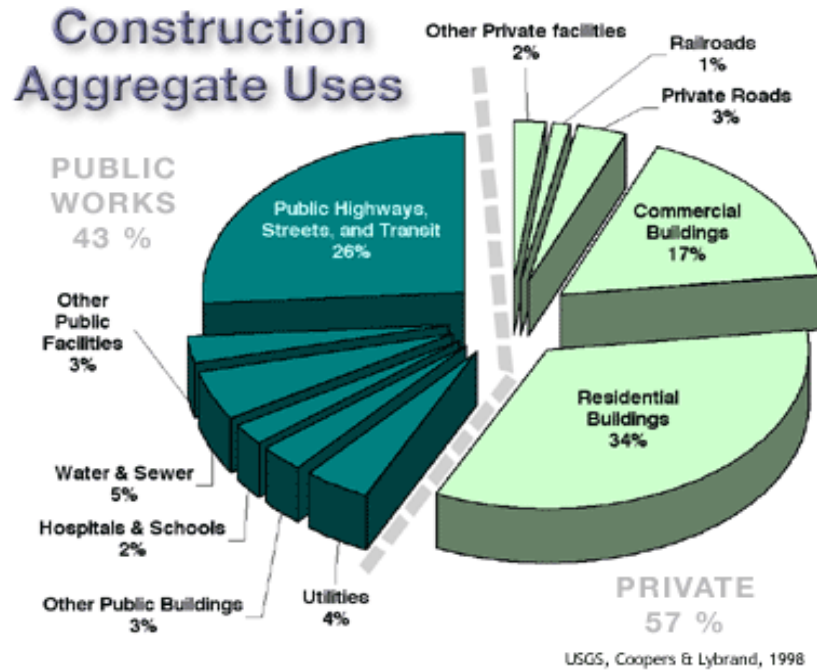


Figure 18. Construction Aggregate Uses.

The raw materials (limestone, shale, iron ore, flyash, bauxite, etc.) are crushed and ground to a fine powder and fed to a kiln that heats the powder to about 2700 degrees F, breaking down the chemical bonds. As the material cools, the resulting chemical reaction forms an intermediate product called "clinker". The clinker is finely ground along with gypsum to make the finished cement.



Figure 19. Photo showing feed hopper of primary crusher at aggregate processing plant at the Colarusso quarry on Becraft Mountain.

SiO₂, Fe₂O₃ and Al₂O₃) for high friction aggregates. Example oxide analyses for sampled rock units are shown in Figure 21..

STOP 9. Thrust Faulting on East Side of Becraft Mountain.

The geologic structure of Becraft Mountain is a doubly plunging, asymmetrical, faulted, north-northeast trending syncline. As a whole, the structure in the eastern portion of the syncline is more complex, owing to the orientation of the stresses impart by continental collisions (i.e. the Acadian Orogeny). The east limb dips at high angles to the west and is highly contorted and thrust faulted. The center of the



Figure 20. Photo showing blacktop plant at Colarusso quarry on Becraft Mountain.

syncline is contorted and thrust faulted. The west limb of the syncline dips to the east at a few degrees. A strong basal, or sole, fault separates the carbonates from the Ordovician shales they overrode. Numerous thrust faults in the eastern portion of the syncline result in an imbricate arrangement of sheets and repeated sections of rock units.

STOP 10. Overview of Mine Impoundment and Description of Past Mine Plans. Universal Atlas Cement (UAC) and their predecessors mined the Holcim property, primarily on the west side of Newman Road, for cement stone until 1976. This portion of Becraft Mountain contains the gently easterly dipping western flank of the asymmetrical

Ft	SiO2	Al2O3	Fe2O3	CaO	MgO	SO3	Na2O	K2O	TiO2	P2O5
0'-11'	35.80	3.58	1.92	30.78	1.60	0.47	0.06	0.73	0.19	0.06
11'-23'	35.73	4.54	1.97	30.12	1.56	0.34	0.06	0.94	0.22	0.08
23'-35'	30.65	6.48	2.34	31.07	1.52	0.39	0.06	1.43	0.29	0.07
35'-47'	30.75	4.44	1.65	32.77	1.52	0.27	0.06	0.92	0.20	0.10
47'-59'	28.29	1.87	1.02	37.55	1.08	0.38	0.06	0.36	0.11	0.05
59'-71'	8.56	1.83	1.12	47.00	1.53	0.59	0.07	0.43	0.11	0.06
71'-83'	8.20	1.59	0.70	46.84	1.97	0.26	0.07	0.37	0.11	0.04
83'-93'	7.44	1.58	0.53	47.94	1.57	0.16	0.07	0.39	0.11	0.05
93'-100'	9.64	2.04	0.76	44.88	1.98	0.17	0.07	0.49	0.12	0.04

Figure 21. Example oxide analysis of chip sample hole.

syncline and includes Helderberg Group units from the Alsen-Becraft down to the Rondout Formation. This area represented some of the thickest exposures of cement stone (Alsen, Becraft and Becraft-Contact) on Becraft Mountain and was the logical area for cement stone quarrying operations to be concentrated. The initial quarrying occurred down to an elevation of approximately 285 to 295 feet above sea level (asl).

UAC developed a drop-cut down to an elevation of about 245 feet (75 m) starting in the late 1950's in order to mine out a wedge of cement stone (Becraft and Becraft Contact) left in the floor. The cut began in the northern end of the property and proceeded southward. Ex-

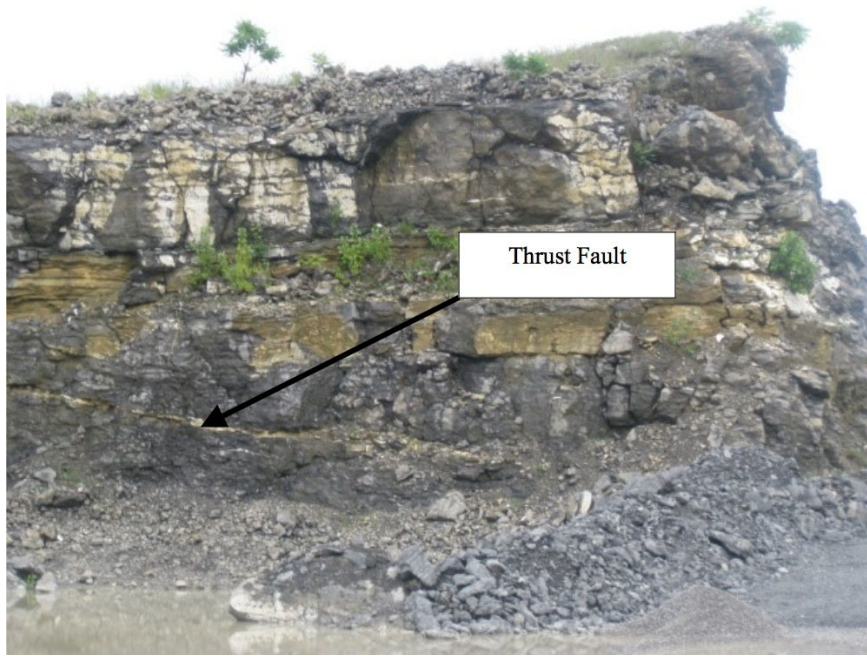


Figure 22. Photo looking east showing end-on view of thrust fault in Kalkberg Formation on east side of Colarusso Quarry.



Figure 23. Photo showing water-filled (mine impoundment) at the Greenport Quarry. The excavated rock from the mined area in this photo was used largely for cement stone manufacturing.

cavation for cement stone (Becraft and Becraft contact) was limited from advancing to the full width of the first cut owing to the easterly dip of the western flank of the syncline. This resulted in a pinching out of the cement stone (Becraft and Becraft Contact) to the west. Mining by Universal Atlas ceased in 1976 and the excavated drop-cut slowly filled with water.

The upper-lift quarry faces westerly of the water-filled drop cut (mine impoundment) include Becraft Contact over the New Scotland Formation. The uppermost quarry lift has been advanced to the west in recent years to prepare the area for deeper excavation.

STOP 11. Overview of Mine Impoundment and North Quarry.

Presently, Colarusso is advancing a cut south from the North Quarry towards the area to the west of the mine impoundment. Colarusso is quarrying the New Scotland Formation and upper part of the Kalkberg Formation for use as construction aggregate. Ultimately, the underlying Kalkberg Formation will be removed as another lift, exposing the Coeymans and Manlius Formations for future cement-stone quarrying.

The North Quarry is being advanced toward the south with a floor elevation of approximately 235 to 245 feet asl. This is some 15 to 25 feet below the elevation of the water surface in the mine impoundment (260 feet asl). The aggregate-grade stone (the New Scotland and the Kalkberg formations) extends down to an elevation of approximately 150 or 160 feet asl in the area immediately west of the impoundment. The underlying cement stone (the Coeymans and Manlius Formations) extend down further to an elevation of approximately 90 or 100 feet asl. The ability to mine to these depths is predicated upon the hydraulic connection (or lack thereof) between the impoundment and the excavated cuts, unless the impoundment is first drained. The North Quarry is not pumped, and the elevation difference between the North Quarry floor and the water level in the adjacent mine impoundment indicates that the intervening rock is comparatively tight (see Figure 24).



Figure 24. Photo showing differing elevations of mine impoundment surface (260 feet asl) on right and dry North Quarry floor (235 to 245 feet asl) on left.

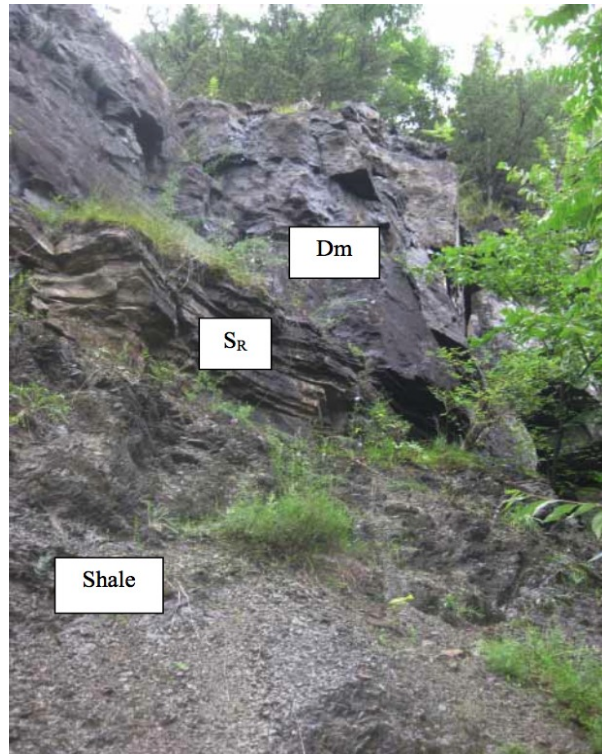


Figure 25. Photo of showing angular unconformity between the Ordovician shale and the overlying Silurian Rondout Formation.

STOP 12. Angular Unconformity and Old West Quarry. The angular unconformity (Taconic Orogeny) between the Ordovician shales and the overlying Silurian Rondout Formation is evident on the west side of Becraft Mountain in an outcrop south of the old West Quarry. The West Quarry was mined for cement stone (Manlius Formation) during the early and middle parts of the past century.

During the Late Silurian and Early Devonian, warm, shallow seas bordered by lands of relatively low relief largely covered New York State. In the Greenport area, the first sediments were deposited in very shallow, quiet (lagoonal), highly saline waters. These sediments formed the relatively fossil poor, thinly bedded, impure dolomites and calcareous shales typical of the *Rondout Formation*. In the Greenport area, sandy portions of the Rondout likely represent near shore sand bar deposits in the generally lagoonal environment.

The Early Devonian seas advanced and covered previous landmasses due to a rise in water level or a downwarping of the land. Water circulation within the deeper oceans gradually increased and living conditions generally improved. Examples of the numerous invertebrates (including brachiopods, trilobites, crinoids, gastropods and bryozoans) that lived in the Early Devonian seas are preserved in the numerous fossils and fossil fragments in the quarry stone.

The *Manlius Formation* was deposited in slightly deeper water than the Rondout. The Manlius sediments were deposited in very shallow, generally quiet, somewhat saline (lagoonal) water and lithified into typically fine-grained, uniformly bedded limestones. Small reefs called biostromes occur scattered throughout the Manlius. The reefy parts of the Manlius were deposited in shallow, clear, agitated waters and typically consist of medium-grained, irregularly bedded limestones. Ripple marks in the Manlius Formation are evident on the sloping floor of the West quarry (see Figure 26).

Cement stone was crushed on site and transported by train and trestle over U.S. Route 9 to the Universal Atlas Cement Plant (see Figure 6). The railroad grade and crusher foundation is still evident in the area west of the notch (see Figure 27).



Figure 26. Photo of ripple marks in Manlius Formation on floor of West Quarry. Arrow shows inferred flow direction.



Figure 27. Photo of concrete foundation (c. 1937) of crusher used to reduce cement stone for use at Universal Atlas' cement plant in Greenport, New York.

STOP 13. South End of Mine Impoundment. Large reserves of aggregate and cement stone remain on Becraft Mountain. Excavation will largely be done below the surface elevation of the water in the impoundment. Future quarrying will eventually necessitate dewatering of the quarry as excavation proceeds to depth. Proper planning necessitates that operators gain a handle on what quantities of water will necessarily be pumped from the quarry to maintain dry operating conditions.

The Greenport Quarry is located within a comparatively small drainage basin. Precipitation, where not falling within the internally draining quarries, tends to run off the site in a focused, but radially outward manner. Water budgets have been performed to assess what fractions of incident precipitation will be lost to evapotranspiration, surface runoff and infiltration. The constantly changing quarry configuration(s) result in altered hydrologic boundaries. The infiltration (always) and runoff (sometimes) components of the incident precipitation must be managed by the operator, in the most environmentally sound and economically viable manner possible, in concert with the geology (and chemistry) of the desired excavation sequence.

Walls and floors of quarries excavated below the water table will only yield groundwater as fast as the host rock permits. Accurate mapping of joints (see Figure 28) and other discontinuities, in relation to the position of recharge areas (and the mine impoundment), is of critical importance, as are the hydraulic conductivities of the excavated units. Hydraulic conductivities can be estimated via pump tests, slug tests, packer tests and flow net construction, combined with proper geologic interpretation.

STOP 14. Blasting Overview. We will stop at a drilled shot if time allows to describe drilling and blasting procedures, current best management practices, view videos of recent blast and discuss and review vibration standards and results.



Figure 28. Photo of water-producing discontinuity in quarry wall. Flow is estimated at approximately 200 gpm.

ACKNOWLEDGEMENTS

The authors would like to thank the quarry owners and operators, Holcim US, Peckham Materials, A. Colarusso and Son and O & G for allowing us to use their property for this trip. Access to these sites is stringently regulated by the Mine Health and Safety Administration and is restricted without the express permission of the quarry owners and operators. Due to their graciousness, the authors are able to lead a large number of educational tours to these quarries and we would like to remain in their good graces. Please do not attempt to visit these quarries without proper authorization.

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Much of the information presented herein is based on geologic and hydrogeologic investigations performed by the authors as consulting geologists to the mining industry over the last 25 years. In addition, the following publications were relied upon to provide general information used in this presentation.

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The Classic Devonian of the Catskill Front: A Foreland Basin Record of Acadian Orogenesis

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INTRODUCTION

Foreland basins sedimentary rocks preserve a record of the evolution of a mountain belt. Indeed, adjacent to ancient, deeply eroded orogens like the Devonian Acadian mountain belt, the foreland basin sedimentary succession is a key source of data on the timing and character of orogenic events and processes long since eroded away.

Making viable interpretations of orogenesis from foreland basin sedimentary rocks is, of course, not simple and straightforward. For example, look at the evolving understanding of Devonian volcanic ash beds (“tephras”, “tuffs”, “K-bentonites”) and their implications for explosive Acadian volcanism, one of the author’s primary research interests (Ver Straeten, 2004a, 2008; Ver Straeten et al., 2005).

Prior to the mid-1970s, only a few thin volcanic tephra were known from Devonian strata across eastern North America. Now approximately 100 beds are known from the Devonian of the Appalachian basin. Older perspectives interpreted each bed to represent a single explosive volcanic eruption. Microstratigraphic and geochemical study of the beds, however, indicate that many are multi-event layers, with a history of amalgamation of multiple eruptions, resedimentation on the sea floor, and/or mixing with detrital sediments. Furthermore, it is unknown how many tephra layers, erupted over ca. 60 million years, were not preserved in the sedimentary rock record. The question then arises, does the existing record of foreland basin tephra yield a viable proxy of Acadian volcanism through time, despite biased and incomplete preservation of individual eruptions?

This field trip and paper examine the Devonian (+/- Upper Silurian) sedimentary rock succession in the Catskill Front of eastern New York, with additional perspectives from across New York and the Appalachian basin. Through a review of the existing literature, and presentation of new data, the author hope to produce a detailed, yet broad perspective of the implications of data from the foreland basin on the Acadian orogeny in the northeastern U.S. It is, of course, a progress report on the subject, a single author’s perspective on what is known at this time, and resulting interpretations. Much more and varied work is needed.

This paper and field trip will present an overview of the Acadian orogeny and its foreland basin. It will then focus on a set of sedimentary rock characters that potentially provide regional perspectives on Acadian orogenesis. These include: 1) sedimentation and sea level history; 2) clastic rock composition; 3) volcanic tephra beds; and 4) soft sediment deformation. These four types of information may, and generally indeed do, permit interpretation of the developmental history of the Acadian orogen and the evolution of the Acadian foreland basin.

The Acadian Orogeny. As defined by some workers, the Acadian Orogeny was a major Late Silurian to Devonian-age mountain-building event in eastern North America. There are other perspectives: van Staal et al. (in press) recognize two separate events, the Acadian and Neo-Acadian orogenies. For the present paper, the Acadian orogeny will be treated as a single event. Long thought to be a Devonian-only event, recent dating of igneous rocks in Maine and adjacent areas indicate a Late Silurian beginning (Bradley et al., 2000). The orogeny is generally interpreted to have been the result of oblique collision of eastern North America (Laurentia) with one or more landmasses (e.g., Avalon, Rast and Skehan, 1993; Avalon and Meguma terranes, van Staal et al., in press), or a series of terranes. It should be noted, however, that Murphy and Keppie (2005) have recently proposed that the orogen may have, at least in the northern Appalachians, resulted from Laramide-type uplift due to flattening of a subduction zone along an Andean-type continental margin.

Acadian orogenesis resulted in construction of an elongate mountain chain that extended from Newfoundland to Alabama. It is characterized by significant plutonic/volcanic activity, regional metamorphism, and large scale deformation along the orogen. Post-orogenic processes and events have destroyed or buried much of the rock record of the event, especially in the central to southern Appalachians.

The Acadian Orogeny is generally interpreted to have resulted in a general southward-progressive, oblique collision against the eastern Laurentian margin, which extended from the Mid-to-Late Silurian to Late Devonian or Early Mississippian (Rogers, 1967; Etensohn, 1985, 1987). Southwestward-migrating pulses of tectonism during the orogeny, indicated by the distribution of clastic wedges along the orogen, were projected by Etensohn (1985, 1987) to be associated with collision at successive Laurentian promontories (Gaspé Peninsula, New York, Virginia, and Alabama Promontories, northeast to southwest, respectively; Thomas, 1977) with a probable single Avalon plate. Other authors proposed collision of separate Avalonian blocks during the Acadian Orogeny (e.g., Rast and Skehan, 1993). Additional select references on the Acadian orogeny include Bradley (1983), Osberg et al. (1989), papers in Roy and Skehan (1993), Rast and Skehan (1993), Robinson et al. (1998), Tucker et al. (2001), and Bradley and Tucker, 2002).

Bradley et al. (2000) provided a detailed history of cratonward advancement of the Acadian orogenic front across Maine and adjacent areas of New England and Canada. Their results, a synthesis of geochronologic, sedimentologic/stratigraphic, biostratigraphic, igneous, and structural data, indicate that between the Late Silurian and the earliest Late Devonian (ca. 40 m.y.), the Acadian deformation front migrated over 240 km cratonward (non-palinspastic distance) from coastal Maine into Quebec. A similar developmental history likely characterized the orogeny across central to southern New England, impacting foreland basin evolution in the New York region.

Foreland Basins. Foreland basins are elongate troughs, or “moats”, that form on continental crust between orogenic belts and adjacent cratons (Dickinson 1974; Miall, 1995; DeCelles and Giles, 1996). The geometry of a basin is the result of one or more interacting factors, including orogenic loading, +/- subduction-related geodynamics, mantle processes, in plane stresses, and redistribution of the load via sedimentation (e.g., Beaumont, 1981; Jordan, 1981; Beaumont et al., 1988; Mitrovica et al., 1989; Cloetingh, 1988; Jordan and Flemings, 1991). The interaction of multiple processes with plates of varying flexural rigidity result in a complex history of foreland basin flexure (DeCelles and Giles, 1996; Gurnis, 1992).

Foreland basin systems have been described by DeCelles and Giles (1996) as consisting of four distinct depozones. These comprise the wedge-top, foredeep, forebulge, and back-bulge basin, from the orogenic front to the craton, respectively (DeCelles and Giles, 1996). These zones, summarized below from DeCelles and Giles (1996), may shift laterally through time.

The wedge-top is the most proximal zone of sediment deposition, comprising the front of the orogenic fold-thrust belt (DeCelles and Giles, 1996). It is characterized by deposition of coarse sediments which thicken toward the foredeep; it is commonly deformed, with numerous unconformities.

The foredeep is a subsiding trough adjacent to the proximal wedge-top, characterized by a thick sedimentary succession that thins cratonward. This area is the focus of many foreland basin studies. It typically is approximately 100 – 300 km wide, with a cumulative sedimentary fill 2-8 km thick. The sedimentary fill is sometimes distinguished as earlier or more distal deeper water “flysch” and more proximal shallow marine to terrestrial “molasse” sediments. The bulk of the sediment is derived from orogenic belt.

The forebulge is a zone of possible flexural uplift on the cratonward side of the foredeep. The forebulge may be on the order of 60-470 km in width, and a few tens to hundreds of meters high. It commonly is an area of little to no deposition and/or erosion, characterized by unconformities; if flooded, carbonate platforms may develop over the bulge. Forebulges generally migrate laterally through time, associated with changes in flexural kinematics in a foreland basin system.

The back-bulge basin of foreland basin systems is a zone of sediment accumulation between the forebulge and craton, with a relatively low subsidence rate. Sediment may be derived from the orogenic belt, but also from the

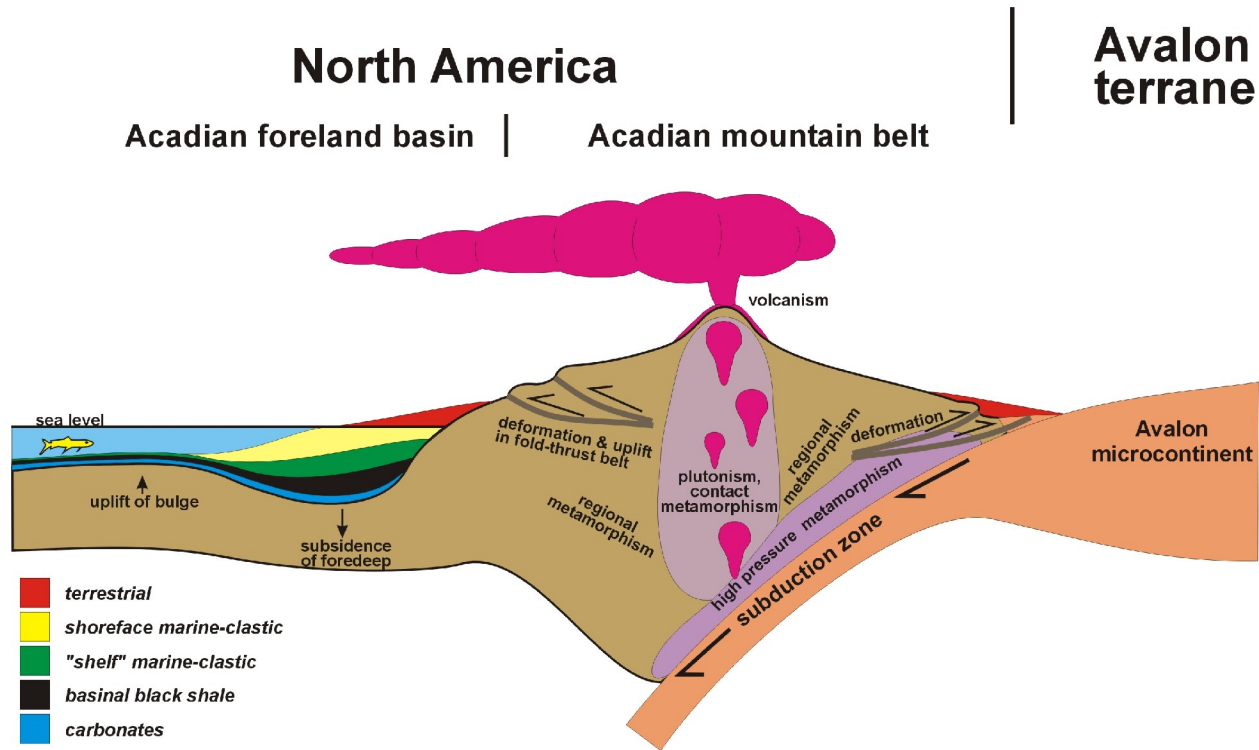


Figure 1. Idealized cross-section of the Acadian orogen and foreland basin.

craton, the forebulge and in situ carbonate production, if it is flooded by marine waters. It is characterized by a relatively thin, tabular-layered sedimentary succession.

The Acadian Foreland Basin and the Appalachian Basin. The Acadian foreland basin was an elongate trough, formed during the Late Silurian to Devonian, due to load-induced subsidence adjacent to the rising Acadian orogen (Figure 1). The basin was oriented approximately northeast to southwest during the Devonian, and extended from Newfoundland to northern Georgia and Alabama. The Appalachian basin portion of the foreland is variously interpreted to have been positioned between approximately 25-40 degrees south of the equator during the Devonian (van der Voo, 1983; Witzke and Heckel, 1988; Scotese and McKerrow, 1990). Additional select references on the Acadian foreland basin include Rogers (1967), papers in Woodrow and Sevon (1985), Ettensohn (1985), Woodrow (1985), Rust et al. (1987, 1989), Lawrence and Rust (1988), Osberg et al. (1989), and Bradley et al. (2000).

The Acadian foreland is generally interpreted to have been a collisional-related foreland basin (Miall, 1995). Ongoing debate about the direction of subduction (e.g., Bradley, 2000; versus van Staal, 2007; van Staal et al., in press) does not permit interpretation of whether the Acadian foreland was a retroarc or peripheral foreland basin.

An important perspective sometimes overlooked in discussions pertains to the issue of the Acadian versus the Appalachian basin. The “Appalachian basin” is only a part of the greater Acadian foreland basin. The former represents a body of rock preserved through the states of New York, New Jersey, Pennsylvania, Maryland, Virginia, West Virginia, Tennessee, Ohio, and parts of southern Ontario. The greater Acadian foreland basin system occurred from Newfoundland to Alabama, and in our region, extended into New England. Today, the proximal portion of Acadian foreland basin system has been removed by uplift and synorogenic to post-orogenic weathering and erosion.

Previous Work. Since the mid-1800s, the Upper Silurian to Devonian succession in the Catskill Front has been the focus innumerable studies. The succession overlying the Taconic unconformity in the area comprises nearly 2.7 km of sedimentary rocks of Uppermost Silurian through Late Devonian age (Pridoli through Frasnian stages; Figures 2-6). The bulk of the rocks are clastic (mudrocks, sandstones, and minor conglomerates); three packages of carbonates

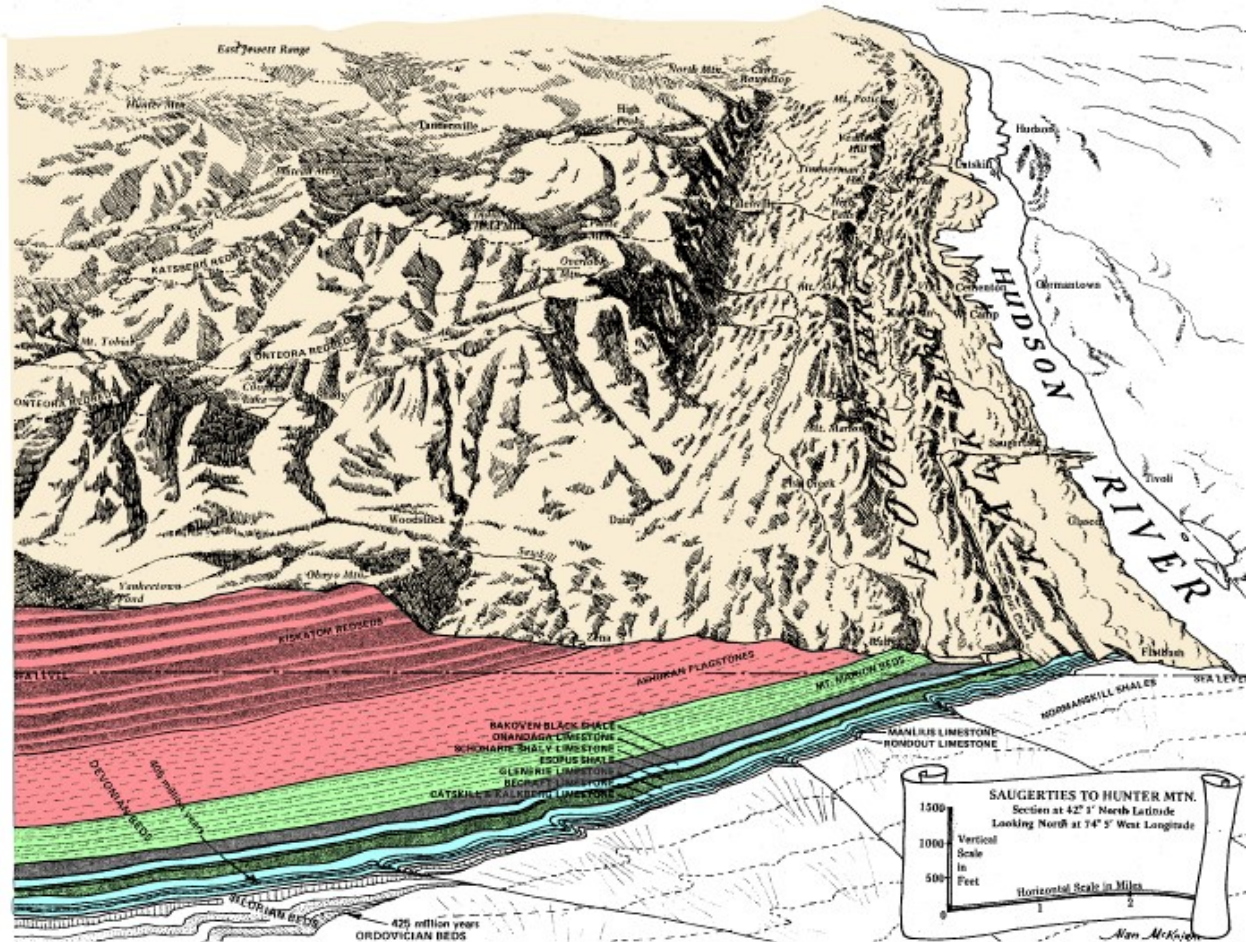


Figure 2. Cross section of the Catskill front. Physiographic map and cross-section of the Hudson Valley and Catskill Front, showing bedrock and relation to topography. Stratigraphic terminology on figure after Chadwick (1944); in modern terminology, Catskill Shaly=New Scotland; Kiskatom=Plattekill and Manorkill; Onteora=Oneonta; and Katsberg=Walton formations. Onandaga=Onondaga. Note that the Mount Marion Formation near Kingston includes much of the lower part of strata shown as Ashokan Formation. Original drawing by Alan McKnight.

occur low in the succession (Helderberg Group, Glenerie Formation, and Onondaga Formation). They were deposited over a period of approximately 45 million years (Tucker et al., 1998; Kaufmann, 2006; Ogg et al., 2008). The strata are divided into 15 formations, most of which consist of relatively thin but distinctive units in the lower part of the succession (Table 1).

The rocks represent a broad range of depositional environments from supratidal to deep ramp marine carbonate facies, and clastics of marine to terrestrial origin, from basinal black shales to shoreface sandstones to fluvial-dominated channel and floodplain facies.

Overall, the succession records initial marine carbonates (Rondout Formation and Helderberg Group), a hiatus of a few million years (Wallbridge unconformity of Sloss, 1962), thin chert-rich carbonates overlain by a first, relatively thin wedge of clastics, and a gradational return to carbonate deposition (Glenerie, Esopus, Schoharie and Onondaga formations; total thickness of lower strata = ca. 275 m). The succeeding second wedge of clastics, approximately 2.4 km thick, consists of a lower package of marine strata (>500 m thick) and overlying terrestrial strata, which extend

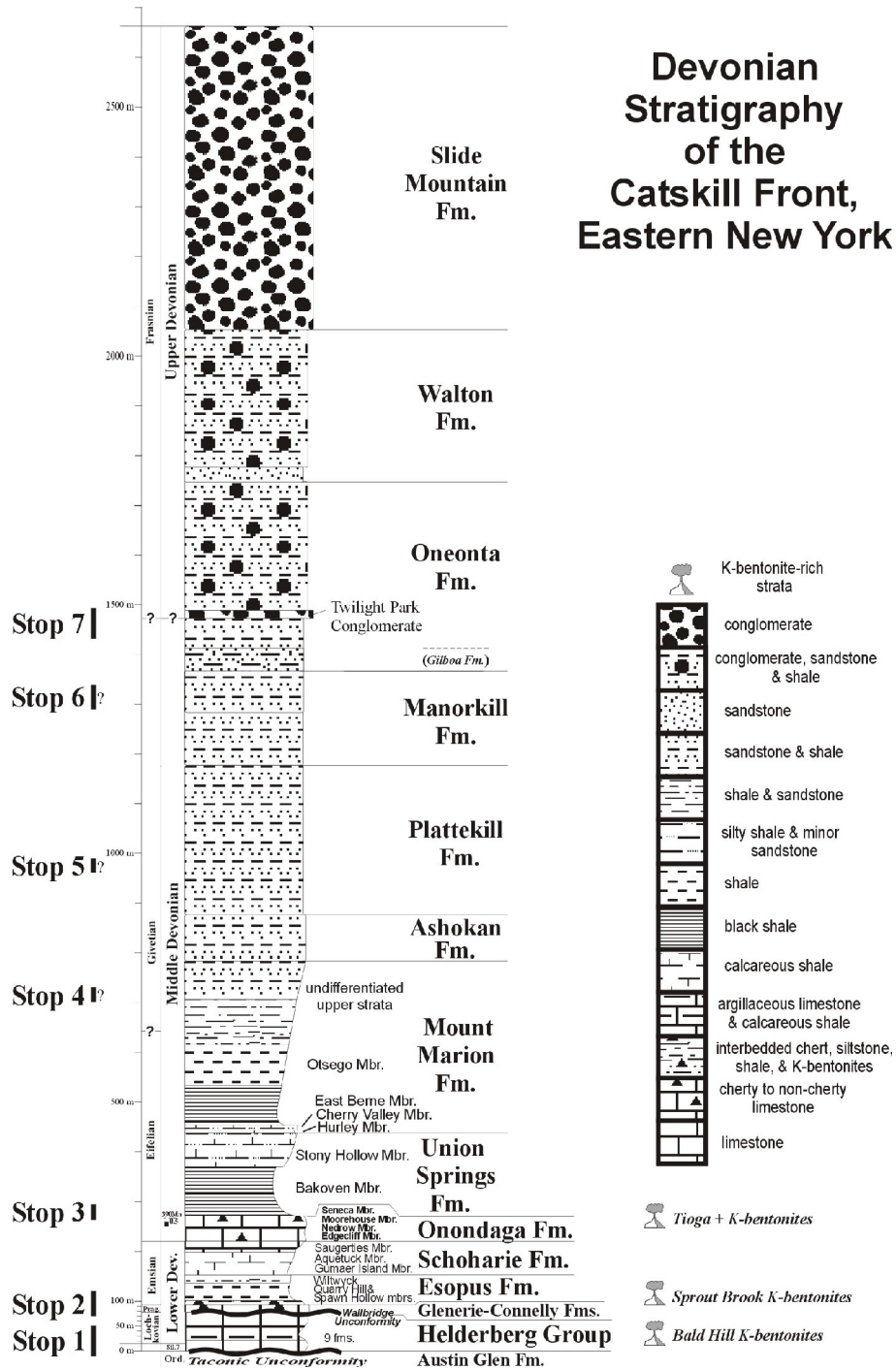


Figure 3. Stratigraphy of the Catskill Front. Composite section of >2.6 km of uppermost Silurian and Lower to Upper Devonian strata, from the Taconic unconformity (Catskill) to the top of Slide Mountain. Stratigraphic positions of Stops 1-8 shown on left. Data from Rickard (1962), Hodgson (1970), Rehmer (1976), the author's field notes, Feldman (1985), the author's field notes, Chadwick (1944), and Fletcher (1967), from low to high, respectively. Unidentified formations overlying the Taconic unconformity include, from low to high: Upper Silurian Rondout Formation; Helderberg Group strata, including possibly Upper Silurian-age Manlius Formation, and Lower Devonian Coeymans, Kalkberg, New Scotland, Becraft, Alsen and Port Ewen formations.

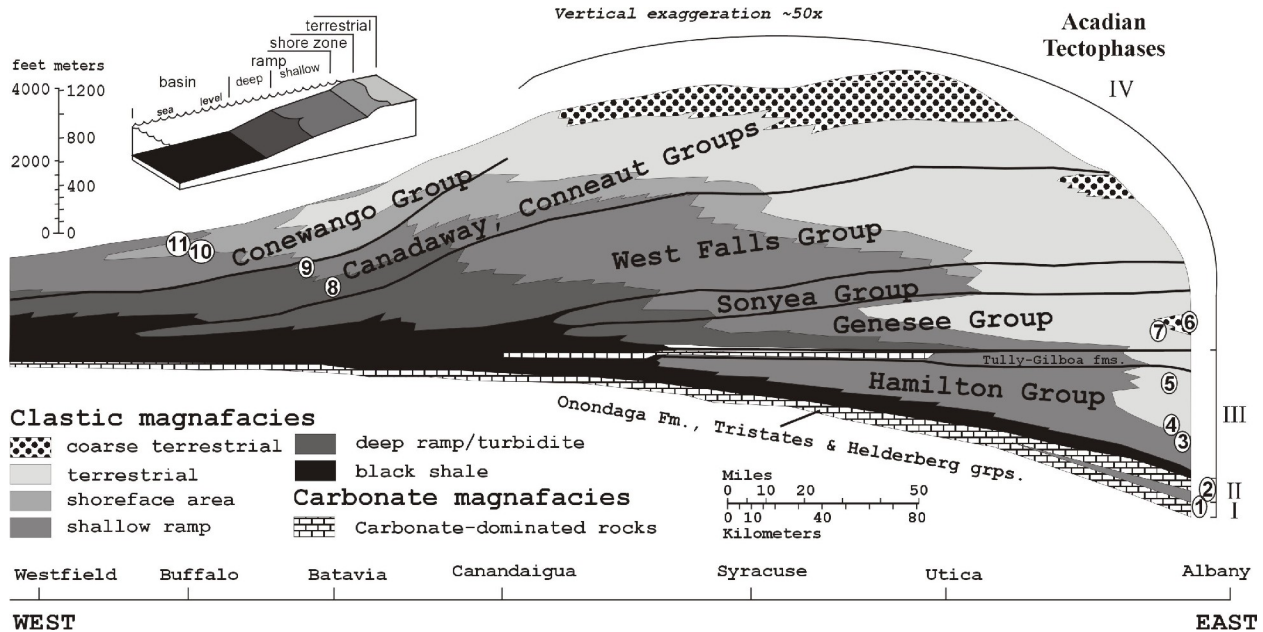


Figure 4. Cross-section of the Catskill delta complex. Figure depicts preserved Devonian strata along the New York-Pennsylvania border. Transect is roughly perpendicular to strike of the basin and the Acadian orogenic belt. Vertical exaggeration ~50x. Modified after Rogers et al. (2000). Cities across bottom of figure occur longitudinally at or north of the transect. Circled numbers 1-11 denote position of conglomerate samples in this study. Roman numerals I-IV denote strata associated with Etensohn's (1985) Acadian tectophases.

to the top of Slide Mountain, the highest peak in the Catskills. Additional information on the Silurian-Devonian strata in the Catskill Front, along with key references, is provided in Table 1.

RESULTS

Sedimentation and Sea Level History

Major Sedimentation Trends. The Devonian succession in New York can essentially be viewed as three major alternations of carbonates (+/-quartz arenites) and terrigenous clastic sedimentary rocks. The carbonates and quartz arenites comprise sediments derived from within the sedimentary basin. The terrigenous clastics, alternatively, were transported into the basin from external sources. These very basic concepts are at the core of this paper and field trip.

Initial Upper Silurian to Lower Devonian carbonates (Rondout Formation and Helderberg Group) are overlain by a major (Wallbridge) unconformity. A succeeding quartz arenite-carbonate suite (Oriskany-Glenerie-Connelly formations) is followed by a first major, if relatively thin, package of synorogenic clastics (Esopus Formation). A gradation from mixed clastic-carbonate to fully carbonate deposition (Schoharie to Onondaga formations) is, in turn, succeeded by a second package of clastic rocks (Hamilton Group).

In central New York, Hamilton clastics are overlain by a third carbonate unit (Tully Formation), succeeded once more by a third package of clastic rocks (Genesee Formation and succeeding strata), which continue into the lower Mississippian beyond the borders of New York. Note that all of the previous mentioned strata were deposited in marine settings.

In eastern New York, lower Hamilton marine rocks grade upward into terrestrial strata, and the third marine carbonate unit is not developed. Terrestrial facies (Plattekill to Slide Mountain formations) characterize the rest of the Devonian succession in the Catskills.

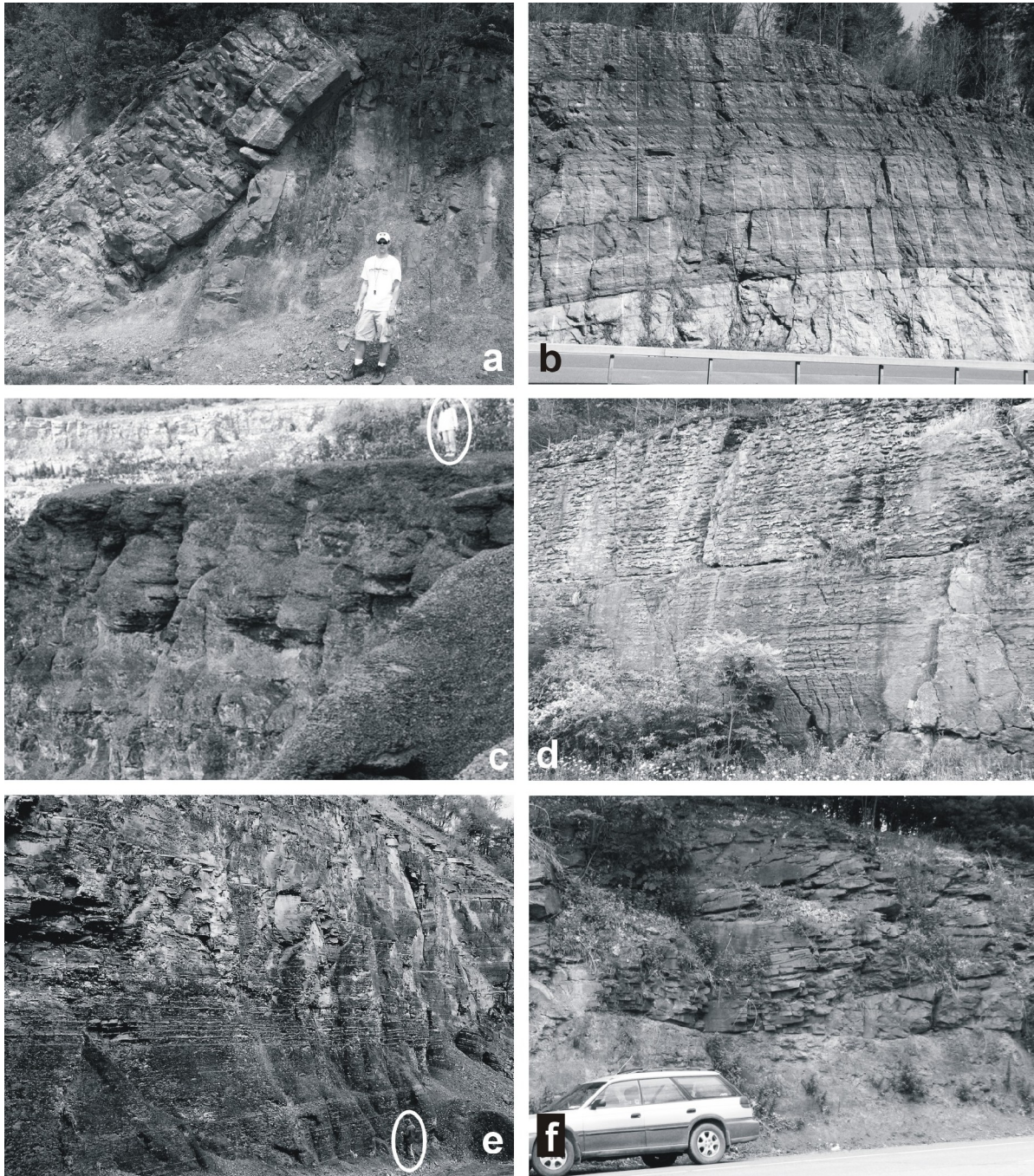


Figure 5. Strata of the Catskill front I. Photo figure of select strata, Catskill front: a) Taconic unconformity, Rte. 23, Catskill. Ordovician Austin Glen Formation below angular unconformity, uppermost Silurian Rondout Formation above; b) upper Helderberg Group limestones, Rte. 199, Kingston. Lower Devonian Becraft, Alsen and Port Ewen formations, from low to high; c) middle of Lower Devonian Esopus Shale, near Hudson, Columbia County. Quarry Hill Member; person (circled) for scale; d) chert-rich Middle Devonian Edgecliff Member, Onondaga Limestone, Rte 23 and Kings Highway SW of Catskill. Note multiple layers of chert. Exposed strata are ~11 m thick; e) shales and thin sandstones, upper East Berne Member, Middle Devonian Mount Marion Formation, near Quarryville, Ulster Co. Person (circled) for scale; f) Middle Devonian Ashokan Formation, along Rte. 28, north-west of Kingston. Floodplain paleosols (below) and channel sandstones (above).

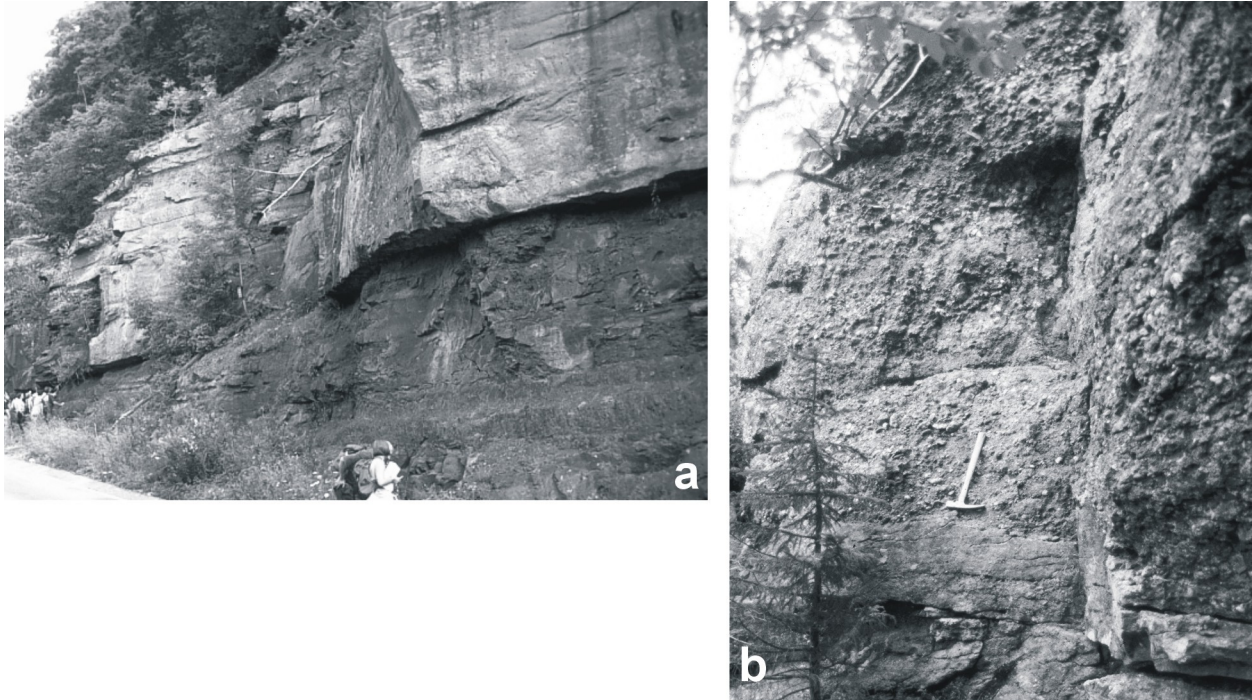


Figure 6. Strata of the Catskill front II. Photo figure of select strata, Catskill front: a) upper Manorkill (Middle Devonian) or lower Oneonta Formation (Upper Devonian?), Rte. 23, Kaaterskill Clove, Greene Co. Red floodplain paleosols (below) and gray channel sandstones (above); b) Upper Devonian (?) Twilight Park Conglomerate, Oneonta Formation, at Sunset Rock, North-South Lakes State Campground (north of North Lake). Conglomerate over pebbly sandstone. Large rock hammer for scale.

Post-Wallbridge Lower to Middle Devonian carbonate (+/- quartz arenite) suites are characterized by three key patterns (Figure 7): 1) a relatively widespread distribution; 2) a relatively tabular thicknesses, and: 3) relatively shallow water litho- and biofacies. This is most clearly shown by the Onondaga Limestone east-west across the state (Figure 7). Similar patterns are seen in the Tully Limestone in central New York; they break down laterally, however, as the Tully passes into thicker, proximal sandstones to the east, and disappears beneath an unconformity to the west. The shallow water Oriskany Sandstone and correlatives do not appear to be widespread or tabular in distribution across the state. However, local pockets of Oriskany across central to western New York indicate that it once was deposited across the area, but later eroded beneath a pre-Onondaga unconformity.

In contrast, those same three characters show opposing patterns in the clastic rocks immediately above the carbonate-rich packages (Figure 7). These initial clastics (Esopus, Union Springs and Geneseo formations) are characterized by: 1) relatively restricted distribution (toward the orogen); 2) distinctive wedge-shaped thicknesses (thickening toward the orogen), and; 3) relatively deep water, dark gray to black shale facies and depauperate dysoxic to anoxic biofacies.

These three key characters undergo sharp, abrupt changes across the contact of the carbonate to initial clastics units. Strata above the initial clastics, however, display an overall, gradational shift to increased CaCO₃/decreased detrital content, a more widespread distribution, a more tabular geometry, and a trend toward shallower water litho- and biofacies (Figure 7).

Sea Level History. Studies of the Devonian succession in New York and, in some cases, across the Appalachian basin have outlined a history of relative sea level cycles, or, in sequence stratigraphic terms, third order depositional sequences (e.g., Johnson et al., 1985; Dorobek and Read, 1986; House and Kirchgasser, 1993; Brett and Baird, 1996; Filer, 2002; Ver Straeten, 2007a; see Figure 8).

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Internati. Stage (Series)	Formation	Lithology	Fauna/flora	Thickness	Environments	Key refs.
Frasnian (Up Dev)	Slide Mountain	ss, cgl; minor red/green sh/mudst	fish, terr. plants,	610 m ³	terrestrial; fluvial channels & floodplains	Burtner 1963, 1964; Fletcher 1964, 1967; Gale 1985
Frasnian (Up Dev)	Walton	ss, red/green/dk gray sh/mudst, minor cgl	fish, terr. plants,	305 m ³	terrestrial; fluvial channels & floodplains, minor lacustrine?	Burtner 1963, 1964; Fletcher 1964, 1967; Gale 1985; Gordon 1986
Frasnian (Up Dev)	Oneonta	ss, minor cgl, red/green/dk gray sh/mudst	fish, terr. plants, arthropods	274 m ³	terrestrial; fluvial channels & floodplains, minor lacustrine?	Burtner 1963, 1964; Fletcher 1964, 1967; Lucier 1966; Johnson 1968; Johnson & Friedman 1969; Nickelsen 1983; Gale 1985; Bridge & Gordon 1985; Bridge & Nickelsen 1985; Gordon 1986
Givetian (Mid Dev)	Manorkill	ss, red/green/dk gray sh/mudst	fish, terr. plants, arthropods	189 m ³	terrestrial; fluvial channels & floodplains, minor lacustrine	Burtner 1963, 1964; Fletcher 1964, 1967; Lucier 1966; Way 1972; Gale 1985; Willis 1986; Willis and Bridge 1988; Bridge & Willis 1994
Givetian (Mid Dev)	Plattekill	ss, red/green/dk gray sh/mudst	fish, terr. plants	305 m ³	terrestrial; fluvial channels & floodplains, minor lacustrine	Burtner 1963, 1964; Fletcher 1964, 1967; Lucier 1966; Way 1972; Gale 1985; Willis 1986; Willis and Bridge 1988; Bridge & Willis 1994
Givetian (Mid Dev)	Ashokan	dk gray/green sh/mudst, ss	fish, terr. plants	152 m (Kingston), 91 m (Kiskatom) ⁴	largely terrestrial (coastal lowlands); fluvial channels & floodplains, lacustrine & palustrine	Burtner 1963, 1964; Wolff 1967, 1969
Eifelian-Givetian (Mid Dev)	Mount Marion	Black/gray sh/mudst, ss, minor cgl	normal marine	~425 m ⁵ (Kingston; thins to north)	marine: clastic anoxic basin to shoreface; local terrestrial tongue	Wolff 1967, 1969; Griffing & Ver Straeten 1991; Ver Straeten 1994, 2007a; Ver Straeten & Brett 1995
Eifelian (Mid Dev)	Union Springs	black sh, buff calc sh to fine ss	restricted to normal marine	~175 m ⁵ (Kingston; thins to north)	marine: anoxic to dysoxic clastic basin	Griffing & Ver Straeten 1991; Ver Straeten et al., 1994; Ver Straeten & Brett 1995; Ver Straeten 2007a
Eifelian (Mid Dev)	Onondaga	ls (wacke- to grainstones)	normal marine	ca. 50 m ⁶	marine: shallow carbonate "shelf" ¹⁰	Oliver 1956; Ver Straeten & Brett 1995; Ver Straeten 2007a
Emsian (Low Dev)	Schoharie	ls, sh, chert, some clastic sh	normal marine	26.7 m ⁷	marine: mixed carb-clastic "shelf"	Johnsen 1957; Ver Straeten & Brett 1995; Ver Straeten 2007a
Emsian (Low Dev)	Esopus	sh/mudst, siltst, ss		⁸	marine: clastic basin to "shelf"	Rehmer 1976; Ver Straeten & Brett 1995; Ver Straeten 2007a

Table 1. (continues) Upper Silurian & Devonian Strata of the Catskill Front.

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Internatl. Stage (Series)	Formation	Lithology	Fauna/flora	Thickness	Environments	Key refs.
Pragian (Low Dev)	Glenerie ² Oriskany Connelly	ls & chert quartz SS quartz cgl	normal marine	4.6 m ⁹ (Glenerie)	marine: shoreface to shallow shelf	Hodgson 1970; Ver Straeten & Brett 1995; Ver Straeten 2007a
Pragian (Low Dev)	Port Ewen	ls (mudstones to packstones), some clastic sh	normal marine	2.1 m ⁹ (thickens to south)	marine; deeper carbonate “shelf”	Rickard 1962; Mazzo & LaFleur 1984; Ebert 1984, 1987
Lochkovian (Low Dev)	Alsen	ls (wacke- to packstones), minor clastic sh	normal marine	9.8 m ⁹	marine; shallow carbonate “shelf”	Rickard 1962; Ebert 1984, 1987
Lochkovian (Low Dev)	Becraft	ls (grainstones)	normal marine	14.6 m ⁹	marine; carbonate shoal	Rickard 1962; Ebert 1984, 1987
Lochkovian (Low Dev)	New Scotland	lIs (mudstones to packstones), some clastic sh	normal marine	30 m ⁹	marine; deeper carbonate “shelf”	Rickard 1962
Lochkovian (Low Dev)	Kalkberg	ls (wacke- to packstones), minor clastic sh	normal marine	16.5 m ⁹	marine; shallow carbonate “shelf”	Rickard 1962
Lochkovian (Low Dev)	Coey- mans	ls (grainstones)	normal marine	4.3 m ⁹	marine; carbonate shoal	Rickard 1962; Ebert and Matteson 2003
Pridoli? ¹ (Up Sil?)	Manlius	ls (mudstones to wackestones), minor clastic sh	restricted marine	15.5 m ⁹	marine; carbonate tidal	Rickard 1962; Ebert and Matteson 2003
Pridoli? ¹ (Up Sil?)	Rondout (upper part)	dolst, basal ss	?	1.2 m ⁹	coastal margin; carbonate, minor clastic supratidal	Rickard 1962;

Table 1. (continued) Upper Silurian & Devonian Strata of the Catskill Front. *For all strata in the Catskill Front, see also Chadwick (1944).*

¹ Previously assigned to the Devonian (e.g., Rickard 1975), recent work by Ebert (e.g., Ebert and Matteson, 2003) suggests a Silurian age.

² North of Catskill, the Glenerie Ls. undergoes facies change into the Oriskany Sandstone; south of Kingston, the Glenerie Ls. is underlain by the Connelly Cgl., which replaces it completely in the Skunnemunk outlies south of Kingston.

³ Thicknesses from Fletcher (1967)

⁴ Thicknesses at type area (near Kingston) and Catskill Front (near Kiskatom), from Chadwick (1944)

⁵ Estimated thicknesses at Kingston, from Rickard (1989)

⁶ Thickness at Saugerties (NYS Thruway cuts), from Feldman (1985)

⁷ Thickness at Rte. 23 (W of Catskill), from Ver Straeten and Brett (1995)

⁸ Thickness from Rte 23a (SW of Catskill), from Rehmer (1976)

⁹ Thicknesses from Austin Glen (W of Catskill), from Rickard (1962)

¹⁰ Strata not deposited on actual continental “shelf”, but in foreland basin/epicontinental sea setting, with a shallow to deep ramp-like basin morphology.

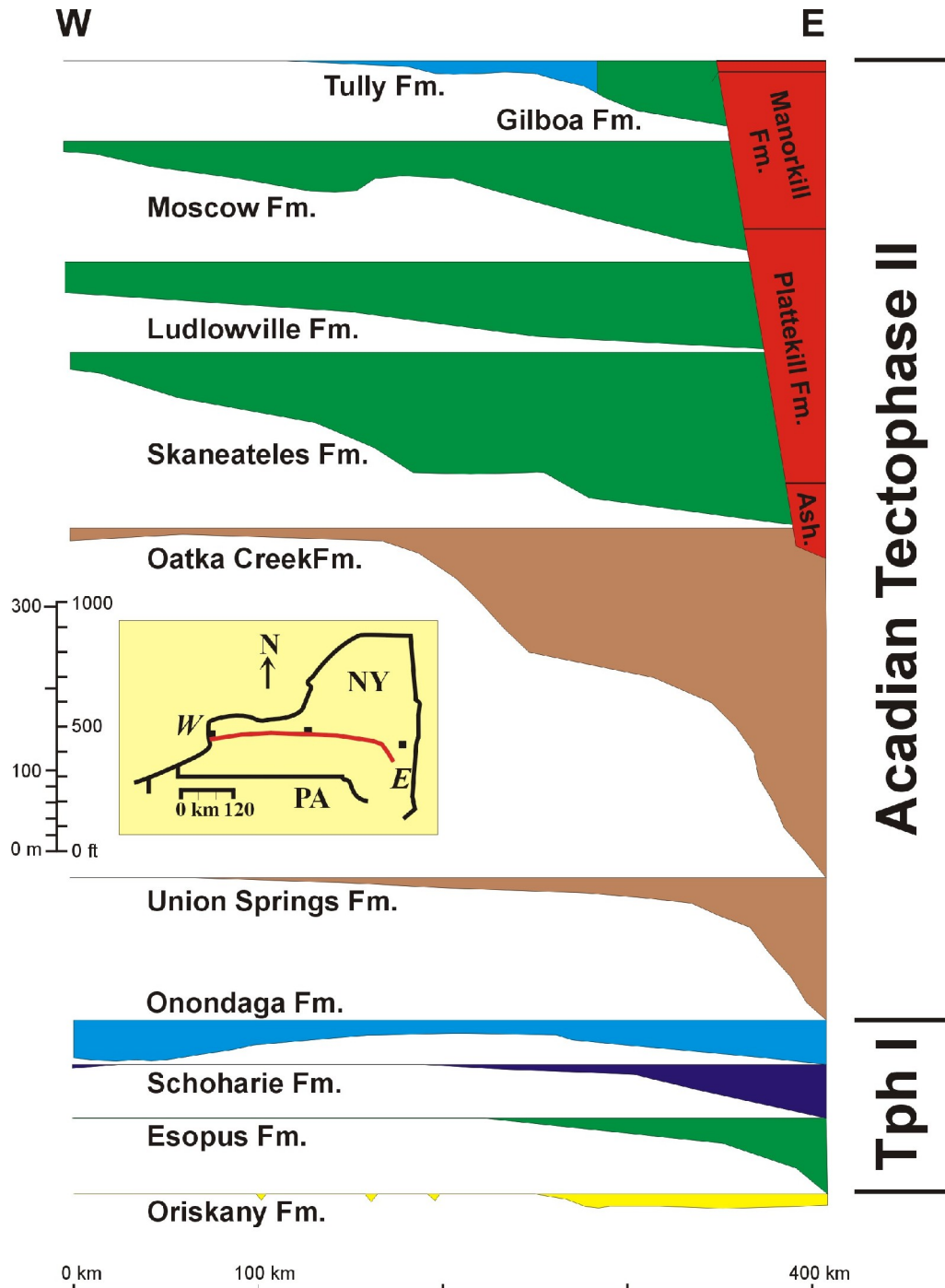


Figure 7. Stratal geometry of upper Lower and Middle Devonian units, New York. Geometry and generalized facies of strata along the New York outcrop belt between Catskill front (east) and Buffalo (west). Note relatively tabular geometry of the Onondaga and Tully formations, in contrast with the Esopus, Union Springs and Oatka Creek-Mount Marion formations. Modified from Ver Straeten and Brett (1995), with additional data from Rickard (1989; Oatka Creek to Moscow fms.), Heckel (1973; Tully Fm.) and Fletcher (1967; Ashokan to Manorkill Fms.). Correlations between the marine and terrestrial units remain tentative to unknown.

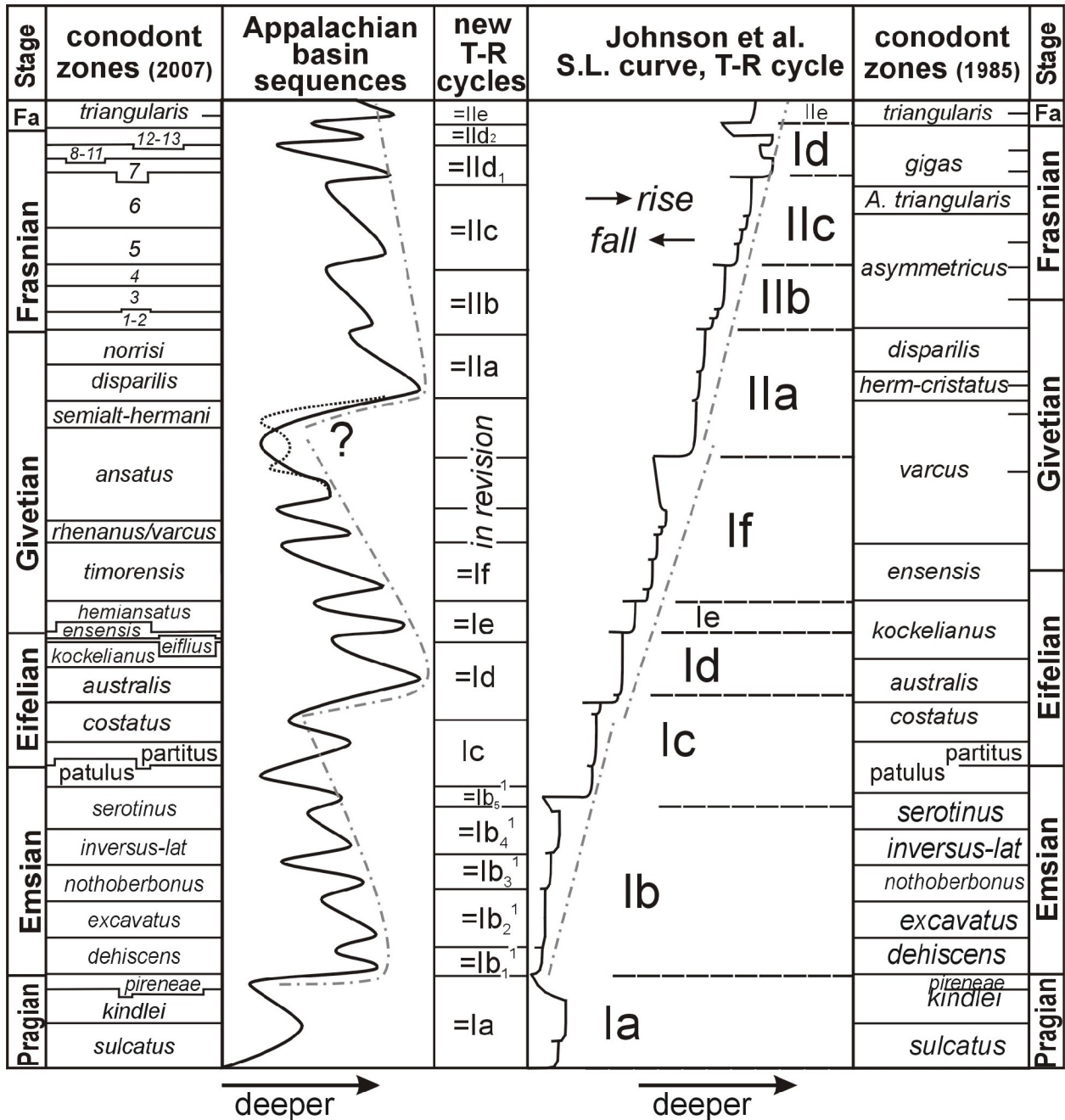


Figure 8. Devonian depositional sequences/sea level cycles. Comparison of Pragian through lower Famennian relative sea level curve (Appalachian basin) and Euramerican eustatic sea level curve (of Johnson et al., 1985). Appalachian curve/sequences include refinements on original Euramerican T-R cycles. Note: 1) correlation of sea level cycles on both sides of the diagram (indicating strong eustatic control on Appalachian basin third order sequences); 2) overall sea level rise from upper Pragian to upper Frasnian for Johnson et al. (1985) curve, and contrasting Appalachian basin pattern of three cycles of major transgressions in the lower Emsian, mid Eifelian, and upper Givetian, followed by overall regressions. These cycles represent Ettensohn's (1985) Tectophases I, II and III (part). Divergence of sea level curves at tectophase-scale cycles is indicative of Acadian influence on relative sea level. Data for Appalachian basin cycles from Johnson et al. (1985), Dorobek and Read (1986), House and Kirchgasser (1993), Brett and Baird (1996), Filer (2002), Ver Straeten (2007a, in press).

In the area of this field trip, Uppermost Silurian to Lower Devonian strata of the Helderberg Group, Oriskany Sandstone and Esopus and Schoharie formations comprise 8 major sequences; two in the Helderberg Group (Rondout to basal Becraft fms.; lower Becraft to Port Ewen fms.), one absent locally, but developed in more basinal settings of the Oriskany Sandstone across the basin; and five in the Esopus and Schoharie formations (Ver Straeten, 2007a). The latter five represent a new interpretation by the author, based on basinwide correlation of Emsian-age Esopus and Schoharie strata, initial field study in Nevada, and comparison with data from Europe and North Africa, along with recent geochronologic age data (Tucker et al., 1998; Kaufmann, 2006) that project a duration of 15 to 17 million years for the Emsian stage.

Seven sequences subdivide the Middle Devonian record across the Appalachian basin (Brett and Baird, 1996; Ver Straeten, 2007a). The lower one occurs within the lower to middle Onondaga Limestone; upper Onondaga strata and the overlying Union Springs Formation (lower part of the Marcellus “shale”) comprise a second sequence. Younger cycles are represented by each of four overlying Hamilton Group formations, and the Tully to Genesee formations. In the field area, only the lower three are identified. As Hamilton strata pass into terrestrial facies above the Mount Marion Formation, the additional Middle Devonian sequences have not been delineated (Ashokan through Manorkill formations). Similarly, multiple Upper Devonian stratigraphic sequences have not been delineated in the terrestrial facies of the Oneonta through Slide Mountain formations.

Johnson et al. (1985) outlined a Euramerican sea level history for the mid Lower to Upper Devonian strata (Pragian to Famennian stages; Figure 7, right side). Although modified and refined since its first publication by numerous workers globally, their interpretations still form the basic standard for Devonian sea level cyclicity.

Clastic Rock Petrology

Previous Results. Between the 1960s and 1980s, a number of petrographic studies examined the mineralogy of New York’s Lower to Upper Devonian clastics. These focused predominantly on the terrestrial rocks of the Catskill Front, from the Plattekill Formation upward. Much of the data is found in unpublished theses. Most studies focused on thin section analyses of sandstones (e.g., Burtner, 1964; Lucier, 1966; Way, 1972; Ethridge, 1977; Gale, 1985), but also included analyses of the clay mineralogy of mudrocks (e.g., Friend, 1966; Wolff, 1967; Leibling and Schirp, 1976, 1980; Hosterman and Whitlow, 1983). Petrology of the conglomerates has received less attention; however, some compositional data is reported (e.g., Nickelsen, 1983; Smith and Jacobi, 1998).

Petrographic data from the Lower Devonian Esopus Formation, the initial package of clastic sedimentary rocks, have a relatively quartz-rich composition (Rehmer, 1976). Excluding fine clay matrix from the data, a single sandstone sample analyzed by Rehmer (1976) was dominated by quartz (77.8%), with only very minor amounts (3.0 to 0.6%) of coarse muscovite, along with plagioclase and K-feldspars, pyrite, chert, and calcite and dolomite. Trace amounts of glauconite, biotite, epidote and zircon were also present. Her analyses of Esopus siltstones yielded relatively similar results (e.g., quartz between 64.2 and 72.8%, excluding matrix). The quartz in the Esopus is dominantly clear, monocrystalline, and unstrained. Detrital phyllosilicates generally comprise >3% of the rock, including matrix; muscovite is more common than chlorite or biotite. Silt-sized shale or phyllitic rock fragments are found, but cannot be distinguished. However, rare micaceous fragments with crinkle-fold textures suggest input of at least some low-grade metamorphic rocks (Rehmer, 1976).

The most abundant grains in the Middle to Upper Devonian sandstones are monocrystalline and polycrystalline quartz, foliated metamorphic rock fragments, and sedimentary rock fragments; these comprise at least 80% of sandstones (Gale, 1985). Metamorphic rock materials include slate, phyllite and possible schistose fragments; sedimentary rocks grains found include shale, siltstone, sandstone and carbonate, along with chert. Other minor components, in decreasing concentration include plagioclase feldspars, with rare orthoclase feldspars and igneous (plutonic/volcanic) rock fragments (Gale, 1985). Chlorite is the chief micaceous mineral present, followed by muscovite and sericite, with only minor biotite (Lucier, 1966).

The matrix in the sandstones, previously interpreted to largely consist of clays, was shown by Gale (1985) to predominantly be comprised of deformed, ductile lithic fragments. Therefore, compositionally the sandstones are litharenites to sublitharenites, not graywackes.

Sandstone compositional trends through the succession (upper Mount Marion to Slide Mountain formations) show no major vertical changes. Smith's (1970) examination of lower Middle Devonian sandstones (upper Mount Marion through Plattekill formations) found some decrease in quartz content (monocrystalline + polycrystalline quartz + chert). This was accompanied by an increase in metamorphic rock fragments and matrix.

Gale (1985) noted some distinctive vertical trends through the overlying Plattekill to Slide Mountain formations. These included: 1) a general increase of mono- and polycrystalline quartz upward; 2) a net upward decrease in foliated metamorphic rock fragments; and 3) an upward increase in grain size. This expanded upon the similar results of Lucier (1966) for the Plattekill and Manorkill formations, who also noted overall vertical decreases in chert and plagioclase feldspar in those strata.

Combined data from Smith (1970) and Gale (1985) indicates an arc of initial decreasing, then increasing quartz content through the Middle to Upper Devonian succession. Chert concentration itself, however, decreases upward through the succession from the upper Mount Marion Formation.

Mudrock studies (Friend, 1966; Wolff, 1967; Rehmer, 1976; Leibling and Schirp, 1976; Hosterman and Whitlow, 1983) indicate that the clay mineral suite in Devonian fine-grained strata is largely comprised of illite and chlorite, with subordinate amounts of mixed-layer illite-smectite and kaolinite. This parallels clay mineral data from the sandstone studies. The chlorite is largely derived from low-grade metamorphic rocks (Leibling and Schirp, 1976, 1980). Hosterman and Whitlow (1983) examined Middle to Late Devonian black marine mudrocks (Marcellus to Cleveland shales) in New York and across the Appalachian basin. They noted an upward decrease in chlorite, accompanied by upward increase of mixed layer illite-smectite. Hosterman and Whitlow's (1983) mudrock analysis reported that kaolinite was largely restricted to younger Devonian shales (Rhinestreet Formation and younger); however, Gale found a decreasing kaolinite content above the Plattekill Formation.

Although Friend (1966) suggested that clays from the Manorkill Formation were largely unaltered, Hosterman and Whitlow's (1983) comprehensive study of the marine black shales stated that the existing illite (2M) and mixed layer illite-smectite clays are largely diagenetic in origin, derived from older illite (Md) and smectite. Interestingly, they also report that at least some of the chlorite was originally derived from smectite.

More recent studies, utilizing other analytical approaches, add additional perspectives to the composition and provenance of these Acadian-derived synorogenic clastics. New approaches include dating of detrital micas and zircons, and rutile geochemistry.

Aronson et al. (1994), using K-Ar methods, dated detrital white micas from Upper Devonian to Lower Mississippian clastics of the Catskill delta complex (NY and OH). The micas all yielded Devonian age dates between 406 – 380 Ma, with a single outlier of 419 Ma (with margins of error of +/- 7-20 m.y.). Micas from the Walton Formation in eastern New York produced an age of 401 +/- 7 Ma. They interpreted their results to indicate that rocks subjected to regional metamorphism or plutonism (>350° C) during the Acadian orogeny were the dominant source of sediments during deposition of the Catskill Delta.

McLennan et al. (2001) examined U-Pb ages of detrital zircons from a sample of the Frasnian-age lower Walton Formation (near Monticello, Sullivan Co., NY). A single sample of red sandstone yielded 41 zircons, from which they obtained a bimodal set of SHRIMP U-Pb ages of 1.23-1.00 Ga (Grenville age) and 470-420 Ma (which they termed "Taconian" in age; it actually encompasses the Middle Ordovician through most of the Silurian Periods, much of it post-dating the Taconian orogeny). They found no detrital zircons of unambiguous Acadian age in their sample.

Certain geochemical characteristics of rutile make it useful in provenance studies, including its ability to act as a geothermometer. Zack et al., (2004) examined the geochemistry of detrital rutiles from the same outcrop of the Upper Devonian Walton Formation sampled by McLennan et al. (2001). Detrital rutiles from their sample were derived from a broad range of low to high (greenschist/blueschist to granulite) grade metamorphic rocks, subjected to temperatures between <450 ° and 1050° C. Significantly, coarser grained rutiles (ca. 60% of rutile grains analyzed) derived from high-grade (granulite facies) metamorphic rocks.

NYSGA 2009 Trip 7 – Ver Straeten

Sample number	1	2	3	4	5	6	7	8	9	10	11
Sample (100 count analysis, or %)	Connelly Cgl.	Kanuse SS	upper Mt. Marion Fm.	Ashokan Fm.	basal Moscow Fm.	Twilight Park Cgl. (% of 400 clasts)	Cgl., top of Platte Clove	Avg. cgl comp, Rushford Fm ¹	low Cattaraugus cgl., Alma, NY	Wolf Creek Cgl. ²	Killbuck Cgl. ²
Age	upper Pragian	basal Eifellian	lower Givetian	lower Givetian	mid Givetian	~lower Frasnian	~lower Frasnian	lower Famennian	mid Famennian	mid Famennian	mid Famennian
vein quartz	100	100	40	*	56	66%	47	*	83	98	97
greenish quartz	*	*	1	*	13	*	2	*	*	*	*
total quartz ³	*	*	*	*	*	*	*	87%	*	*	*
gray chert	*	*	25	*	*	*	*	*	*	*	*
lt gray, white chert	*	*	3	*	1	*	*	*	*	1?	*
dark gray chert	*	*	12	*	*	*	*	*	*	*	*
black chert	*	*	3	*	*	*	*	*	*	*	*
banded gray chert	*	*	1	*	*	*	*	*	*	*	*
red chert/jasper	*	*	*	*	*	1%	2	*	*	p	2
sandstone/quartzite	*	*	6	*	10	30.5%	49	*	*	*	*
Siltstone	*	*	*	*	17	1.5%	*	*	*	*	*
“lithic fragments”	*	*	*	*	*	*	*	4%	*	*	*
Feldspar	*	*	*	*	*	*	*	2%	*	*	*
Mica	*	*	*	*	*	*	*	3%	*	*	*
Opagues	*	*	*	*	*	*	*	2%	*	*	*
Calcite	*	*	*	*	*	*	*	2%	*	*	*
clay-rich clast	*	*	*	1	*	*	*	*	*	*	1
calc. nodule clast	*	*	*	99	1	*	*	*	*	*	*
pyrite nodule clast	*	*	*	*	1	*	*	*	*	*	*
siderite nod. clast	*	*	*	*	*	0.5%	*	*	*	*	*
limonite nod. clast	*	*	*	*	1	*	*	*	*	*	*
hematite, SS matrix	*	*	*	*	*	*	*	*	16	*	*
marine shells	*	*	c	*	c	*	*	c	*	*	*
fish bone	*	*	*	*	*	*	*	*	1	*	*
fossil wood	*	*	*	*	*	*	*	*	p	*	*
fossil charcoal	*	*	*	*	*	*	*	*	p	*	*
unidentified	*	*	9	*	*	0.5%	*	*	*	1	*

Table 2. Composition of some Devonian conglomerates, New York State. Preliminary results of Lower to Upper Devonian (Pragian to Famennian) conglomerate analysis in New York State, with data from Smith and Jacobi (1998; sample 8). Based on a 100 clast count, or % of total clasts counted. Abbreviations: c = common; p = present; nod. = nodule.

¹ mean % of 11 conglomerates, Upper Devonian Rushford Fm, from Smith and Jacobi (1998)

² samples from NY State Museum collection; inaccurate locality data

³ total quartz = monocrystalline, polycrystalline and undulatory quartz.

Stratigraphic and geographic positions of samples denoted by numbers on Figure 4.

New Results: Devonian Conglomerates. Up to the present, no systematic analysis of conglomerates is reported from the Devonian Catskill delta complex in New York. Their presence are noted in publications, but few studies examine clast composition in detail (however, see Allen and Friend, 1968; Nickelsen, 1983; Smith and Jacobi, 1998; Friedman, 1998). Recent study provides an overview of compositional changes of conglomerates through the Devonian succession in New York.

Table 2 presents data for 11 conglomerate beds, including results from the two previously noted studies. The conglomerates were deposited in eastern and western New York through approximately 40 million years (upper Pragian to middle Famennian stages, ca. 410-370 Ma). Grain composition through the succession comprise extra- and intrabasinal clasts, including vein quartz, sedimentary and low-grade metamorphic clasts, intraformational nodules and mud clasts, and fossil material. Compositions were generally determined using a standard 100 clast count method, via naked eye, 10x hand lens, or a binocular microscope. Photographs of select conglomerate beds can be seen in Figures 9 and 10.

The coarse clastic beds in Table 2 were variously deposited in shallow marine to fluvial environments. The majority of beds are true conglomerates; some terrestrial beds, however, feature abundant angular, blocky calcareous nodules of pedogenic origin (Figure 9d), and are better defined as breccias. Some conglomerates consist largely to fully of intrabasinal clasts, including reworked nodules of varying composition or ripped-up mud fragments (Figure 9c, d, e). Other ones are chiefly to fully composed of extrabasinal granules, pebbles and/or cobbles. Older and younger beds analyzed are oligomictic in composition, characterized by dominantly quartz grains (e.g., vein quartz +/- chert; Figures 9a, 10c). The intervening beds have a petromict (polymict) character (Figures 9b, 10b), with more diverse clast types including quartz, chert, and pre-lithified sandstone/siltstone pebbles. Many samples are paraconglomerates, with a significant amount of finer-grained (sand- to clay-sized) matrix. Clasts are dominantly pebble-sized, except for granule-dominated sample 2.

Initial Lower and Middle Devonian conglomerates (Samples 1, 2; Oriskany-equivalent Connelly Formation and lower Onondaga-equivalent Kanouse Formation) are composed exclusively of white vein quartz. Succeeding samples, however, show diverse types of clasts, including of both sedimentary and low-grade metamorphic composition. Chert content is highest in Sample 3 (Middle Devonian Mount Marion Formation), where its abundance is roughly equal with macrocrystalline quartz; a small number of sandstone clasts also occur in Sample 3. Chert concentration declines above the Mount Marion conglomerates. Stratigraphically higher beds, from the Upper Devonian Oneonta through Slide Mountain formations, feature significant percentages of sandstone clasts, some of which commonly feature chlorite. Red sandstone clasts, similar to red sandstones in the Devonian Catskill succession, are common in the Slide Mountain Formation (Fletcher, 1967).

The youngest conglomerates studied, from the Late Devonian of western New York, feature abundant white vein quartz (Samples 8-11, Figure 10c). In the two youngest samples, vein quartz concentration approaches the purely vein quartz composition of the two oldest conglomerates (samples 1 and 2).

Volcanic Airfall Tephra in the Foreland Basin Fill. Explosive plinian-type volcanic eruptions result in atmospheric transport of fine ash and crystals, which settle out across all environments downwind. The resulting ash layers, which are subjected to various physical, biological and chemical processes in the respective environments, may or may not be preserved in the rock record. These beds are diagenetically altered over deep time, and become rich in clays or other mineral phases to which other terms are sometimes applied (e.g., bentonite, K-bentonite, tonstein). The author follows the usage of USGS volcanologist A.M. Sarna-Wojcicki (pers. comm., 2008), and applies the term “tephra” to ancient volcanic airfall beds in sedimentary rocks.

Though they comprise a small proportion of sedimentary rocks, volcanic airfall tephra provide key beds for analyses of a rock succession. Regional correlation and geochronologic age-dating of airfall tephra provide high resolution timelines to analyze the relative and geochronologic timing of events and processes active in both a foreland basin and the adjacent orogen (e.g., tectonic flexure, changes in sedimentation, timing and character of explosive volcanic activity).

Approximately 100 tephra, occurring as clay-rich K-bentonites or coarse-grained tuffs, are now known from the Devonian (+/- uppermost Silurian) of the Appalachian basin (Figures 11-12). These include five clusters of 6-15 or

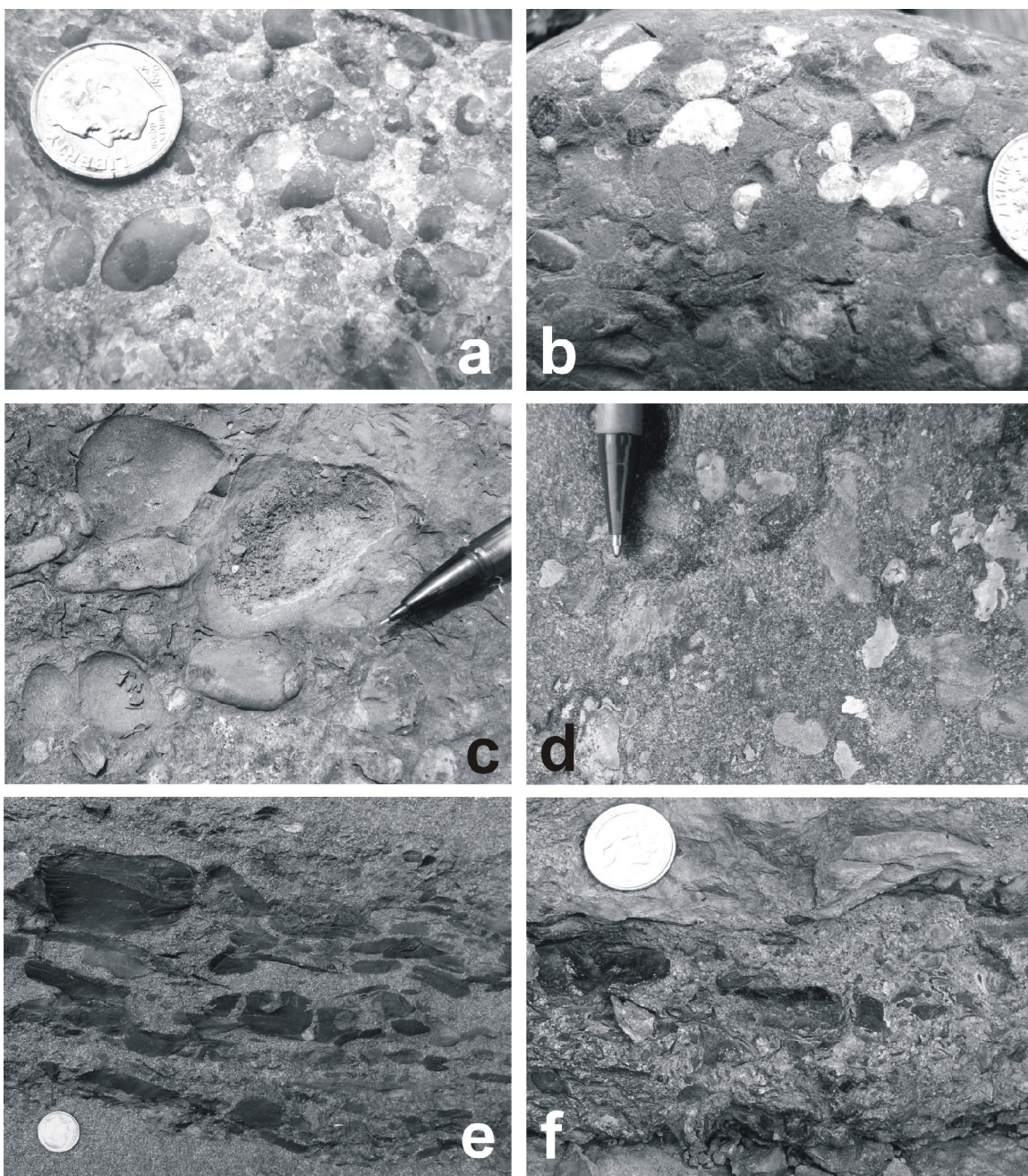


Figure 9. Devonian Conglomerates I. Photos of Devonian conglomerates: a) Milky quartz pebble conglomerate, Lower Devonian Connelly Formation. Pebbles slightly stained by iron. Along west side of NYS Thruway, Schunemunk Outlier, Orange Co.; b) polymict, extraformational conglomerate, upper part of the Middle Devonian Mount Marion Formation, near Catskill. Mix of milky quartz and chert, with sandstone and reworked nodules. Dime for scale; c) intraformational conglomerate of reworked nodules, upper Mount Marion Formation, along Rte. 23, near Leeds (Stop 5). Pencil tip for scale; d) intraformational breccia of calcareous nodules in sandstone, reworked from terrestrial paleosols. Plattekill Formation, Catskill Creek at Rte. 32, Cairo. Point of pen for scale; e) intraformational conglomerate of dark gray mud clasts in sandstone, reworked from floodplain deposits. Loose block of unknown Upper Devonian sandstone, Oneonta or Walton Formation. Dime for scale; f) intraformational biogenic conglomerate of fish bones, plant debris, and possible charcoal. Manorkill Formation, quarry north of Windham. Quarter for scale.

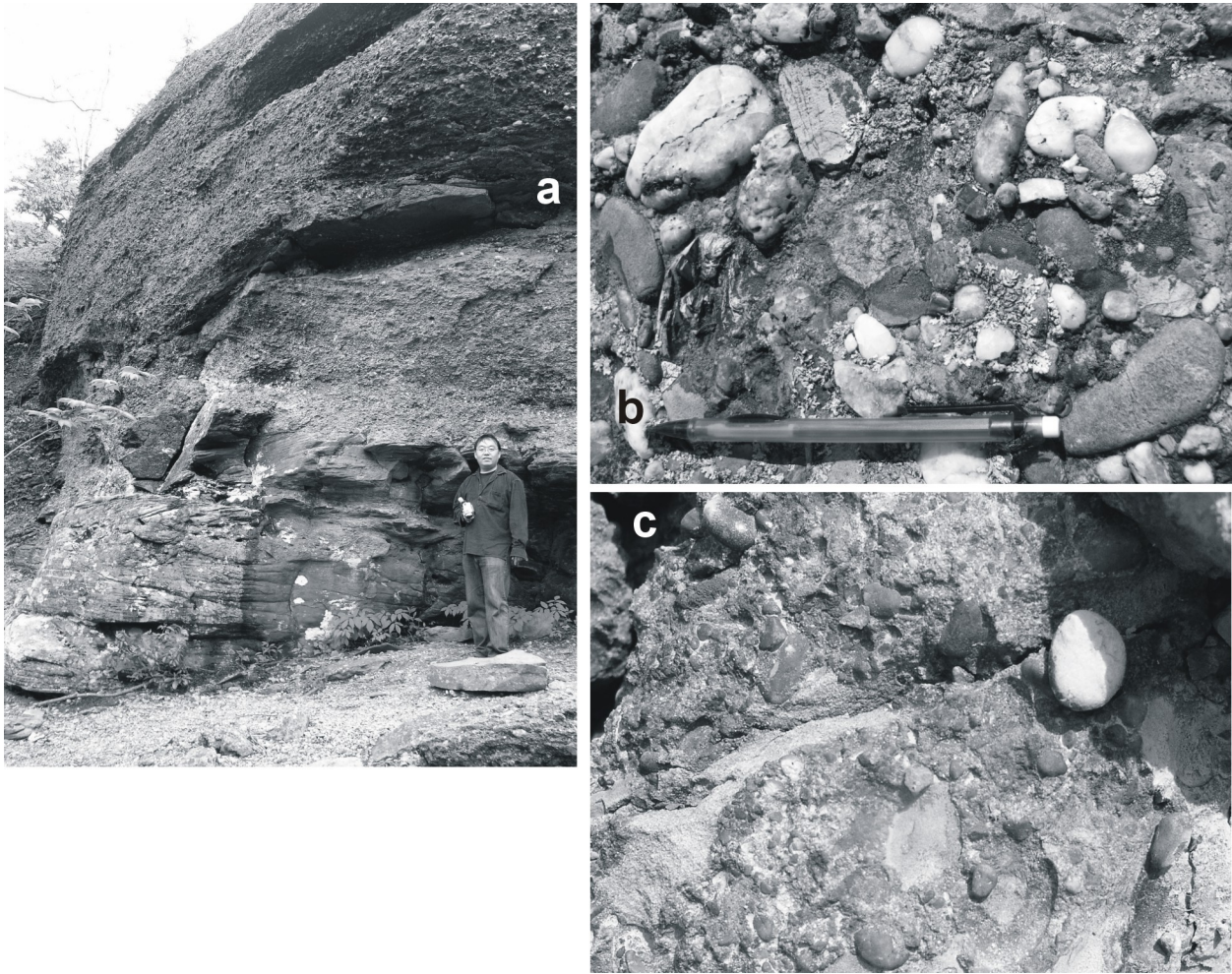


Figure 10. Devonian conglomerates II. Photos of Devonian conglomerates: a) Outcrop of Upper Devonian (Frasnian) Twilight Park Conglomerate (Oneonta Formation), at Sunset Rock, North-South Lake (Stop 8); b) polymict, extraformational conglomerate of mixed clast types. Mix of milky quartz and sandstone clasts, with other less common clasts. Twilight Park Conglomerate, at Sunset Rock, North-South Lake (Stop 8). Pencil for scale; c) Milky quartz-rich extraformational conglomerate, with lesser amounts of sandstone clasts, wood and charcoal. Upper Devonian (Famennian) Cattaraugus Formation, near Allentown, Allegany Co. western NY. Larger pebble is approximately 1 cm in length.

more closely-spaced beds (Figures 12, 13a, b); additional single beds or lesser clusters occur through parts of the succession (Figure 13c-f). Some are found widely across the eastern U.S. (e.g., one or more beds of the Eifelian Tioga A-G zone, and the Frasnian Center Hill K-bentonite), whereas others are only regionally to locally recognized. At present, Lower to lower Middle Devonian (Lochkovian- through Eifelian-age) tephra have been more extensively documented. Additional work is needed through upper Silurian and upper Middle to Upper Devonian strata. The highest concentration of documented tephra beds are in Eifelian strata (>50 beds) across the basin; in contrast, some intervals have as yet yielded few if any beds.

Key references on Devonian tephra include Dennison and Textoris (1970, 1978) Dennison (1983). Conkin and Conkin (1979, 1984), Smith and Way (1983), Way et al. (1986), Smith et al. (1988, 2003), Shaw et al. (1991), Hanson (1995), Ver Straeten (1996, 2004a, b, 2008), Over and Rhodes (2000), Shaw (2003), Ebert and Matteson (2003), Wilcott and Over (2005), Ver Straeten et al. (2005, 2007a), and Over (2007). Information from other geologists include personal communications with G. Baird, D.J. Over, and P. Rubin (2007).

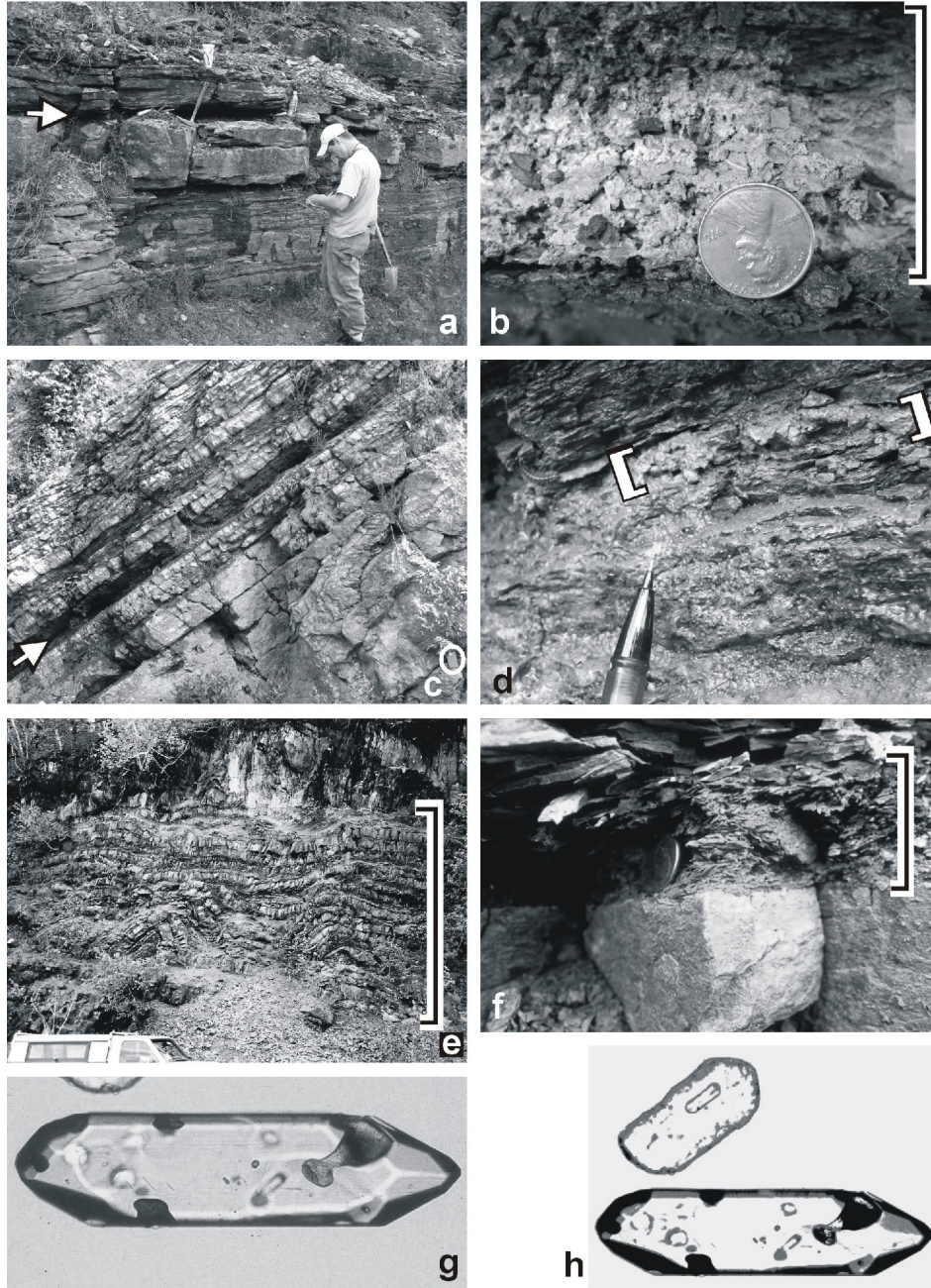


Figure 11. Devonian volcanic tephra I. Devonian airfall tephra beds in foreland basin sedimentary rocks, eastern New York: a) Arrow points to Bald Hill K-bentonite C. Lower Devonian Kalkberg Formation, Rte. 20, Cherry Valley, Otsego Co.; b) close up of clay to claystone of Bald Hill K-bentonite C (bracketed), at same outcrop; c) a thin Bald Hill K-bentonite bed within prominent dark gray shale bed. Middle of Lower Devonian Kalkberg Formation, Rte. 23 Catskill (Stop 1). Fieldbook in lower right (circled) for scale; d) close up of same thin Bald Hill K-bentonite bed as previous photo (bracketed). Pencil tip for scale; e) interval with 15 Sprout Brook K-bentonites. Spawn Hollow Member, Lower Devonian Esopus Formation, Rte. 23a SW of Catskill (Stop 2). Top of pickup truck for scale. f) close-up of a thin Sprout Brook K-bentonite (bracketed), from same outcrop. Dime for scale, just left of center; g) close-up of zircon from a Sprout Brook K-bentonite bed, showing inclusions. Scale not known; h) close-up of same zircon crystal, with apatite crystal.

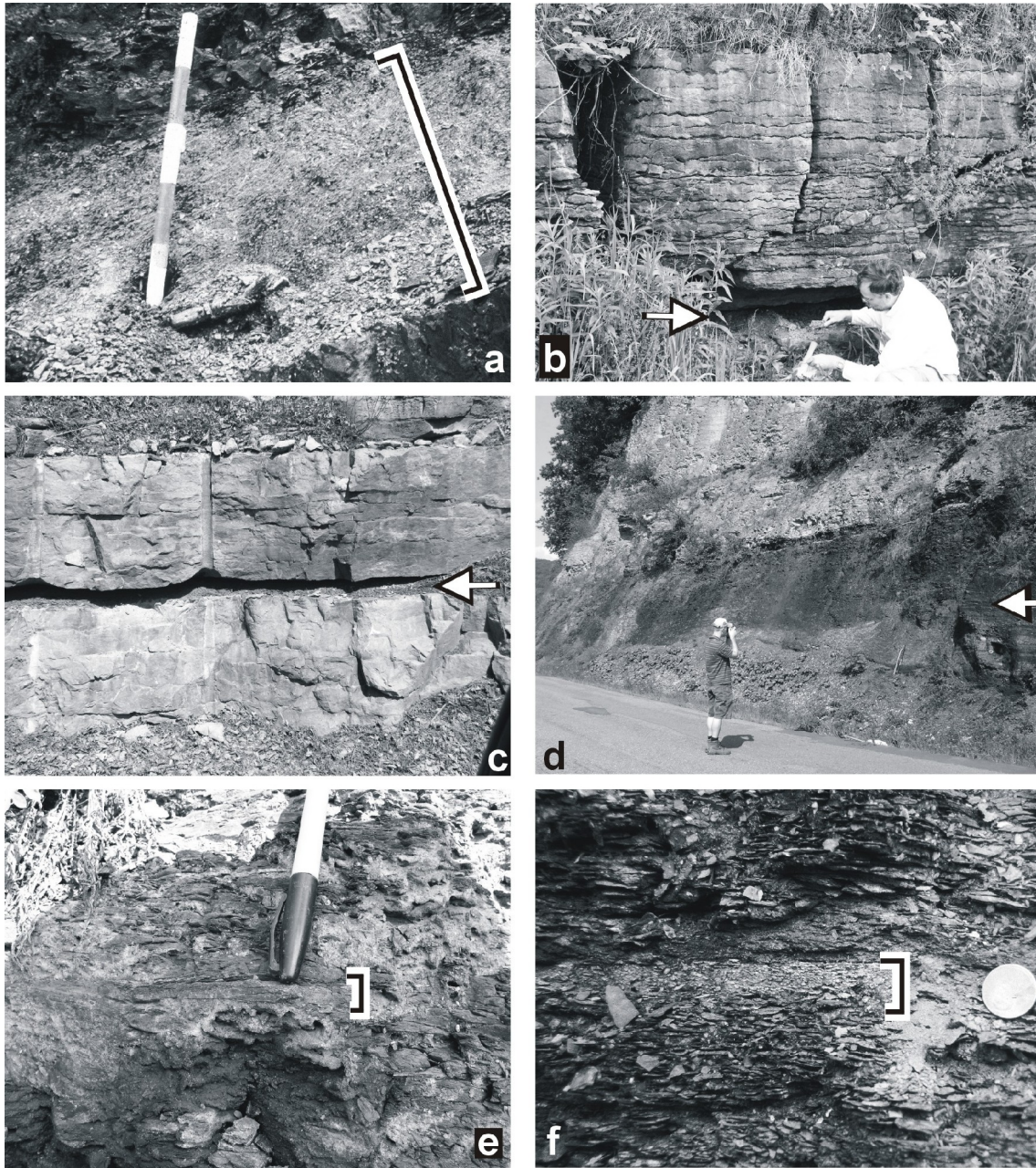


Figure 12. Devonian volcanic tephra II. Devonian airfall tephra beds in foreland basin sedimentary rocks, eastern New York: a) Topmost Sprout Brook K-bentonite (bracketed), from same outcrop as Figure 12e, f. Bed locally thins and thickens along the outcrop, associated with structural deformation of strata. Bands on stick = 10 cm; b) Tioga B K-bentonite in recessed interval, from Tioga A-G K-bentonites cluster. At base of Seneca Member, Middle Devonian Onondaga Formation, Rte. 20 cuts at Cherry Valley. George Shaw for scale; c) arrow points to thin (ca. 7 cm-thick), unidentified K-bentonite in upper part of Onondaga Formation (Moorehouse Member). Rte. 85, north of Clarksville, Albany Co.; d) arrow points to mid-Union Springs K-bentonite of Ver Straeten (2004a). In upper part of Bakoven Member, below contact with Stony Hollow Member, Middle Devonian Union Springs Formation. Along City View Terrace, off Rte. 28, northwest of Kingston; e & f) close-up of mid-Union Springs K-bentonite at same outcrop. Brackets denote position of cm-scale bed. In photo f, the bed is distinctly replaced by pyrite. Marker pen and penny for scales.

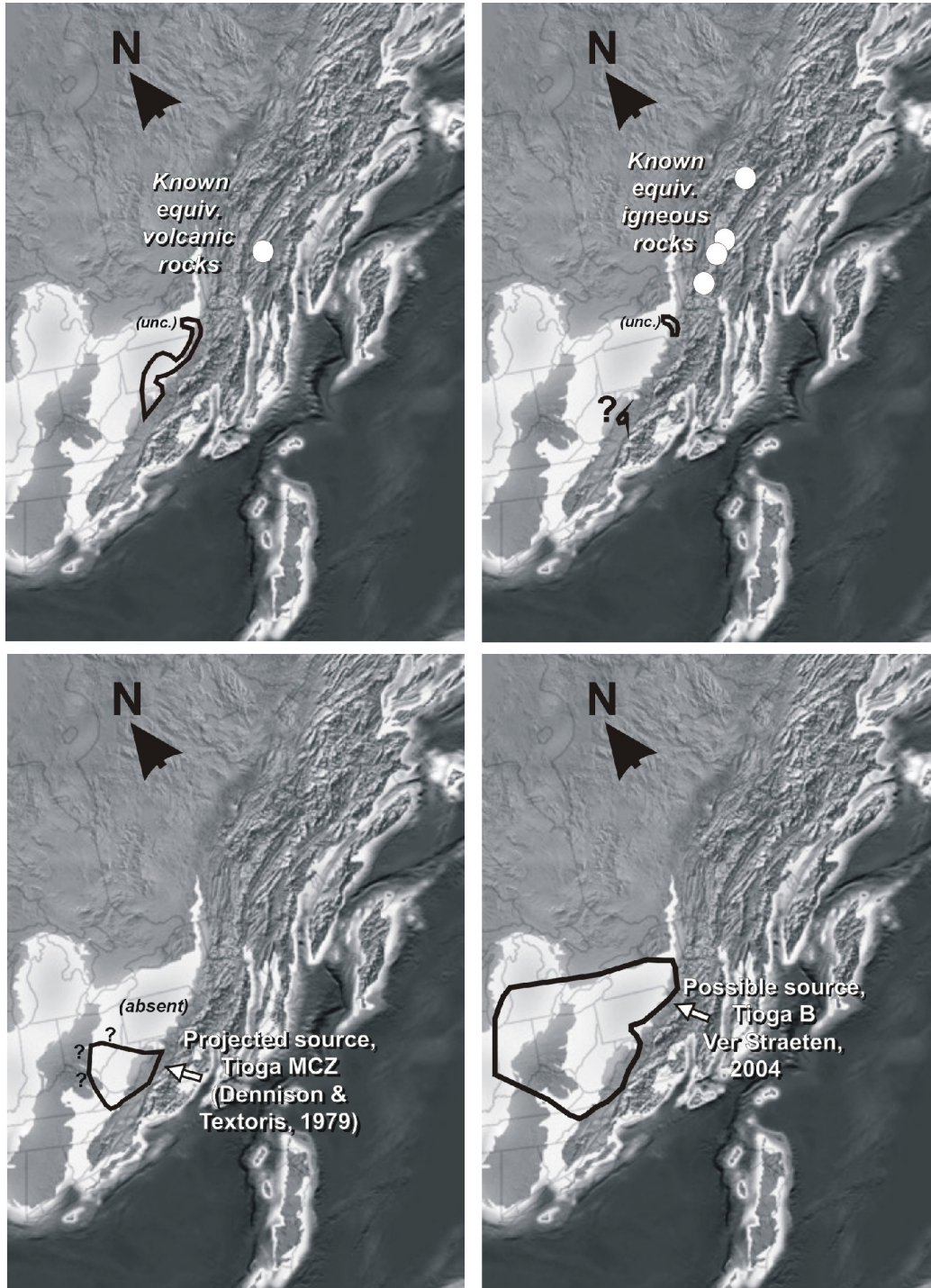


Figure 13. Lower to Middle Devonian tephra clusters, Appalachian basin, and possible source areas. Map distribution of the Bald Hill, Sprout Brook, Tioga Middle Coarse Zone, and Tioga A-G tephra clusters are plotted over a paleogeographic map of eastern Laurentia in the Middle Devonian. Potential source areas in a and b based on known age-equivalent volcanic and plutonic rocks in the Acadian orogen and greater Acadian foreland basin; projected source areas in c and d after Dennison and Textoris (1978) and Ver Straeten (2004a). Base map from Blakey (<http://jan.ucc.nau.edu/~rcb7/namD385.jpg>).

Most of the Devonian airfall tephtras occur as clay rich beds; the clays are derived from alteration of volcanic glass. In contrast, some beds are essentially sandstones, formed of sand-sized minerals (phenocrysts) that were ejected from the magma chamber with volcanic ash during eruptions. The clay-rich beds are generally termed K-bentonites or metabentonites, due to the high content of potassium-rich illite; sandy phenocryst-rich beds may be termed “tuffs”. Early diagenetic weathering of the abundant volcanic glass in the K-bentonites would have initially yielded smectitic clays (e.g., montmorillonite), which over deep time were altered to illite or mixed layer illite-smectite. Colors of the clay beds may vary, but commonly have a yellow-tan appearance. The clay rich K-bentonites beds commonly form recessions within the surrounding strata. Most all beds also at least a small percent of phenocrysts; these generally appear pristine and unabraded, euhedral in form or broken and shattered, with sharp edges. Common phenocrysts include zircon, apatite and bipyramidal quartz +/- quartz shards. Geochemical analysis of melt inclusions within metastable quartz crystals indicate that source magmas for the Bald Hill, Sprout Brook, and Tioga A-G clusters were of a high-silica rhyolitic composition (Hanson, 1995).

The Major Clusters Of Lochkovian To Eifelian Airfall Tephtras. The Lower Devonian (Lochkovian) Bald Hill K-bentonites cluster comprises of up to 15 airfall tephtras (Figure 11a-d). They occur widely across the Appalachian foreland basin, from New York to Virginia and West Virginia (Smith et al. 1988, 2003; Shaw et al. 1991). One of the beds yielded a U-Pb age of 417.6 +/- 1.0 Ma (Tucker et al. 1998). In New York, they occur within the Kalkberg and New Scotland formations. The documented distribution of the Bald Hill K-bentonites across the Appalachian basin is shown in Figure 13a.

The Lower Devonian (lower Emsian) Sprout Brook K-bentonites, also consist of up to 15 airfall tephtra beds (Ver Straeten, 2004b; Figures 12e-h, 13a). U-Pb dating of zircons yielded an age of 408.3 +/- 1.9 Ma (Tucker et al. 1998). They occur in the lower part of the Esopus Formation (Spawn Hollow Mbr.), in eastern New York (from Cherry Valley, Otsego Co. eastward). Locally in Virginia, a bed of mixed terrigenous and volcanic grains has been found in the same position, in the lower part of the Needmore Formation (lower part of the Beaverdam Member; Conkin and Conkin, 1979; Ver Straeten, 2004b). The Sprout Brook K-bentonites appear to be restricted to the northeastern part of the Appalachian basin (Figure 13b).

The Middle Devonian (lower Eifelian) Tioga Middle Coarse Zone of Dennison and Textoris (1970, 1978) is restricted to the southern part of the basin, where it is characterized by three prominent tuff beds, with up to 29 additional minor beds. Zircons from one bed have been dated at 391.4 +/-1.8 Ma (Tucker et al., 1998). The Tioga Middle Coarse Zone occurs in upper strata of the Needmore Formation, correlative with the lower part of the Moorehouse Member of the Onondaga Formation in New York (Ver Straeten 2004a, 2007a). Figure 13c illustrates the distribution of the Tioga Middle Coarse Zone across the Appalachian basin.

The Middle Devonian (lower Eifelian) Tioga A-G K-bentonites cluster (Smith and Way, 1983) consists of eight to nine beds that are widely correlatable through the Appalachian basin, from New York to southwestern Virginia, and into central Ohio (Ver Straeten, 2004a, 2007a). The Tioga B bed yielded an age of 390 +/- 0.5 Ma (Roden et al. 1990). In the northern part of the basin the A-G cluster occurs in the upper part of the Onondaga Formation and basal Union Springs Formation (Figure 13b); to the south they are found in the Selingsgrove Member (PA) or calcareous shale and limestone member of the Needmore Formation, and/or the lower part of the Marcellus Shale or Millboro Formation (MD, VA, WV). In Ohio they occur in the upper part of the Columbus Formation and lower part of the Delaware Formation (Ver Straeten, 2007a; DeSantis et al., 2007). At least one Tioga bed has been identified as far west as the Illinois basin. The distribution of the Tioga A-G cluster is shown in Figure 14d.

The Sprout Brook K-bentonites are the best represented tephtras in the Catskill Front (Figures 12e-h, 13a). Though not commonly exposed, a number of complete outcrops can be found from Columbia and Greene to Otsego counties (including Stop 2). Up to 15 K-bentonites are interbedded with siliceous siltstones, impure cherts and minor shales in the lower part of the Spawn Hollow Member (lower part of the Esopus Formation). They commonly appear as yellow-tan unlithified clay beds, ranging in thickness from >1cm to ~20 cm, except where they are structurally modified in deformation zones (e.g., Stop 2, Figure 13a).

Independently derived correlations of Sprout Brook K-bentonites are shown in Figure 14 (Ver Straeten et al., 1993). The figure shows an upper set of physical correlations made by the author, and a lower set of independently derived

correlations based on the geochemistry of glass inclusions in quartz phenocrysts by B. Hanson (Hanson, 1995). Clear physical and geochemical correlations closely match for outcrops in the Hudson Valley. However, correlations to the west (e.g., Knox, Cherry Valley) by either method was less clear.

Close observation of the columns in Figure 14 from the Hudson Valley delineates distinct packages of tephra beds with siliceous siltstones, cherts and minor shales. Six such packages seem to occur through the interval of the Sprout Brook K-bentonites.

Soft Sediment Deformation. Some Devonian strata in the Catskill Front show clear indications of deformation previous to lithification (Figure 15). These are marked by distorted to convolute layering, ball and pillow structures, and other features characteristic of soft sediment deformation zones (SSDs). SSDs commonly, though not exclusively, occur in beds of alternating grain size (e.g., muds and sands/carbonates), and may form through several different mechanisms.

At present, observations in eastern New York by previous workers and the author indicate a concentration of SSDs in the upper part of the marine succession (ca. upper middle Mount Marion Formation; “storm rollers” of Chadwick, 1944; Wolff, 1967; Stop 4, Figure 15a-e). Higher in the succession, in terrestrial strata, Fletcher (1967) reported SSDs in the base of the Oneonta Formation, and Gordon (1986) reported rare load casts and convolute bedding; in addition, some “highly disturbed” zones of Willis and Bridge (1998) may represent additional SSD zones. The author has locally noted SSDs in the Manorkill Formation (Stop 6) and the lower Oneonta Formation (Stop 7; Figure 15f, g).

Through the entire Devonian succession in the Catskill Front, however, SSDs are rarely observed or reported. Though this may in small part due to the difficulty of recognizing SSDs in homogenous facies (e.g., mudrock-only lithologies), it largely reflects their absence.

In upper middle to upper parts of the Mount Marion Formation, SSDs can be seen all along the outcrop belt from Kingston to the Helderbergs (Figure 15a-e). One area to observe multiple SSDs in the Mount Marion Formation is along and adjacent to NY Rte. 28 northwest of Kingston (Ulster Co.). Stratigraphically lower SSDs in this succession are seen along Moray Hill Road, near Stony Hollow, where three deformed zones can be seen in the upper approximately 12.5 m of a road cut. The lowest SSD zone comprises a 0.6 m-thick bed of internally-deformed sandstones; lower and upper bounding strata are undeformed. The middle zone features isolated pillows of sandstone in a sandy mudstone through a 20-30 cm-thick interval, underlying an undeformed argillaceous sandstone unit. The upper zone is not so subtle; the deformed zone is up to 3.3 meters thick, with contorted laminations and bedding, ball and pillow structures with a distinct wrinkled surface that marks the passage of fluidized muds +/- water along their margins. Also visible are four large, isolated sandstone boudins, ranging from 0.35 x 2.2 m to 0.50 to 4.3 m in size (Figure 15e). Three of the boudins are horizontal, but one is distinctly tilted at an angle. In at least one spot, a zone of plastic deformation is dissected by small vertical to near-vertical faults,

Northwest of Moray Hill Road on Rte. 28, additional outcrops show increasing degrees of soft sediment deformation up through the succession. One very large pillow, or “bowl”, is visible in one of the middle outcrops (Figure 15a). A stratigraphically higher outcrop, a short distance northwest, features multiple zones of SSDs.

Similar features are noted where upper-middle Mount Marion strata are exposed, such as NY Rte. 23 near Quarryville (Greene Co.), NY Rte. 32 east of Westerlo (Albany Co.), and along the Hamilton escarpment in the Helderbergs (Albany Co.). Many of these SSDs are associated with interbedded sandstone and shale layers.

A different style of soft sediment deformation is visible in the upper part of the Mount Marion Formation off of NY Rte. 23, northwest of Catskill (Stop 4). Along the south side of the 12.3 m-thick outcrop, the uppermost, 0.9 m-thick sandstone bed features obvious SSD features (Figure 15b). Ball and pillow deformation and contorted strata is accompanied by a wrinkled surface, indicative of the plastic flow of fluidized sediments along the surface of the ball sediments. In this case, it appears that no mud was involved in the deformation. Instead, liquification of the sands following a triggering event led to dewatering, repacking of the sand grains, and deformation. An additional SSD occurs in sandstones about 40 cm below the base of the uppermost sandstone bed.

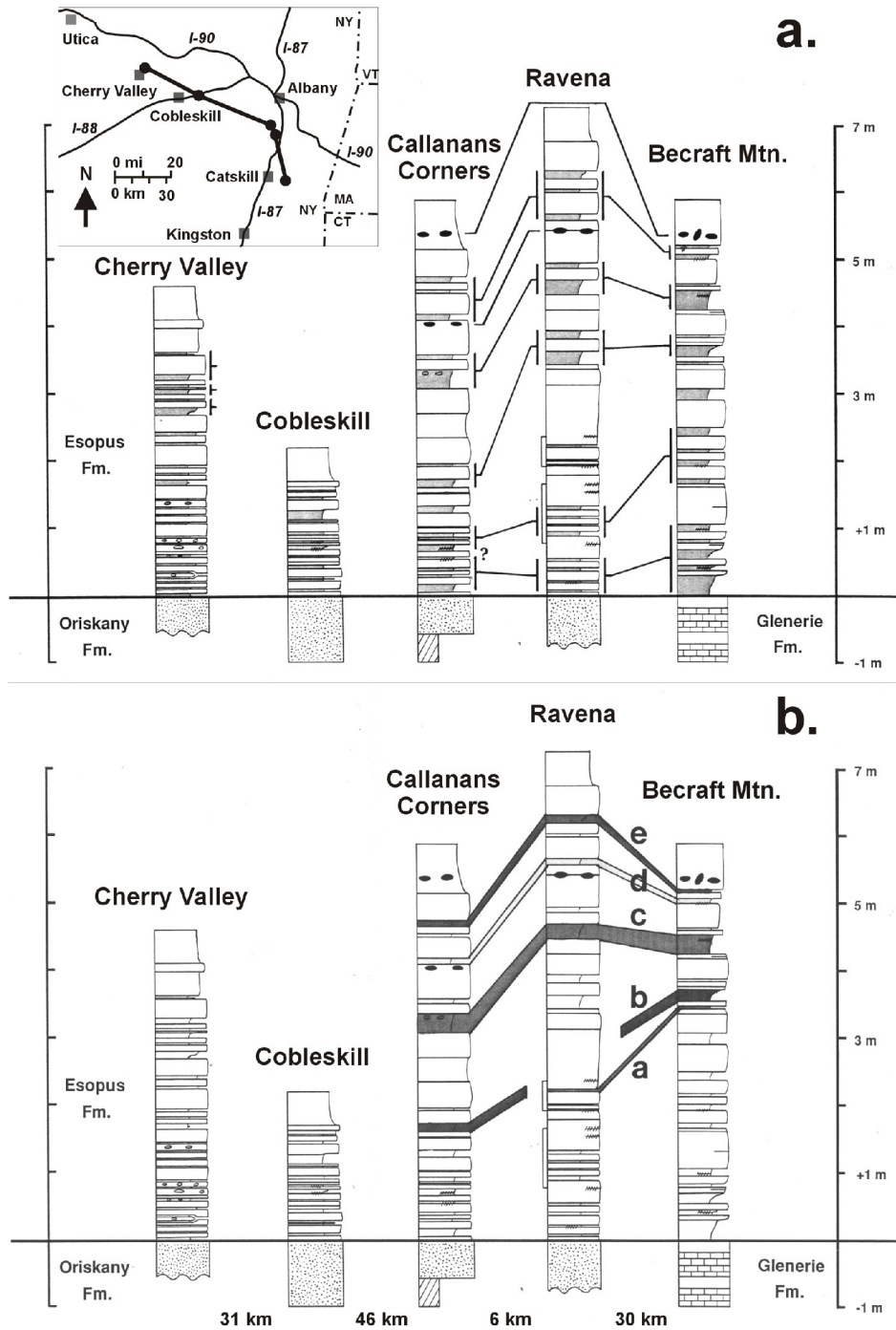


Figure 14. Correlations of the Sprout Brook K-bentonites, eastern New York. Independently-derived correlations of the Sprout Brook K-bentonites cluster: a) physical correlations, by the author; b) geochemical correlations by B. Hanson. Originally presented by Ver Straeten et al. (1993). Both methods yielded essentially the same results in the three easternmost outcrops; correlations to the west are less clear. Geochemical correlations based on composition of rhyolitic inclusions within volcanogenic quartz within individual Sprout Brook beds (Hanson, 1995). Note distinct packages of K-bentonites, separated by marine strata, which can be correlated physically in easternmost outcrops.

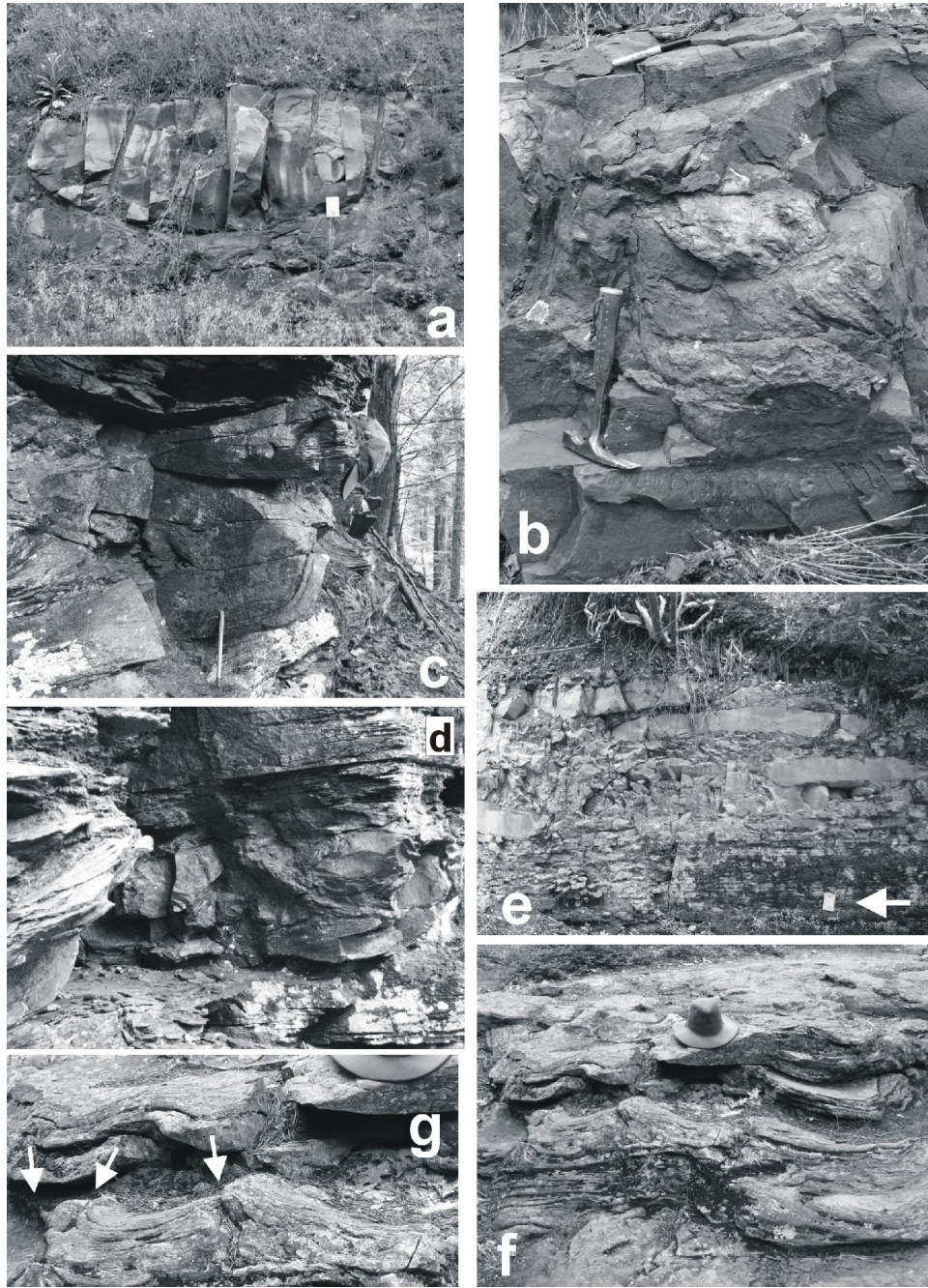


Figure 15. Soft-sediment deformation features, Catskill front. a) large dish/pillow structure of sandstone in mixed sandstone and shale. Upper middle part of the Middle Devonian Mount Marion Formation, Rte. 28, northwest of Kingston. Fieldbook for scale, at lower right of dish; b) pillow-like structures, with “wrinkled” surfaces, developed in sandstone-only strata. Upper part of the Mount Marion Formation, Rte. 23 near Leeds (Stop 5); c) various soft sediment deformation structures in mixed sandstone-shale facies. Upper middle part of the Mount Marion Formation, on the author’s land, East Berne, Albany County. Pen for scale at lower center; d) structurally complex SSD zone, between undisturbed strata of similar facies; same outcrop as photo c; e) large, isolated sandstone boudins within ca. 3.5 m-thick deformed zone. Upper middle part of the Mount Marion Formation, Moray Hill Rd., NW of Kingston. Field book in lower right for scale; f) SSDs in sandstone only fluvial channel facies. Lower part of Upper Devonian (?) Oneonta Formation, along trail to Artists Rock and Sunset Rock, North-South Lake. Note rough similarity of structures to trough cross-bedding; however, margins of troughs are vertical water escape structures.

In the Oneonta Formation, along the crest of the Catskill Escarpment, one or more SSDs have been noted in Kaaterskill Clove and near North-South Lake (Fletcher, 1967; Gale, 1985; this paper; Stop 7, Figure 15f, g). When first viewed along the trail to Artists Rock at North-South Lake, one interval of sandstones appear to feature numerous trough cross-beds. However, a closer examination indicates that the edges of the troughs are at near-vertical to vertical angles, far beyond the angle of repose. Those edges actually represent dewatering structures, and the troughs, at least in part, represent lows where sand grains settled lower, via foundering and post-liquifaction repacking into more condensed deposits.

One of the interesting characters of many SSDs in the Catskill Front is their occurrence within specific layers, surrounded by layers of similar lithology throughout the rest of an outcrop. Most or all of the surrounding strata show no deformation.

DISCUSSION

Sedimentation and Sea Level History

Sedimentation Trends and the Acadian Orogeny. The major carbonate-quartz arenite and initial terrigenous clastic packages of strata above the Wallbridge unconformity feature markedly different patterns of sediment type, depth-related facies, and stratal distribution and geometry.

Variations in sediment type between the intrabasinal carbonates (+/-quartz arenites) versus extrabasinal clastics are, of course, related to changes in sediment source/provenance. Most of the carbonate is derived from biogenic production of shell matter by marine organisms (e.g., brachiopods, crinoids). Quartz arenites, including quartz conglomerates, are largely derived from reworking of supermature sediments within or on the fringes of the basin. Terrigenous clastics, in contrast, derived from erosion of previously existing rocks eroded from regional highlands and transported into the basin

The relatively widespread distribution of the limestones of the Helderberg Group, along with the Glenerie-Oriskany, Onondaga, and Tully formations reflect widespread shallow seas, characterized by relatively high carbonate production rates by shelly organisms living in shallow, relatively clastic-free tropical seas. The distinct wedge-shaped geometries, especially of the initial clastics overlying the carbonates indicate the influx of detrital materials eroded from a regional source area, accompanied by close to full cessation of carbonate production with clastic input.

The contrasting pattern of relatively shallow versus deep water litho- and biofacies of carbonate versus initial clastic deposits is associated with flexure of the crust underlying the foreland basin system. Crustal loading associated with orogenic buildup leads to subsidence of the crust, the shutoff of the carbonate production, and initial sediment starvation with transgression. Subsequent low sedimentation rates of suspended fine-grained sediments, combined with deposition/preservation of organic matter in deep water anoxic sediments, leads to the formation of basinal black shales.

Another contrast between the carbonate versus initial clastic suites is the widespread nature of the former and the more proximally-restricted distribution of the latter. Carbonate production, active across the basin during limestone deposition, shut down with the onset of clastic deposition, even in distal, clastic-starved areas of the foredeep basin in New York.

These major patterns, associated with development of three to four separate clastic wedges punctuated by carbonates, were interpreted by Ettensohn (1985) to reflect tectonically-active to -quiescent phases (“tectophases”) during the Acadian orogeny. He proposed four separate Acadian tectophases, from the upper Lower Devonian through Lower Mississippian. The onset of tectonism in each tectophase resulted in subsidence of the basin foredeep and deposition of the basinal dark gray to black shales. Subsequent development through a tectophase culminated in a return to tectonic quiescence and deposition of carbonates. In general, Ettensohn’s (1985) basic model still appears to explain a number of trends in the foreland basin sedimentary rock record. The relationship of his model to data and interpretations by researchers working in the orogen itself is still largely unclear to the author. More discussion of sedimentary evidence of orogenesis and the tectophase model are needed with peers working in the orogenic belt.

We now know that Acadian orogenesis began in Late Silurian in Maine and New England (e.g., Bradley et al., 2000), and progressively migrated cratonward through the Lower to Upper Devonian. The earlier initiation of the orogeny was not readily visible to Ettensohn (1985). This is now understood by the author to be due to: 1) early in the Acadian event, the orogen, its deformational front, and most of the Acadian foreland basin system (wedge-top, foredeep and forebulge) were still geographically positioned in New England to easternmost New York, and; 2) Ettensohn (1985, 1987) was working with data from the eroded remnant of the greater Acadian foreland basin, which starts on the west side of the Hudson River, far cratonward of the Late Silurian interface of the orogen and foreland. During the initial stages of the Acadian orogeny, the Catskill front was geographically situated in the far cratonward backbulge basin of the foreland basin system – marked by deposition of the Rondout and Helderberg carbonates. Portions of the early Acadian foredeep are preserved in northern New England, in thick, flysch units like the Lower Devonian Littleton Formation of Vermont and New Hampshire.

Through time, the orogenic front and successive portions of the foreland basin (wedge-top, foredeep, and forebulge) migrated cratonward, some of which passed west of the position of today's Hudson River. By the Late Devonian, the wedge-top portion of the basin, characterized by deformational thrusting and uplift, may have migrated into the Catskill Front. Interpretation of this is dependent on whether any of the structural deformation in the Catskill Front is of Devonian/Acadian age – a point of ongoing debate (e.g., Geiser and Engelder, 1983; Marshak, 1986; Marshak and Tabor, 1989; Zadins, 1989).

Eustatic and Tectonic Effects on Relative Sea Level. Two of the most important factors affecting relative sea level change are global eustatic sea level processes and regional, tectonic-induced subsidence. There have been many debates over the relative effects of these two processes in foreland basin systems. A series of third order depositional sequences/sea level cycles from the Devonian Appalachian basin Devonian are outlined in Figure 8. Many of the sequences have been correlated outside of the basin (e.g., Johnson et al., 1985; House and Kirchgasser, 1993; Bartholomew, 2006), indicating their global, eustatic nature. Figure 8 compares relative sea level curves for the Appalachian basin sequences and Johnson et al.'s (1985) Euramerican curve. The Appalachian basin sequences show some refinements over the Euramerican curve (e.g., base of transgressions placed at position of lowstand or initial transgressive litho- and biofacies; cycles added where Johnson et al., 1985 overgeneralized the sea level history). Accounting for these refinements, data from New York and globally indicate correlation of individual sequences/cycles (e.g., House and Kirchgasser, 1993; Brett and Baird, 1996; Over, 2002; Filer, 2002; Ver Straeten, 2007a). However, where the Euramerican curve shows overall transgression from the middle Lower Devonian to middle Upper Devonian, the Appalachian basin curve is marked by three major transgressive to regressive pulses, superposed over the record of third order cyclicity. These are the effects of tectonically-induced loading and subsidence in the foreland basin. The three major transgression mark the onset of three separate Acadian tectophases (Tectophases 1 to 3 of Ettensohn, 1985). Of importance, it is clear that these major subsidence events are superposed over the distinct record of eustatic sea level cycles in the foreland basin.

Clastic Rock Petrology and Provenance Changes in detrital grain mineralogy through foreland basin successions provide information about changes of the rocks exposed within source areas through time (provenance). The presence of abundant quartz, K-rich feldspars and granitic rock fragments, or minerals such as staurolite, kyanite and high grade metamorphic rocks fragments indicate very different rocks exposed in their respective sedimentary source areas. These may reflect changes of provenance, although the record may also be affected by grain size variation, transport distance, or depositional environment (Gale, 1985).

Conglomerates Discussion. The new conglomerate analysis presented here provides additional perspectives on changing clastic composition and provenance over approximately 40 million years, (ca. 410 to 370 million years ago). Initial vein quartz conglomerates (mid Lower Devonian Connelly Conglomerate and uppermost Lower to basal Middle Devonian Kanouse Sandstone) at ca. 410 and 395 Ma were likely derived from erosion of intrabasinal sources, such as the Silurian Shawangunk and correlative Green Pond Formations in southeastern New York and adjacent New Jersey.

Succeeding quartz- and chert-rich polymict conglomerates in the upper Mount Marion Formation (Sample 3, ca. 388 Ma) clearly indicate a change in source area. The introduction of the chert and sandstone clasts, not present in the lower samples, is the result of deposition of synorogenic sediments from outside the basin. The varied colors of the

cherts (i.e., gray, light gray to white, dark gray to black, and red) imply the erosion of multiple levels of strata in the source area. Various colored cherts are found in Cambrian and Ordovician strata east of the Hudson River – although no green cherts, characteristic of some of those rocks (e.g., Mount Merino Member, Normanskill Group) have been noted.

The overlying Upper Devonian conglomerates of the Oneonta Formation (Samples 6 and 7, ca. 384 Ma) to Slide Mountain Formation (based on general statements in Fletcher, 1967; Gale, 1985; ca. 376 Ma) are characterized by abundant quartz and sandstone clasts. Chert content has declined to almost negligible amounts. Chlorite is found in a number of the sandstones, which may indicate low grade metamorphism of some of the strata, or incorporation of low-grade metamorphic sediments into the sandstones. The latter could have happened during the Lower to Middle Devonian, off to the east before the foredeep of the basin migrated westward into the Hudson Valley area. Their occurrence could be explained by uplift, exposure and erosion of those sandstones above thrust fault slices in the wedge-top of the foreland basin system. This would be consistent with Fletcher (1967) proposal that the red sandstone pebbles in the Slide Mountain Formation may have been cannibalized from Devonian sedimentary rocks further to the east. However, further petrologic and possibly palynological study is needed to resolve this issue.

The four Famennian-age conglomerates (ca. 376-370 Ma) from western New York are quartz-rich (83-98%), especially the two younger mid-Famennian samples. These nearly compare to the oldest conglomerates in the succession, which were 100% macrocrystalline vein quartz. However, the composition of the younger Famennian conglomerates may have biasing influences different from all of the older samples. Clastics in western New York may have been derived from a broader or different source area along the orogen (e.g., beyond eastern New York and/or Pennsylvania outcrops), yielding a different pebble composition from those in the Catskill Front. Their occurrence far into the foredeep basin also implies a greater transport distance for the clasts. The conglomerate from the lower Cattaugus Formation, near Amla, New York lies approximately 340 km west of Catskill escarpment. How far it would be to similar sources of clasts southeast of the Pennsylvania outcrop belt is unknown to the author. Long distance transport through high gradient wedge-top to low gradient delta plain streams wear down and destroy less durable clasts (e.g., Cameron and Blatt, 1971; Ethridge, 1977; Davies and Moore, 1970). So, the great transport distance would favor the preservation of highly durable quartz pebbles over other clasts. How much influence this had on the composition of the Famennian conglomerates in New York is unknown. However, at least in part, it likely represents a continuance of increasing quartz content noted in the Catskill Front through the Middle to Upper Devonian Plattekill through Slide Mountain formations.

To summarize, extrabasinal Devonian conglomerates record changing input of coarse clastics into the Catskill Front and beyond. Initial milky quartz gravels (at ca. 410 and ca. 395 Ma, upper Pragian and lower Eifelian) were succeeded compositionally by mixed chert- and quartz-rich gravels (at ca. 388 Ma, lower Givetian), to slightly increasing quartz-rich gravels with abundant sandstone-/metasandstone-rich pebbles (ca. 383 to 376 Ma, lower Frasnian). Through the latter interval, increased numbers of red sandstone pebbles in the gravels appear to indicate erosion and cannibalization of Acadian-derived sandstones (Fletcher, 1967). These would have most likely been exposed by thrusting in the proximal wedge-top depozone of the foreland, and subsequently transported into the foredeep. The youngest conglomerates from this study (ca. 376-370 Ma, lower to mid Famennian) continue the arc of increasing quartz content, culminating in nearly pure milky quartz compositions approaching those of the lowest conglomerates.

Sandstone and Mudrock Provenance. Sandstone petrographic studies of the Lower to Upper Devonian Catskill Delta succession (e.g., Lucier, 1966; Allen and Friend, 1968; Way, 1972; Rehmer, 1976; Ethridge, 1977; Gale, 1985) repeatedly found little to no mineralogical evidence of significant igneous or medium to high grade metamorphic rocks exposed within Acadian drainage basins supplying sediments to the Catskill Front. Instead, the mineralogical evidence largely points to exposure and erosion of low-grade metamorphic rocks (up to greenschist grade) and sedimentary clastics and minor carbonates within the source area. Some very minor exceptions to this pattern were reported by some authors in the sandstones. For example, Lucier's (1966) data on heavy minerals in five samples found traces of medium to high grade metamorphic and igneous rocks (e.g., staurolite, hypersthene, hornblende and diopside-augite). In addition, Allen and Friend (1968) report clasts of pyroclastic tuffs and metamorphic granulites in the Twilight Park Conglomerate.

The low-grade metamorphic and sedimentary rock grains predominantly found in the Middle to Upper Devonian sandstones are lithologically similar to Cambrian to Ordovician sedimentary to low grade metamorphic rocks now exposed in eastern New York (Lucier, 1966; Fletcher, 1987). However, unlike proposed by Lucier (1966) and others authors, it seems unlikely that those rocks were exposed at the surface in eastern New York between approximately 388 to 375 million years ago.

Hosterman and Whitlow (1983), in their study of Middle to Upper Devonian marine shales, hint at a another, disguised sediment source. Their analytical results led them to interpret that during the Devonian, much of the illite and mixed layer illite-smectite now in the rocks was originally composed of illite and smectite clays. As smectite clays form from the weathering of feldspars, volcanic ash, and other similar rocks, their occurrence would indicate a more significant input of sediment from igneous sources than generally interpreted.

In support of Hosterman and Whitlow's (1983) interpretation of substantial smectite content in the Devonian rocks, paleosol studies in the Catskill magnafacies of New York and Pennsylvania note the abundance of vertisols, along with entisols, inceptisols, and alfisols (Driese and Mora, 1993; Mora and Driese, 1999; Oliver and Terry, 2009). Modern vertisols are characterized by a relatively high content of smectite (montmorillonite) clays. Vertical cracks, and deformational features such as pedogenic slickensides, bowls/gilgai/pseudoanticlines and other structures are diagnostic features of vertisols. These result from wetting and drying of the soils (commonly seasonal), and resultant expansion and contraction of the smectitic clays. Though substantial amounts of smectite are not preserved in the Catskill succession today, the widespread occurrence of vertisols also appear to indicate its strong presence during the Devonian.

Apparent substantial smectite content in the strata during the Devonian appear to indicate a hidden source of igneous-derived sediment. Interestingly, Hosterman and Whitlow (1983) note the similar alteration of Devonian airfall tephra to K-bentonites. These beds, originally composed of volcanic glass and phenocrysts, were diagenetically altered to smectitic clays, and then further altered to mixed layer illite-smectite clays and illite.

To summarize, these clay mineralogy and paleosol studies indicate that during the Devonian, fine-grained sediments in New York and the Appalachian basin had a significant component of smectitic clays, apparently derived from the weathering of igneous rocks or volcanic ash. This sharply contrasts with the data from sandstones and conglomerates outlined above, which portray a source very restricted to sedimentary to low grade (slate to greenschist) metamorphic facies.

More recent provenance studies, using different techniques, also portray a more complex picture of sediment provenance in the Catskill front and adjacent areas. Aronson et al.'s (1994) dates of detrital micas (mostly 406-387 Ma) from Upper Devonian to Lower Mississippian clastics in New York and Ohio appear to indicate the much of the Catskill clastic wedge have an Acadian-age provenance. They calculated that Taconic- and Grenville-age sources could comprise no more than about 30% and 5% of sedimentary sources, respectively.

McLennan et al.'s (2001) U-Pb ages of detrital zircons from lower Walton Formation yielded SHRIMP U-Pb ages of 1.23-1.00 Ga (Grenville age) and 470-420. In the absence of clear Acadian age detrital zircons in their single sample, they projected that Acadian sediments appear to recycle pre-existing, pre-Devonian rocks from along the continental margin of Laurentia.

Zack et al.'s (2004) geothermometry analysis of detrital rutiles from the same outcrop as McLennan et al. (2001) indicated that the rutiles were derived from a broad range of low to high (greenschist/blueschist to granulite) grade metamorphic rocks. The authors interpret the rutiles to be eroded from Pre-Cambrian gneissic terranes, and deposited in pre-Taconian orogeny sedimentary successions. During the Devonian, they were eroded from post-Grenville/pre-Taconian sedimentary or sub-greenschist grade metasedimentary rocks in the Acadian orogen.

Two recent studies of Devonian to Pennsylvanian sedimentary successions in the eastern U.S. (Thomas et al., 2004; Eriksson et al., 2004) hypothesize that synorogenic foreland basin sediments incorporate clastic detritus from the erosion of previous orogenic events, but not from the current orogeny. This appears to agree overall with the data and interpretations of McLennan et al. (2001), and the numerous petrologic studies of Middle to Upper Devonian strata in the Catskill Front (e.g., Burtner, 1964; Lucier, 1966; Fletcher, 1967; Gale, 1985). However, all of these

studies are in sharp contrast with the results of Aronson et al.'s (1994) detrital micas dates, and projected dominance of Acadian sources of sediment during deposition of the Catskill delta.

As pointed out by McLennan et al. (2001) these contrasting data and interpretations are in part a result of analysis of very different detrital grains (e.g., zircons versus white micas), which may have different deep versus shallow sources in an orogen, and be biased by weathering/abrasion and transport processes. It is intuitive that Acadian-age, non-volcanic zircons would be sourced from deeply buried igneous or metamorphic rocks, whereas Acadian-age white micas could be readily sourced from shallowly buried, low-grade metamorphosed rocks. Deep Acadian sources, exposed at present in New England, would not have been unroofed during the Devonian. In contrast, Upper Silurian to Middle Devonian synorogenic sediments, or Taconic metamorphics re-exposed to >350°C temperature conditions during the Acadian orogeny, would be less deeply buried, and become exposed and weathering in the orogen.

And, as also pointed out by McLennan et al. (2001), while highly durable zircons last through multiple events that recycle sediments, micas are readily weathered, or broken down by transport processes, and disappear relatively quickly. One would expect to find older zircons in clastic sediments, but less so older micas.

Another key issue in this debate may be the small number of analyses performed, on a stratigraphically- and regionally-limited number of samples. A more systematic geochemical and geochronologic analysis of the Lower to Upper Devonian succession in the Catskill Front (and other areas) is needed.

One more point should be expressed on this issue. Low grade metamorphic rocks are known to be the source of a significant component of Catskill delta sediments. While interpreted to be Taconian in age by earlier workers (e.g., Burtner, Lucier etc.), Aronson's dates constrain their source to be largely Acadian. Where would such Devonian sediments be sourced from?

The author provides the following hypothesis as a plausible explanation. Early in the Acadian orogeny, a massive volume of Upper Silurian to at least Lower Devonian synorogenic sediments were deposited in the foredeep basin in New England (e.g., Littleton Formation). A great thickness of these was deposited over the top of the rocks of the Taconian orogen in the basin foredeep. As the Acadian orogenic front migrated cratonward through time, these early foredeep sediments, and underlying rocks, were subjected to regional metamorphism. Later in the orogeny, these early Acadian synorogenics would have been uplifted along thrust sheets and exposed in the wedge-top of the foreland basin, and cannibalized, providing unaltered to low-grade metamorphosed Acadian sediments to younger Acadian synorogenic sediments. Although some older rocks may have been thrust to the surface, much of what should have been exposed in the orogenic front should have been the younger Upper Silurian to Lower Devonian rocks. This thick younger succession would largely have to be unroofed first to get to underlying Taconian-age rocks.

The author finds it most plausible that the fine-grained sediments, sedimentary and low-grade metamorphic rock fragments, and conglomerate clasts in the Catskill Front were derived from Acadian, not Taconian sources. At least in part, older more durable grains, like zircons and rutile, were likely eroded from pre-Devonian sources early in the orogen, and deposited in the Devonian succession of the early Acadian foredeep, and later uplifted and cannibalized.

In summary, the provenance of at least Middle to Upper Devonian sandstones and mudrocks in the Catskill Front appear to be derived largely from sedimentary and low grade metamorphic rocks in the Acadian orogen. Between approximately 388 and 376 million years ago (Mount Marion to Slide Mountain formations), there were no major changes in sediment composition, and hence no major changes in rock types exposed and eroding within the paleodrainage basin feeding into the Catskill Front. In contrast with previous interpretations, it seems plausible that Catskill delta clastics were largely derived from Acadian sources, not Taconian. Little if any significant magmatic or high-grade metamorphic rocks were exposed in the orogen, within the paleodrainage basin. A hidden source of abundant smectitic clays, derived from weathering of igneous rocks, may have come from airborne volcanic ash.

Foreland Basin Tephra and Acadian Volcanism/Magmatism. At this time, more than 80 beds of volcanic airfall origin are documented from Lower and Middle Devonian Lochkovian to Eifelian stages) strata across the Appalachian basin. Additional beds are known from the upper Middle to Upper Devonian, but more work is needed in that interval. Stepping back, however, what are the broader implications of this data for the history of Acadian paleo-

volcanism? Can the record of tephra beds preserved in foreland basin sediments be used as a proxy for the timing and character of explosive plinian volcanism in a magmatic belt like the Acadian orogen?

Tephra Beds as a Proxy for Paleovolcanism. The traditional view of volcanic airfall tephtras (including the New York's Devonian K-bentonites) is that a single tephra bed is the result of a single volcanic eruption. Recent studies, however, indicate that many beds have a complex depositional history, resulting from reworking of tephra sediments and/or the amalgamation of multiple eruptive events into a single layer. Furthermore, a broad range of physical, biological, and chemical processes active in individual environments can lead to preservation, mixing or destruction of airfall tephra layers.

Ver Straeten (2004a) explored these issues and their implications for explosive Lower to Middle Devonian volcanism in the Acadian orogen (see also Ver Straeten, 2005, 2007b, 2008; Benedict, 2004; and Ver Straeten et al., 2005). Based on the record of tephtras in the foreland basin fill, and recognizing potential preservational biases in that record, Ver Straeten (2004a) proposed that the mid-Lochkovian, lower Emsian and lower to middle Eifelian stages were times of increased volcanism in the mountain belt. These times correspond to deposition of the Bald Hill, Sprout Brook, and Tioga Middle Coarse Zone and Tioga A-G tephra clusters. Ongoing search of the literature on Devonian magmatism and volcanism from the Acadian orogen seems to support those interpretations (see lower Emsian discussion below).

As noted in the provenance discussion, other lines of evidence (smectite clay mineralogy and vertic paleosols) appear to indicate the presence of an otherwise disguised igneous source in the Devonian strata. This could be derived from weathered igneous rocks in the orogen, reworking of pre-Acadian sedimentary rocks in the orogen rich in smectite/igneous minerals, or weathered Acadian-derived volcanic ash deposited by airfall in the foreland basin

Lower Emsian magmatism and volcanism in the Acadian Orogen. Figure 14 appears to indicate that packages of multiple Sprout Brook K-bentonites can be physically correlated from the Catskill front to Helderbergs in eastern New York. This is dependent on the separation of a few closely-spaced K-bentonites from others by thicker beds of background siliceous siltstones to cherts. Thickest siltstone/chert beds appear to underlie the insertion of the sets of thin K-bentonites.

These patterns resemble small-scale parasequences, developed in clastic-dominated facies. Increased sedimentation of marine sediments during a fall to lowstand of sea level (small-scale falling systems and lowstand systems tracts in sequence stratigraphy terms) would yield thicker siliceous siltstones to cherts. Succeeding sea level rise (a small-scale transgressive systems tract) and resultant shutoff of clastic sedimentation would allow for accumulation of whatever alternative sediment was available in the environment. In this situation, that would be airborne volcanic tephra from a single to multiple volcanic eruptions in the orogenic belt. The accumulation of exclusively volcanic tephra could occur over sediment-starved intervals of time because no other sediment was available. A detailed discussion of tephra deposition and sediment condensation in a context of sea level change is provided in Ver Straeten (2008).

The lower part of the Esopus Formation (Spawn Hollow Member) marks a major sea level rise, via a combination of a third order eustatic sea level rise, and superposed tectonically-induced basin subsidence (Ver Straeten, 2007a; Figure 8). Shutoff of carbonate production, and relative clastic sediment starvation due to sea level rise could potentially result in periods of time where the only sediment available for deposition in the environment might be airfall volcanic ash (Brett and Baird, 1990; Puspoki et al., 2008; Ver Straeten, 2008).

Supporting evidence for application of such a model in lower Esopus time come from geochemical analysis of apatite phenocrysts from the Sprout Brook K-bentonite beds by Benedict (2004; and in Ver Straeten et al., 2005). His work indicates that a single tephra bed sometimes yields phenocrysts with different geochemical signatures. Those signatures imply deposition of ash from different volcanic sources, or at least different eruptive events, within a single bed.

Puspoki et al. (2008) carefully documented such a case of enhanced sedimentation of tephra from multiple eruptions with clastic sediment starvation during a major Miocene transgression. In their remarkable study, deposition during

some parasequences consisted wholly of volcanic ash sediments; the delineation of individual parasequences was in some cases only possible by the degree of weathering and alteration of volcanic to clays at transgressive surfaces.

Let's extend this line of thought further. If the patterns of a few clustered Sprout Brook tephra (separated by thicker beds of marine deposits) are interpreted to represent small-scale cycles, the sections at Becraft Mountain to Callanans Corners could possibly represent six such cycles. If each cycle is interpreted to represent a Milankovitch precessional cycle, of approximately 23,000 years duration, then the entire Sprout Brook succession could represent $23,000 \times 6 =$ approximately 138,000 years. If individual couplets of K-bentonite + background marine beds represent a single parasequence and the six packages represent approximately 100,000 year Milankovitch eccentricity cycles, then the succession would comprise on the order of 600,000 years.

Where is all of this volcanic ash coming from? Is there supporting data in the Acadian orogen to interpret such ongoing and extensive volcanism during deposition of the Sprout Brook K-bentonites?

Lower Emsian-age igneous rocks occur in both the greater Acadian foreland basin and the Acadian orogen. A number of these have an overlapping error range with the Sprout Brook K-bentonites cluster in the Appalachian basin (408.3 +/- 1.9 Ma; Tucker et al. 1998). These include at least six different deposits of volcanic rocks, and numerous plutons in New England, Quebec and New Brunswick (e.g., Bradley et al., 2000; Ver Straeten, 2004b).

The Sprout Brook K-bentonites, in the lower part of the Esopus Formation (Spawn Hollow Mbr.), are found in northeastern New York (Ulster and Otsego counties; Ver Straeten, 2004b). Locally, in the southern part of the basin, beds with mixed volcanogenic and detrital grains, are found (Conkin and Conkin, 1979; Ver Straeten 2004b).

A relatively large number of lower Emsian volcanic and magmatic rock units with an overlapping error range with the Sprout Brook K-bentonites (ca. 409 to 405 Ma) are reported from the Acadian orogen, chiefly from Maine (see references below). These include the Traveler, Kineo and other rhyolitic ashflow tuffs erupted from five major volcanic centers in north-central to western Maine. All along the belt, the rhyolites are underlain by quartz-rich sandstones (Matagamon Formation) equivalent to the Oriskany Formation in New York (Boucot, 1969; Rankin, 1968, Rankin and Hon, 1987). Furthermore, some of the rhyolites are overlain by marine sedimentary rocks correlative with the Schoharie Formation in New York (Boucot, 1969).

The widely known Mount Katahdin in north-central Maine is one of the lower Emsian granitic plutons. Katahdin's northeastern flank is draped by its co-magmatic ash flow tuffs (Traveler Rhyolite; Rankin, 1968). Rankin and Hon (1987) conservatively estimate the preserved volume of the Traveler Rhyolite alone is approximately 800 km³. This volume compares with major Cenozoic tuffs in the western U.S. (e.g., Timber Mountain, Paintbrush, and Lava Creek tuffs, 900 km³, 1000 km³ and 1000 km³, respectively; Christiansen, 1979, p. 31). The bottom and top of the Traveler Rhyolite has been dated at 407.3 +/-0.5 and 406.7 +/-1.4 Ma. The nearby Kineo Rhyolite yielded an age of 406.3 +/-3.8 Ma (Bradley et al., 2000). These volcanic rocks are all within the range of error for the Sprout Brook K-bentonites.

The ages of numerous additional lower Emsian-age felsic plutons in Maine and New Hampshire are reported by Bradley et al. (2000). Further plutonic rocks with dating errors that overlap with the Sprout Brook K-bentonites are reported by Tucker and Robinson (1990), Bevier and Whalen (1990), Rankin and Tucker (1995), Robinson and Tucker (1996), Ludman and Idleman (1998), Solar et al. (1998), Eusden et al. (2000), and Tucker et al. (2001).

More lower Emsian volcanogenic strata from Acadian foreland basin deposits on the Gaspé Peninsula, Quebec, in northern New Brunswick, and in New Hampshire, (Billings, 1937, 1956; Douglas, 1970; Boucot, 1970, pers. comm., 1993; Poole and Rogers, 1972; Doyon and Valiquette, 1987; Tucker and Rankin, in Bradley et al., 1999; and Wilson, 2004).

No igneous rocks of Emsian age are known from south of New England. This may be associated with erosion and/or cover by younger rocks. However, with close to no record of airfall tephra in the southern part of the Appalachian basin, it is possible that there was only minor explosive lower Emsian-age volcanic activity in the southern Acadian orogen.

The high concentration of Lower Emsian magmatic and volcanic rocks in the northern Appalachians, combined with the restricted distribution of airfall tephra largely in the northeast portion of the Appalachian basin suggest that the Sprout Brook K-bentonites originated from sources in New England, possibly even in part from the Traveler Rhyolite (Ver Straeten, 2004b). Devonian-age westward directed transport by winds would have carried tephra plumes from New England out over the northeastern edge of the Appalachian basin, where it would have been deposited in the seaway in present-day eastern New York.

Soft Sediment Deformation = Seismites? The purpose of this section is to examine potential causes of soft sediment deformation in the Catskill succession, and how they may potentially relate to Acadian orogenesis. Too little detailed information is yet available to make clear interpretations; it is hoped this will draw attention to such features, and lead to further investigation.

The deformation of unlithified sediments may result from various processes. These include the formation of a density inversion via rapid deposition of dense sediments over dilute water-rich sediments; repacking of under-compacted sediment layers; the escape of gases from sediments; sliding or slumping of sediments; impacts or earthquake/seismic shocks; waves or flood surges; tsunamis and density flows; and pressure changes on the sea floor from storm currents (Jones and Omoto, 2000; McLaughlin and Brett 2004; Montenat et al., 2005). They commonly occur (though not exclusively) in beds of alternating grain size (e.g., muds and sands or carbonates), which are sensitive to changes in sediment yield strength (Montenat et al., 2005). Sediments by themselves do not deform without a trigger to reduce their yield strength. According to Jones and Omoto (2000), deformation of unconsolidated sediment requires: 1) a deformation mechanism, which enables the material to be deformed; 2) a driving force, which brings about deformation; and 3) a trigger, which initiates 1, 2, or both.

In recent years, some SSDs have been interpreted to be the result of seismic shocks, generated by earthquakes. Termed “seismites”, they form due to powerful shocks that strongly affect water-saturated sediments, triggering a thixotropic reaction, which leads to liquefaction of sediments. Deformation then results from the expulsion and or intrusion of fluidized materials (Montenat et al., 2005). Selected references on seismites include Sims (1975), Pope et al. (1997), Jones and Omoto (2000) McLaughlin and Brett (2004), Montenat et al. (2007).

Key features of seismitic beds include convolute bedding/laminations; ball and pillow/“thixotropic bowls”/“saucer” structures; mudstone/sandstone volcanoes/diapirs; boudins and brecciated fabrics; inclined blocks; and truncation surfaces (McLaughlin and Brett 2004; Montenat et al., 2005). Sims (1975) discussed criteria for correlating soft sediment deformation structures with seismic events. These include: 1) Proximity to active seismic zones; 2) presence of potentially liquefiable sediments; 3) similarity to structures formed experimentally; 4) structures related to liquefaction; 5) structures restricted to single stratigraphic intervals correlatable over large areas; and 6) absence of slope influence and failure. When interpreting ancient SSDs, criteria number 1 is often difficult to assess; geologists rely more on the other criteria, including their correlatability over large areas.

The SSDs in the Mount Marion and Oneonta formations in the Catskill front (Figure 15) meet many of the criteria for seismites. Their internal structures appear to be related to liquefaction of the sediments, and they feature structures (e.g., ball & pillows) similar to experimentally-generated forms. Furthermore, they are largely restricted to relatively thin intervals, separated from other occurrences in similar facies, and adjacent under- and overlying beds appear undeformed.

The sedimentary conditions needed to commonly form SSDs in shallow marine and terrestrial facies (e.g., fluid muds below rapidly deposited sands) occur extensively, especially in the upper Mount Marion Formation. However, deformed zones are generally uncommon. As noted by McLaughlin and Brett (2004) this pattern is indicative of a limited frequency of triggering events capable of deforming the strata.

Some SSDs in both marine and terrestrial strata in the area occur within sandstone-only layers (e.g., Figure 15b, f, g; Stops 4 and 7), where the sands did not founder into fluid-rich muds. This would seem to call for a significant triggering event to initiate liquefaction, settling and tighter repacking of the sand grains, along with dewatering. Powerful seismic shocks, generated by along the Acadian deformation front, would provide a likely trigger to deform these sediments.

One key line of evidence for interpreting the Catskill SSDs as seismites, however, is still undocumented. It is unknown whether individual SSDs can be correlated from outcrop to outcrop, across broad areas. And at least in the upper part of the Mount Marion Formation, the author's experience indicates that it may be difficult to establish such correlations, due in part to the homogeneity of facies, an apparent lack of distinctive marker beds, the thickness of the interval, and its limited exposure.

Although the geographic distribution of these soft-sediment deformational units are unknown, others in the Devonian succession of New York have relatively widespread distribution. Sutton and Lewis (1966) report soft-sediment deformation in the Upper Devonian (Frasnian) Point Bluff Siltstone Bed in western New York. They found SSD at all studied localities over a ca. > 775 km² area of outcrop of the thin (ca. 12 cm-thick) unit. This is very likely a previously unreported seismite related to Acadian earthquake activity, either in the orogenic belt or along faults active at that time in western New York. In another study, Smith and Jacobi (1998) document soft sediment deformation in Upper Devonian (Famennian) strata of the Canadaway Group, and interpret them to be seismites related to syn-depositional movement along the Clarendon-Linden fault system.

The stratigraphic distribution of the SSDs in the Catskill Front present an additional possible line of support for their interpretation as seismites. The examples from the Mount Marion and lower Oneonta formations occur in strata associated with the early stages of Acadian tectophases, as outlined by Etensohn (1985). If the tectophase model is viable, these strata should represent times of increased seismic activity, with renewed or at least increased uplift and deformation in the orogenic belt.

In summary, SSDs in the Catskill Front potentially represent seismites, formed in unlithified sediments. If they are seismites, they provide insights into the timing of Acadian seismic activity. More work is needed documenting their character, distribution and, importantly, their correlatability across the region.

SUMMARY

This paper has been an attempt to provide a broad perspective of the Acadian orogeny, based on a synthesis of new and old data from New York's portion of the Acadian foreland basin. Approximately 2.7 km of Upper Silurian and Devonian strata in the Catskill front, deposited through roughly 45 million years (ca. 420 – 375 Ma) provide a key source of data about the timing and character of orogenic and foreland basin evolution, erosional unroofing the orogen, explosive volcanism, and perhaps seismic activity.

Beginning in the Late Silurian, collision of the North American margin and another landmass initiated the Acadian orogeny in the northeastern U.S. Uplift and loading in the orogen led to subsidence and development of an adjacent foreland basin system (wedge-top, foredeep, forebulge and back-bulge basin). A tremendous volume of sediments eroding off the orogen was deposited across the foreland.

Initially, the orogen and foreland basin were largely developed in New England. Through time, both the orogen and foreland migrated cratonward. As a result, initial foreland basin deposits were thrust up and exposed, some of them subjected to low-grade metamorphism, and then eroded and transported via rivers to cratonward.

Following the Taconian orogeny, the area of today's Catskill front was elevated and eroding from the Late Ordovician through much of the Silurian. However, in the Late Silurian, approximately 420 million years ago, the eastern New York high subsided and was submerged. Through deposition of the Rondout Formation and Helderberg Group, eastern New York was positioned in the distal, cratonward margin of the foreland basin system, in a back-bulge basin. Following Helderberg time, the forebulge region of the foreland migrated into and through the Catskill front; however, Oriskany time was an interval of relative quiescence in the mountain belt, and the forebulge was relatively subdued as it moved through the region. With the onset of a new tectonically active phase of the orogeny in later parts of the Early Devonian, the cratonward margin of the basin foredeep migrated into eastern New York, and the first major wedge of synorogenic clastics were deposited (Esopus Shale), followed by a gradational return to carbonate deposition during another period of relative tectonic quiescence (Schoharie to Onondaga formations).

The onset of another tectonically active phase of the Acadian orogeny (ca. 390 Ma) led to subsidence and black shale deposition (lower Hamilton Group, Union Springs Formation), and cratonward migration of the complete fore-deep of the foreland basin system into the Catskill Front. Within a few million years, the foredeep became overfilled to above sea level with mud and sand, and the Catskill front became terrestrial. The beginning of a third major tectonically active phase in the orogen is less visible in the Catskill Front, but is recorded in the marine basin at about 385 Ma, with deposition of the black Genesee Shale.

In New York, we have insufficient data to interpret much of the older history of the Acadian orogen. But beginning in the middle Lower Devonian, we see that three tectonically active to quiescent phases (Tectophases I-III of Etensohn, 1985) began at about 408, 390, and 385 Ma.

The sediments eroding off the orogen and being shed into the Catskill front through the Middle to Upper Devonian indicate that rocks exposed in the source area were predominantly low grade metamorphic (up to greenschist-grade) and sedimentary rocks. They were most likely derived from older Devonian foreland basin sediments, not Cambrian-Ordovician as proposed in older studies.

The foreland basin record of volcanic airfall tephra appears to indicate that there were peaks of explosive volcanic activity in the Acadian orogen at approximately 417, 408, and 391-390 Ma. Interestingly, these appear to correspond with the beginnings of tectonically active stages in the mountain belt.

ACKNOWLEDGEMENTS

The author would like to thank many people for sharing their knowledge and insights during the author's work on these Devonian problems. These include: C. Brett, G. Baird, A.J. Boucot, J. Dennison, J.B. Epstein, J. Inners, D.J. Over, D. Woodrow, R. Kirchgasser, M. DeSantis, A. Bartholomew, E. Landing, J. Ebert, R. Lindemann, D. Griffing, M. House, W.F. Koch, G. Klapper, the late W.A. Oliver Jr., and E. Schindler, on New York and Appalachian basin sedimentology and stratigraphy; P. DeCelles and P. McLaughlin on foreland basins in general; F. Etensohn, C.E. Brett, and A.J. Boucot, on foreland basin perspectives on Acadian orogenesis; D. Bradley, C. van Staal, P. Karabinos, D. Rankin and others on the Acadian orogeny; and W. Huff, G. Shaw, L. Benedict, J. Dennison, J. Delano, B. Hanson, D. Rankin, R. Smith III, D. Bradley, and A.J. Boucot on volcanic airfall tephra and Acadian volcanic and plutonic rocks.

FIELD TRIP ROAD LOG

Mileage

- 0.0 Start at Plattekill Parking Lot, on the SW corner of Plattekill Ave. and Manheim Ave. (NE corner of the SUNY-New Paltz campus).
Turn right onto Plattekill Ave.
- 0.0 Immediately, at stop sign, turn left onto Manheim Blvd.
- 0.2 Turn right at light onto Rtes. 32 & 299
- 1.0 Get into right lane.
- 1.1 Turn right at entrance to NY State Thruway.
- 1.3 Get toll card at booth.
- 1.4 Fork right for I-87/NYS Thruway Northbound (toward Albany).
- 5.9-7.3 Outcrops of Martinsburg Formation. Interbedded turbidites and shales.
- 6.4 Catskills visible ahead.
- 7.3 Cross over Wallkill River.
- 9.9 Cross over Rondout Creek. Begin climb up onto Devonian bedrock.
- 11.1 Lower Devonian Glenerie Formation.
- 11.8 Lower Devonian New Scotland Formation?
- 12.5 Lower Devonian, upper part of New Scotland to Becraft formations.
- 13.0 New Scotland Formation.
- 13.4 New Scotland to Becraft formations.
- 14.8-15.6 Lower Devonian Schoharie Formation
- 15.7 Middle Devonian Onondaga Limestone (Edgecliff Member).
- 16.0 Onondaga Limestone (Nedrow to Moorehouse members).
- 16.8 Cross over Esopus Creek.
- 17.0 Bypass Kingston exit.
- 17.6-18.3 Middle Devonian Union Springs Formation (Stony Hollow Member).
- 19.0 Esopus Creek on right.
- 19.5 Cross under Rte. 209, with a classic set of Lower to Middle Devonian road cuts along it.
- 19.8 Cross over Sawkill Creek.
- 19.9 Middle Devonian Mount Marion Formation (East Berne Member).
- 22.0 Mount Marion ahead on left. Type section of Mount Marion Formation.
- 22.7 Cross over Plattekill Creek.
- 23.1 Stony Hollow Member again.
- 25.3 Mount Marion to left.
- 26.2 Note large quarry on north flank of Mount Marion. East Berne and Otsego members, Mount Marion Formation.
- 27.1 Bypass Saugerties exit. Most complete section of Onondaga Limestone in eastern New York, visible along Thruway and northbound entrance; Schoharie Formation visible at far end, right side.
- 28.5-29.1 Excellent section of Lower Devonian Becraft, Alsen, Port Ewen, and Glenerie formations.
- 29.8 Schoharie Formation (Aquetuck & Saugerties members).
- 32.6 Schoharie Formation.
- 33.2 Extensive limestone quarries in Lower Devonian Helderberg Group beyond ridge to right.
- 34.6 Becraft Formation (?).
- 34.5 Schoharie Formation.
- 35.3 Schoharie Formation +/- uppermost underlying Esopus Formation.
- 35.5 South end of long exposure along abandoned Thruway exit, ending at Stop 2 of this trip. Strata visible from Thruway include Schoharie and Esopus formations. Complete section of Esopus exposed.
- 36.2 Onondaga to Schoharie formations.
- 37.0-37.2 Schoharie exposed on southbound side of Thruway.

- 36.8 Onondaga Formation.
- 37.4 Cross over Kaaterskill Creek.
- 37.6 Schoharie Formation on west side, Onondaga Formation on east side.
- 38.1 Schoharie Formation.
- 38.3-39.5 Esopus Formation (mostly Quarry Hill Member, in middle of formation).
- 39.6 Cross over Catskill Creek.
- 39.7 Becraft to New Scotland formations.
- 40.0 Exit NYS Thruway at Catskill. New Scotland Formation along exit ramp.
- 40.5 Pay toll. New Scotland Formation on left.
- 40.7 Turn left onto Rte. 23b, toward Catskill.
- 40.8 New Scotland Formation on left, Kalkberg Formation on right.
- 41.0 Pull over and park on shoulder. Cross and continue ahead on Rte. 23b, then walk up exit ramp off of Rte. 23 to prominent angular Taconic unconformity.

STOP 1. TACONIC UNCONFORMITY AND HELDERBERG GROUP LIMESTONES, CATSKILL (45 MINUTES). This classic locality has been the subject of many field trips. For this trip, we will visit two road cuts along Rte. 23: a) exit ramp to Leeds/Rte. 23b, off of Rte. 23 westbound; and b) the succession on the north side of Rte. 23 west of Rte. 23b. We will largely focus on various features with implications for Acadian orogenesis, as interpreted from the sedimentary rock record. These include the Taconic unconformity and Manlius-Coeymans contact, carbonate deposition in a back-bulge basin of the foreland basin system, volcanic tephra beds (K-bentonites), and possible Devonian deformation of the “Little Mountains Fold-Thrust Belt” here.

Near the base of the succession, the prominent angular unconformity (Taconic unconformity) places supratidal dolostones and sandstones (Upper Silurian (?) Rondout Formation) over deep water turbidite sandstones and shales (Middle Ordovician Austin Glen Formation). The hiatus represents approximately 30 million years of time, and marks a series of events associated with the Taconian orogeny. This history begins with deep water deposition of synorogenic clastics, overthrusting of the rocks from western Massachusetts into eastern New York, and later uplift of the area due to erosional unroofing and rebound of the Taconian orogen through the latest Ordovician to Late Silurian. The latter resulted in the elevation of eastern New York above sea level throughout most of the Silurian, and restriction of the sea to western +/- central New York during most of the Silurian. Transgression of marine waters over eastern New York in the latest Silurian to earliest Devonian is at least in part due to tectonic-induced subsidence and migration of the foreland basin to the east. This was associated with collision and crustal loading during Late Silurian, with the onset of the Acadian orogeny on the far margin of eastern North America.

The limestones, dolostones and minor shales of the Rondout Formation and Helderberg Group along Rte. 23 here comprise a carbonate ramp succession. They were deposited in the back-bulge basin of the greater Acadian foreland basin, at a time when the orogen and the main body of the foreland basin were still far to the east. The strata examined on this trip (Rondout, Manlius, Coeymans, Kalkberg, New Scotland, and Becraft formations) record an overall deepening- to shallowing up succession through supratidal, tidal, shoal, and shallow to deep ramp facies (Rondout to lower New Scotland fms.), followed by a gradational shallowing up to shoal to tidal facies (middle New Scotland to lower Becraft formations). In sequence stratigraphic terms, the succession comprises a single “third order” depositional sequence, and the lowstand base of a second sequence. In total, the Rondout-Helderberg succession represents two major “third order” sea level cycles/depositional sequences.

The Manlius-Coeymans contact in the Catskill area was interpreted by Chadwick (1944) to represent an erosional hiatus. However, the wider distribution and implications of this break has only been recently documented by Ebert and Matteson (2003). Their detailed work through the two formations has documented at least two significant unconformities, including the formational contact. Noting a subtle but documented angularity to the unconformity, Ebert and Matteson (2003) interpret its formation to be associated with the orogen-ward migration of a bulge-like feature, associated with early stages of the Acadian orogen on the distant margin of North America near the Silurian-Devonian boundary.

Ancient volcanic airfall tephra beds, altered to clay-rich K-bentonites, occur in the Kalkberg and New Scotland formations along Rte. 23. This cluster of approximately 15 beds, termed the Bald Hill K-bentonites, are widely reported across the Appalachian basin, from eastern New York to Virginia and West Virginia. They preserve a record of explosive, plinian-type volcanism in the Acadian orogen. Additional K-bentonites have been noted in the Manlius Formation at other localities (Ebert and Matteson, 2003; P. Rubin, pers. commun. 2007).

The age of deformation of the strata along Rte. 23 is the subject of debate. Some workers (e.g., Marshak, 1986; Marshak and Tabor, 1989; Zadins, 1989) interpreted the folding and faulting to be Devonian, at least in part. In contrast, Geiser and Engelder (1983) interpret the structures to have formed later, during the Late Carboniferous-Permian Alleghanian orogeny. It is possible that both orogenic events led to the deformation. At present, the timing of deformation is unclear.

- At end of Stop 1a, return to cars.
- 41.3 Proceed ahead on Rte. 23.
- 41.4 Cross over NYS Thruway.
- 41.5 Beginning of another classic cut through the “Little Mountains” fold-thrust belt.
- 41.8 Cross over Catskill Creek. Downstream the creek passes for approximately 1 mile through the gorge of Austin Glen, descending stratigraphically through Lower Devonian to uppermost Silurian strata of the Esopus through Rondout formations, and into the Ordovician Austin Glen Formation. Classic, beautiful site.
- 41.9 Esopus through basal Onondaga Formations (Stop 1B of Ver Straeten and Brett, 1995).
- 42.4 Turn left at light, onto Cauterskill Road.
- 43.7 Cross over NYS Thruway.
- 43.8 Schoharie Formation.
- 44.1 Onondaga Formation.
- 45.0 Cross over Kaaterskill Creek. Ordovician Austin Glen Formation exposed in creek bed to left.
- 45.05 Turn right at stop sign, continuing on Cauterskill Road. Kaaterskill Creek will follow road for some distance.
- 46.1-.6 Onondaga Limestone. Section between 46.5-.6 exposes Edgecliff, Nedrow and lower Moorehouse members.
- 46.9 Fork left onto Rte. 23a.
- 47.1 Park along shoulder, and walk ahead to prominent outcrop.

STOP 2. LOWER DEVONIAN RAMP CARBONATES AND THE INITIAL ACADIAN CLASTIC WEDGE, THE SPROUT BROOK K-BENTONITES, AND THE WALLBRIDGE UNCONFORMITY (45 MINUTES).

This is another classic Devonian locality, chiefly noted for its prominent structural folds. However, several additional characters of the outcrop provide a record Acadian Lower Devonian activity in the Acadian mountain belt. These include the Wallbridge unconformity; a carbonate-quartz arenite suite of rocks succeeded by the first major influx of Acadian synorogenic sediments; migration of the foreland basin foredeep into the Catskill Front; and a second major cluster of altered volcanic tephra beds. The Glenerie-Esopus contact at this outcrop lies approximately 90 m stratigraphically above the Taconic unconformity, seen at Stop 1.

The units visible at Stop 2, from low to high include the top of the Port Ewen Formation, the Wallbridge unconformity, local chert facies of the Glenerie, and a rare, complete section of the overlying Esopus Formation. At the far end of the outcrop, along the NYS Thruway, lower strata of the Schoharie Formation are visible.

The top of the Port Ewen Formation, a shaly limestone analogous to the New Scotland formation, directly underlies the Wallbridge unconformity here. The Wallbridge, which marks one of the major Phanerozoic sea level lowstands in North America (Sloss, 1963), is of relatively short duration in the Hudson Valley. Deposition across the interval is continuous in the Port Jervis area, where New York, New Jersey, and Pennsylvania meet.

Immediately overlying the Wallbridge unconformity is a conglomeratic lag bed at the base of the Glenerie Formation. The conglomerate in the area of Catskill is largely composed of phosphatic pebbles with scattered milky quartz. To the south, beginning near Kingston, a conglomerate unit of milky quartz wedges in below the Glenerie. This unit,

termed the Connelly Conglomerate, is the oldest conglomerate found in the New York Devonian (Table 2). Its quartz composition (>99% milky quartz) contrasts with younger pebbly to conglomeratic strata in the Catskill front, which feature a more diverse composition (Table 2). The conglomeratic base of the Glenerie locally near Catskill is dominated by chert; to the south, it transitions into silica-rich limestones, and to the north to the quartz sand-rich Oriskany Sandstone.

Glenerie-Esopus contact is relatively gradational at Stop 2 and in the eastern New York region. Elsewhere across the central to southern part of the basin, the contact is generally more sharp (Ver Straeten, 2007a). It marks a time of foundering of widespread shallow marine ramp conditions, and progressive subsidence and migration of the foreland basin foredeep into eastern New York, during the onset of the first Acadian Tectophase recognized by Ettensohn (1985). In reality, at least one previous tectonically active to quiescent “tectophase” likely occurred during the Late Silurian to Early Devonian parts of the Acadian orogeny.

The mudstones, shales, siltstones and sandstones of the Esopus Formation mark the first significant influx of Acadian synorogenic clastics into the Catskill Front and the Appalachian basin. Migration of the orogenic front across New England has by the Emsian Stage moved far enough cratonward for the foredeep segment of the foreland basin system to migrate west of the present day Hudson River. Petrologic data from Rehmer (1976) indicates that fine sandstones to siltstones of the Esopus are rich in quartz (~33-55%), with a high concentration of matrix (~19-49%), and minor amounts of fragmentary mica and chlorite, detrital and diagenetic chert, and pyrite.

Detailed work by the author on Emsian-age (upper Lower Devonian) strata of the Esopus and Schoharie formations and equivalent strata across the Appalachian basin (Ver Straeten, 2007a), along with geochronologic age dating (Tucker et al., 1998; Kaufmann, 2006), outline a longer, more complex history to the Emsian stage in the eastern U.S. than has been appreciated. U-Pb dating indicates a duration on the order of 15-18 million years for the Emsian (Tucker et al., 1998; Kaufmann, 2006), and that its global sea level history comprises five major third order cycles (Ver Straeten, in press), not one as previously interpreted by Johnson et al. (1985).

To the author, one of the key stories linking the foreland and orogenic belt is tied a series of thin, tan-colored clay beds in the lower part of the Esopus Formation here (Spawn Hollow Member). These are the Sprout Brook K-bentonites, 15 altered volcanic tephra dated at 408.1 +/- 1.5 Ma (Tucker et al., 1998). In contrast with other Devonian clusters of K-bentonites in the Appalachian basin, these are geographically restricted to eastern New York (Ver Straeten, 2004a, b). The age of the Sprout Brook K-bentonites cluster overlaps with dates of numerous volcanic and plutonic rocks in the northern Appalachians, predominantly in Maine (e.g., Bradley et al., 2000; and additional references in main body of this paper). This includes the Katahdin Granite and co-magmatic Traveler Rhyolite of central Maine; the Traveler alone, from only one of many volcanic centers of lower Emsian age, is conservatively estimated to have a volume on the order of the largest Cenozoic-age tuffs in the western U.S. (Rankin and Hon, 1987; Ver Straeten, 2004b). It is plausible that explosive, plinian-type volcanoes in northern New England were the source of the tephra deposited in eastern New York.

As at Stop 1, the age of deformation of the strata here is unknown. An interesting point here is the relationship of deformation to the Sprout Brook K-bentonites. The soft, unlithified clay beds form a sharp rheological contrast with their interbedded thin chert and shale beds. Slippage largely appears to follow the clay beds, which in places appear to have been squeezed through the folds (“like toothpaste”). This is well seen in the uppermost K-bentonite, which varies in thickness from zero to over 1.5 meters along the outcrop over the central anticline. The K-bentonites probably helped concentrate deformation along this zone in the Catskill Front and into the subsurface, the position of a major decollement in the Catskill Front/Little Mountains fold-thrust belt according to Marshak (1986).

At end of Stop 2, return to cars.

- 47.1 To continue, pull ahead and turn around near base of outcrop. Proceed west on Rte. 23a.
- 47.3 Pass Cauterskill Road.
- 47.4 Cross over NYS Thruway. Prominent exposures of Lower Devonian Schoharie Formation.
- 47.6 Pass Old Kings Highway on left. Topmost Schoharie and Edgecliff Member of Middle Devonian Onondaga Limestone exposed along beginning of Road (Stop 3B of Ver Straeten and Brett, 1995).

- 47.65 PARKING OPTION 1 FOR STOP 3: Pull onto shoulder and park. *If you wish to use option 2 for parking, proceed ahead on Rte. 23.* Folded Nedrow and Moorehouse members (Onondaga Limestone) exposed on left. To proceed from here to Stop 3, carefully walk 0.3 miles ahead to east (proximal) side of bridge over Kaaterskill Creek. Walk down steep slope on south (left) side of bridge to exposure along creek.
- 47.95 Cross over Kaaterskill Creek. Drive across valley of easily eroded Bakoven Member black shales.
- 48.45 Turn around at intersection of Rte. 23 and Underhill Rd. Then return east on Rte. 23
- 48.6 PARKING OPTION 2 FOR STOP 3: Pull onto shoulder and park in grassy area. To proceed from here to Stop 3, carefully walk 0.3 miles ahead to the far side of bridge over Kaaterskill Creek. Walk down steep slope on south (left) side of bridge to exposure along creek.

STOP 3. MIDDLE DEVONIAN RAMP CARBONATES AND THE SECOND ACADIAN CLASTIC WEDGE, AND TEPHRAS (30 MINUTES). Note: No hammers or collecting from the Onondaga-Bakoven contact – it is a rare exposure. The creek exposure is on private property - ask permission for access.

Strata exposed at Stop 3 include the uppermost limestones of the Middle Devonian Onondaga Formation and overlying black shales of the Bakoven Member of the Union Springs Formation (lower part of the Marcellus subgroup of Ver Straeten and Brett, 2006). At least one K-bentonite of the Tioga A-G K-bentonite cluster is exposed here. The first part of the outcrop is along Kings Highway and Rte. 23a, where the lower to middle Onondaga Formation is exposed (Edgecliff, Nedrow and Moorehouse members). Walk westward down the hill to the south side of the bridge over Kaaterskill Creek, where the top Onondaga (Seneca Member?) and lower Bakoven Shale are exposed along the creek. This outcrop lies approximately 270 m stratigraphically above the Taconic unconformity, seen at Stop 1.

Above the Esopus Formation at the previous stop, the Schoharie Formation is transitional from extrabasinal clastics to intrabasinal carbonates. This shift culminates in deposition of the Middle Devonian Onondaga Limestone. Onondaga-equivalent carbonate-rich facies occur widely across eastern North America; they mark a shutdown of clastic sedimentation in the northern Appalachian basin, although to the south in deeper parts of the basin (PA through VA-WV), the interval marks more of a decline in % clastic content, and deposition of mixed carbonate-clastic facies.

Base of Onondaga is a lowstand of sea level. In the Schunnemunk outlier north of NYC, and in the area of Palmer-ton, eastern PA, lower Onondaga +/- upper Schoharie strata are represented by shallow marine quartz sandstones conglomerates. Exclusively quartz pebbles also occur scattered through strata at two positions within the Schoharie Formation. Furthermore, rare quartz pebbles have also been found in the same position in the Moorehouse Member and equivalent strata in central New York and central Pennsylvania. Curiously, widespread occurrence of quartz-rich strata are a contrast with the synorogenic sediments of the Esopus Shale below. Conditions that permitted progradation of Esopus extrabasinal clastics no longer existed.

Here at Stop 3 we stand at a major depositional shift, where relatively shallow marine carbonates are succeeded by basinal organic-rich black shales (by some estimates representative of ca. 100-200 m depth). A prominent bone-rich phosphatic lag interval at the formational contact marks a period of sediment starvation across the transition. A similar lag bed occurs along the Onondaga-Bakoven contact into western New York, overlying progressively younger uppermost Onondaga strata in that direction. The top of the Onondaga progressively youngs to the west, indicated but by the progressive upward appearance of the Tioga B, B', C, D, E, and F K-bentonites with the thickening of upper Onondaga (Seneca Member) strata to the west. This younging is associated with earlier foundering and subsidence of the foredeep of the foreland basin in eastern New York; subsequently, the shallow Onondaga ramp progressively subsided to the west, due to loading and uplift during the onset of Acadian Tectophase II of Etensohn (1985).

The Onondaga to lower Union Springs Formation and equivalent strata across eastern North America are well known for the occurrence of the Tioga K-bentonites. Actually, the so-called "Tioga" interval comprises two major clusters of volcanic tephtras (Ver Straeten, 2004a, 2007a). In the northern and central part of the Appalachian basin, an upper cluster of eight beds (Tioga A-G K-bentonites) occurs widely in upper Onondaga and Union Springs-equi-

valent strata. A lower cluster of up to 32 beds, previously correlated with the Tioga A-G K-bentonites, is found only in middle Onondaga-equivalent strata in the southern part of the Appalachian basin (Virginia and West Virginia; Ver Straeten, 2004a, 2007a). The lower cluster has been dated at 391.4 \pm 1.8 Ma (Tucker et al., 1998), the upper one at 390 \pm 0.5 Ma (Roden et al., 1990). They appear to be sourced from volcanic centers in northern Virginia and south-east of the Stroudsburg, PA area, respectively (Dennison and Textoris, 1970, 1978; Ver Straeten, 2004a). At Stop 3, some of the long bedding planes exposed along Rte. 23a may represent thin K-bentonites. Along the creek exposure, a thin centimeter-thick clay bed 35 cm below the contact is a K-bentonite. Another thicker bed, possibly the widely known Tioga B K-bentonite bed from the base of the Seneca Member in New York, is covered just above the base of a small gully closer to the bridge.

At end of Stop 3, return to cars. *If you used the first option for parking, follow directions to Underhill Rd., turn around and follow directions from 48.6 miles.*

- 48.6 Proceed ahead, eastward, on Rte. 23a.
- 49.4 Cross over NYS Thruway again.
- 49.5 Fork left onto Cauterskill Road, then immediately turn left again at stop sign.
- 50.1 Kaaterskill creek visible along road again, on left.
- 51.4 Turn left at bridge, and remain on Cauterskill Road.
- 52.6 Cross over NYS Thruway.
- 53.5 Pass Vedder Mountain Road on left
- 53.55 Turn left onto Vedder Road.
- 54.4 Exposures of upper part of the Mount Marion Formation on left.
- 54.6 Pull onto shoulder of Vedder Road and park.

STOP 4. NEARSHORE MARINE CLASTICS, UPPER MOUNT MARION FM. (30 MINUTES). This stop examines shallow marine clastics (upper part, Middle Devonian Mount Marion Fm.), not far below the transition into terrestrial strata (Ashokan Fm. and higher strata). Exposed along Rte. 23 are approximately 12.3 m of sandstone-dominated strata. Well defined hummocky cross beds, indicative of storm processes low in the outcrop are replaced above by mega-ripples (dunes), suggesting an overall shallowing up succession. Intraformational conglomerates and soft sediment deformation zones are also visible in the upper part of the outcrop. The strata at Stop 4 are approximately 750 m stratigraphically above the Taconic unconformity at Stop 1.

Two sandstone beds with reworked, intraformational pebbles and brachiopods are visible in the upper part of the outcrop. Along Catskill Creek, approximately 0.8 km to the north-northwest, a conglomerate with abundant macrocrystalline milky quartz and chert, with lesser numbers of sandstone and other clasts, was reported by Wolff (1967) and exposed in the late 1980s. It is presently covered in the creek bed. The author has noted multiple thin, sometimes lensing conglomerate beds in the upper Mount Marion Formation along and in the forest off of Rtes. 28 and 28a northwest of Kingston. They also mark the progradation of gravels into the Catskill Front at approximately 388 Ma, possibly concentrated into beds during the lowstand or basal transgressions of small scale (fifth to sixth order) cycles (e.g., Bergman and Walker, 1987; Smith and Jacobi, 1998).

Compositionally (Table 2), these conglomerates contrast sharply with the only white quartz pebble compositions of conglomerates and scattered pebbles in older strata (Connelly Formation; Schoharie Formation; Kanouse, Palmerton and Onondaga/upper Needmore formations) in New York and Pennsylvania. The relatively high concentration of various-colored cherts in the conglomerates also contrast with relatively chert poor compositions of overlying conglomerates in the Devonian succession. This indicates a relatively significant source of chert in rocks exposed in the Acadian orogen at this time. The upper Mount Marion conglomerates

Smith's (1970) petrographic studies of upper Mount Marion sandstones in the Catskill front found that the sands were composed of a mix of mono- and polycrystalline quartz, chert, metamorphic and sedimentary rock fragments with less abundant chlorite, other micas, and plagioclase feldspars. His reported average compositions for the upper Mount Marion strata ("Solsville" and overlying "Pecksport" equivalents, 5 and 8 samples, respectively) are: Quartz + chert = 50.0 & 37.4%; matrix = 9.1 & 13.2%; rock fragments = 37.6 & 41.1%; and carbonate = 1.6 & 8.3%. The

presence of slate and phyllite fragments, along with recycled sandstone and limestone grains indicated erosion of low grade metamorphic and sedimentary rocks in the Acadian orogen by approximately 388 Ma.

On the south side of the outcrop, along Vedder Road, two zones of soft sediment deformation can be seen in the upper part of the outcrops (Figure 15b). These potentially represent “seismite” beds, formed when severe seismic shocks from the Acadian orogen triggered liquification of the loosely packed sands, and their subsequent, more condensed repacking, and expulsion of excess water.

The first outcrop west of Five Mile Woods Road, on the south side of Rte. 23, exposes roughly five meters of interbedded sandstone and dark gray mudstones. Two sandstone bodies (ca. 1.5 and 3 m-thick) are not notably cross-bedded or erosively based; small delicate traces in the intervening mudstones may represent small plant root traces. The outcrop may be the lowest terrestrial deposits exposed along Rte. 23. If so, it could represent a thin tongue of terrestrial facies within the upper Mount Marion Formation, as seen near Kingston (Stop 8 of Ver Straeten and Brett, 1995), or deposits low in the overlying Ashokan Formation.

Strata from the lowest redbeds to the top of Slide Mountain, the highest peak of the Catskills, comprise approximately 1.9 km thickness of Middle to Upper Devonian strata (lower Givetian to upper Frasnian stages, Rickard, 1975). They were almost exclusively deposited in fluvial-dominated, terrestrial environments. These strata will be seen at subsequent Stops 5-7, and in road cuts along the route.

At end of Stop 4, return to cars.

- 54.6 Proceed ahead on Vedder Road.
- 54.65 Turn right onto Five Mile Woods Road, then turn left onto Rte. 23 (westbound).
- 55.0 Interbedded sandstones and shales on south side of Rte. 23.
- 56.2 Lowest exposure of terrestrial “redbeds” of Plattekill Formation along Rte. 23. Lower red and green mudrocks (including paleosols) deposited on floodplain, overlain further along road by channel sandstones. Outcrops for next ~10 miles (to ~66.1 miles) are in the Plattekill Formation.
- 57.8 Intersections with Silver Spur Road. Good exposure of Plattekill Formation along Silver Spur Road to left.
- 59.2 Intersection with Rte. 32, which joins Rte. 23 here for a short distance. **For a “pit stop” at McDonalds, turn left onto Rte. 32 south, and then right into parking lot.**
- 60.0 Very good exposures of Plattekill Formation. Pull onto shoulder and park for optional stop.

(OPTIONAL) STOP 5. MIDDLE DEVONIAN FLUVIAL CHANNEL AND FLOODPLAIN DEPOSITS (30 MINUTES). This outcrop of the Middle Devonian Plattekill Formation exposes typical terrestrial facies of the Catskill delta complex, as developed in the Catskill front. The strata at Stop 5 are roughly a little less than one kilometer stratigraphically above the Taconic unconformity (Stop 1).

Along the outcrop (ca. 11.6 m-thick), two channel sandstone bodies and two sandy to muddy floodplain deposits, including ancient soils (paleosols) are visible. Fining up pairs of channel sandstones to floodplain deposits in the Catskills are interpreted to represent thousands to tens of thousands of years (Bridge, 2000).

The lower “mudrock” unit (ca. 2.0 m-thick) is characterized by dark gray to gray shales that grade upward into interbedded thin sandstones and green mudstones. The upper 40 cm have a blocky texture and feature small-scale slickensided surfaces (pedogenic slickensides), associated with soil development.

Lower strata of the overlying sandstone body (ca. 4.2 m-thick) grade laterally between gray sandstones, dark gray shaly sandstones, and intraformational conglomerates, of which at least one is dominated by calcareous nodules, eroded and reworked from erosion of paleosols upstream. Cross-bedded sandstones above feature more than one channel (multi-storied) up through the succession.

The top of the channel is marked by a 0.7 m-thick interval of more tabular sandstone beds, which grade upward into ca. 1.2 m of red mudrock-dominated strata, deposited across a floodplain after migration of the channel. Centimeter-scale vertical traces, often green in color represent root traces of *Eospermatopteris*-type cladoxylopid trees, similar to the famous tree stumps found near Gilboa, higher in the Catskills. In places, the roots have extensively bioturb-

ated the sediments. In addition, the upper 40-60 centimeters feature multiple “dish/gilgai/pseudoanticline” soil deformation structures, and pedogenic slickensides occur through the strata. These features form by wetting and drying of expandable smectite (montmorillonite) clay rich, fine-grained sediments, and indicate the development of vertic paleosols.

The upper unit (ca. 3.5 m exposed) is another multi-storied channel sandstone body. Mudrock lenses occur locally along the outcrop, and prominent, down-cutting erosional surfaces are overlain by additional channel deposits. The sandstone bodies are interpreted to be deposited in single channel, sinuous (“meandering”) rivers, which migrated across vegetated alluvial plains (Bridge, 2000).

Sandstone petrologic studies by Gale (1985) through the Plattekill to Slide Mountain formations in the Catskills found that composition of the sandstones is relatively consistent. Mono- and polycrystalline quartz, foliated metamorphic rock fragments and sedimentary rock fragments comprise 80% or more of the sandstones (Gale, 1985). Low in the succession (e.g., Plattekill Formation), sandstones contain a greater percentage of foliated metamorphic rock fragments and a correlative lower concentration of quartz. The sandstones are relatively clay poor, and are best defined as lithic to sub-lithic arenites (Gale, 1985). Overall, the sand-size fraction increases in grain-size upward through the succession, as environments change from lowland to transitional lowland-upland alluvial plain environments.

Plattekill Formation sandstones analyzed by Gale (1985; 10 samples) feature concentrations of foliated metamorphic rock fragments between 19-47%, in sharp contrast with the Slide Mountain Formation (3-17%; 5 samples); total macrocrystalline quartz (mono- + polycrystalline) in the Plattekill comprises 25-50 % of the rock, compared with 47-62% in the Slide Mountain Formation. In addition to quartz, metamorphic rock fragments and sedimentary rock fragments (10-23%), lesser amounts of chert (0.4-3.6%), plagioclase and orthoclase feldspars (0.2-1.9%), along with chlorite and micas, illite and kaolinite are found in Plattekill Formation sandstones (Gale, 1985).

Analysis of the clay minerals in the Catskill succession, and marine strata beyond, indicate a dominance of illite with lesser amounts of chlorite and minor kaolinite (e.g., Friend, 1966; Hosterman and Whitlow, 1983). However, vertic paleosols, as relatively well developed in the upper floodplain deposit, form in smectite-rich sediments. These clays, which swell and shrink with wetting and drying, are derived from igneous sources. The abundance of vertic paleosols in the Catskill succession appear to represent an otherwise hidden component of sediments derived from igneous sources. The sources could be derived from weathering of plutonic or volcanic rocks in the source area, or from volcanic ash erupted from explosive volcanic eruption in the Acadian orogenic belt.

At this time, only intraformational conglomerates/breccias are known by the author in the Plattekill Formation (Figure 9d). The clasts in these beds consist of reworked calcareous nodules (“peds”) reworked from paleosols, or chips of mud. They commonly occur in the base of channel sandstone deposits.

- At end of Stop 5, return to cars.
- 60.0 Proceed ahead on Rte. 23.
- 60.3 Continue straight ahead. Rte. 32 turns to right (North). More excellent exposures of Plattekill Formation are visible along Catskill Creek, 1.4 miles north. Outcrops include well exposed bedding plane and cross-sectional exposures of fluvial channel sandstones and floodplain mudstones. Also found (some distance upstream of abandoned dam) is a ~1.5 m-thick interval of carbonate-rich facies, capped by a thin (ca. 20 cm-thick) limestone bed indicative of lacustrine facies in floodplain environments. NOTE: Exposures along the creek are on private property. And pay attention to extensive No Parking signs along Rte. 32 near Catskill Creek.
- 60.6 Get into left lane.
- 60.85 Stop sign at intersection with Rte. 145. Proceed ahead.
- 61.0 Exposures of Plattekill Formation along Rte. 23.
- 61.5 Peaks of Acra Point, Burnt Knob, and beyond, Windham High Peak (3524') visible in distance.
- 66.1 Cairo-Durham town line. Approximate position of Plattekill-Manorkill (“Moscow”) formational boundary, according to Fisher et al., (1970).

- 66.9 Good exposures of Manorkill Formation along south side of road, near interesting and unusual buildings. Additional exposures of the Manorkill along highway ahead.
- 68.0 Fork right into large parking area. Pull ahead and park.

LUNCH STOP (20 MINUTES).

STOP 6. FLOODPLAIN AND CHANNEL SANDSTONE DEPOSITS, EAST WINDHAM (45 MINUTES).

Over 60 m of mudrocks, sandstones and a thin limestone bed in the Manorkill Formation at Stop 6 represent deposition in floodplain, fluvial channel, and lacustrine environments on the subaerial delta plain of the Catskill delta complex. It is possible that a part of the section may represent some brackish water conditions however. This outcrop is roughly 1.25 km stratigraphically above the Taconic unconformity at Stop 1.

Low in the part of the outcrop facing the parking area, another red paleosol zone with green, cm-scale diameter root traces of cladoxylopsid trees is visible, more easily seen than those at the last stop. Additional paleosols along the outcrop show varying development (Mintz et al., 2006). Along the outcrop, floodplain deposits vary between red, green, yellow-tan and dark gray/black mudstones and lesser fine-grained sandstones. Several sandstone bodies occur along the outcrop also, though few are thicker than 1-2 m.

Around the bend and uphill from the parking area, a prominent sloping-to-the-right interval of yellowish-green strata is succeeded by a zone of thin sandstones and dark gray mudrocks, at a distinctly different angle to the underlying beds. A number of isolated soft sediment deformation pillows occur along the outcrop within the lower approximately 1-3 m of the upper unit.

Higher in the section, downhill from the first driveway uphill of the bend, a prominent, thin ledge of limestone sticks out from the outcrop. Fallen slabs of the bed can be seen in the talus. Light gray to brown-gray, varyingly smooth to knobby in appearance, the fauna noted in the bed consists of ostracodes. Mintz et al. (2006) state that the bed appears to have been pedogenically modified and brecciated. It is overlain and underlain by red to green paleosols.

Apparent lacustrine (lake) and palustrine (wetland) facies, including similar thin limestones, are not often discussed but not unknown in Upper Devonian Catskill magnafacies (DeMicco et al., 1987; Dunagan and Driese, 1999). Similar limestones are documented from Devonian terrestrial facies in Canada, Great Britain and Australia (e.g., Donovan, 1975). They are generally interpreted to have formed toward the center of ponds and lakes, beyond the transport of fine-grained clastics. Carbonate is derived from calcareous, photosynthetic algae (e.g., charophytes). The author has found multiple apparent freshwater limestones in the Plattekill and Manorkill formations along the Catskill Front. Lacustrine and palustrine/wetland environments comprise an interesting and relatively overlooked facies in the Catskill Front, which deserves more attention.

A short distance above the limestone, along the lower part of a driveway, an interval of olive-colored, mudrock-dominated strata above the limestone, best seen along the lowest part of a driveway, features common fish bone material, ostracodes and desiccation cracks. Also found in the interval are *Spirophyton* trace fossils, which have been interpreted by some to have lived in freshwater setting (Bridge and Gordon, 1985; DeMicco et al., 1987), but by others (Gordon, 1988; Miller, 1991) to indicate brackish water conditions. Miller (1991) proposed that the animals producing *Spirophyton* lived in ephemeral ponds on the coastal floodplain, with fluctuating fresh- to brackish water salinities, perhaps tied to floods of brackish water that flowed upstream and spread across floodplains during major storms.

As in other fluvial-dominated strata of the Catskills, Manorkill Formation sandstones analyzed by Gale (1985; 3 samples) feature common mono- and polycrystalline quartz (29.6-60.0%), and foliated metamorphic and sedimentary rock fragments (30.4-41.8% and 9.3 to 22.0%, respectively). Lesser amounts of chert (0.2-0.7%), plagioclase and orthoclase feldspars (1.6-4.3%), along with chlorite and micas, illite and kaolinite are found in Manorkill sandstones (Gale, 1985). This represents a subtle shift toward increased quartz content from the underlying Plattekill Formation. With the exception of a single bed at or near the base of the Manorkill Formation, conglomerates known to the author at this point are intrabasinal ones, with reworked mud or pedogenic carbonate nodules.

At end of Stop 6, return to cars.

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- 68.0 Drive ahead to exit from parking area. Then, TURN LEFT, and return back downhill on Rte. 23 toward Cairo.
- 69.1 Peak of Burnt Knob and/or Acra Point ahead.
- 75.4 Intersection with Rte. 145.
- 75.9 Intersection with Rte. 32, which joins Rte. 23 for 1.1 miles. Village of Cairo on right.
- 77.0 Turn right, and follow Rte. 32 south.
- 77.25 Stop light, village of Cairo on right.
- 79.0 Channel sandstones of Plattekill Formation. All exposures along Rte. 32 ahead are in the Plattekill Formation.
- 82.4 Area between here and turn at Rte. 23a are called the Kiskatom Flats, which we'll see from above at Stop 7.
- 84.8 Turn right onto Rte. 23a, toward Palenville and Kaaterskill Clove.
- 85.7 Note rise of the highway, ascending up alluvial fan out front of Kaaterskill Clove.
- 87.1 Stop light in Palenville.
- 87.6 Enter Kaaterskill Clove.
- 88.0 Lower bridge in Kaaterskill Clove. Cross over Kaaterskill Creek. Excellent exposures of Plattekill Formation upstream, including channel sandstones, floodplain deposits, and at least one more apparent freshwater limestone.
- 88.3 Area of "High Rocks" of Chadwick (1944, Fig. 47).
- 89.0 Middle bridge in Kaaterskill Clove. Excellent exposures of red mudstone/paleosol facies of Manorkill Formation at and above bridge.
- 90.5 Upper bridge in Kaaterskill Clove. Excellent exposures of Oneonta Formation upstream, along trail to base of Kaaterskill Falls.
- 90.7 Parking area for trail to Kaaterskill Falls.
- 90.8 Classic exposures of Oneonta Formation along road, with well developed paleosols.
- 91.4 Twilight Park entrance on right, near village of Haines Falls.
- 91.6 Top of Kaaterskill Clove
- 92.1 Turn right onto North Lake Road.
- 93.9 Bypass road to Kaaterskill Falls to right.
- 94.4 Entrance to North-South Lake Campground, in Catskill Park. Drive ahead to booth and pay entrance fee (\$8/car at time of field trip). Proceed forward to parking area for North Lake. At stop sign ahead, continue ahead to North Lake.
- 96.1 Parking area for North Lake and trails along the Catskill Escarpment.

REST STOP: Restrooms on the east side of the parking area at North Lake.

Stop 7. UPPER DEVONIAN CLASTICS, NORTH-SOUTH LAKE (1.5 HOURS). Note: No hammers or collecting at this stop (a Catskill Park campground, run by NYS-DEC).

At this stop, we will walk north along the Catskill escarpment to Artists Rock and Sunset Rock. The rocks exposed comprise lower strata of the Upper Devonian Oneonta Formation, and will include the Twilight Park Conglomerate Member (at Sunset Rock). The latter outcrop is approximately 1.5 km stratigraphically above the Taconic unconformity at Stop 1. Another ~1.2 km of strata overlie the Twilight Park Conglomerate, to the top of Slide Mountain, the highest peak in the Catskills.

The Escarpment trail going north from the parking area at North Lake slowly rises upward through a major sandstone ledge that caps the Catskill escarpment here. Termed the "Kaaterskill Sandstones" by Chadwick (1944), they form the lower part of the Upper Devonian Oneonta Formation. Sandstones dominate the strata along the trail to Artists Rock. An abandoned trail a short distance beyond Artists Rock exposes red mudstones that lie between major sandstone packages.

Partway along the trail to Artists Rock, shortly after a several-meter rise up through the lower Oneonta sandstones, an odd set of sedimentary structures occur in the sandstones in the trail. At first glance, they appear to be trough

cross beds. On closer observation, however, the edges of the troughs are vertical to near vertical, well beyond the angle of repose. The vertical edges are water-escape structures, along the margins of foundered bowls of sandstone. This represents another soft sediment deformation zone, developed in sand-only facies (as found at Stop 4). The author has not examined their correlatability, although previous workers have noted disturbed soft sediments elsewhere in the area, in the lower part of the Oneonta Formation. If the feature is relatively widespread, it could possibly represent a “seismite”, triggered by a significant seismic shock during the Acadian orogeny. Without more evidence (e.g., correlatability across a broad area), this is only conjecture.

Mono- and polycrystalline quartz comprise approximately 37.9-59.1 % of the rock in petrographically analyzed sandstones of the Oneonta Formation (Gale, 1985). The concentration of foliated metamorphic and sedimentary rock fragments (15.7-33.4% and 8.7 to 21.1%, respectively) have decreased relative to underlying strata, while the concentration of quartz has increased. The concentration of other components (e.g., chert, feldspars, chlorite and micas, etc.) remain about the same as in the underlying Manorkill and Plattekill formations.

There is a rise of approximately 530 m (1750') from the Kiskatom Flats below. In the distance to the east, the Taconic Mountains are visible along the New York-Massachusetts border; high peaks of the Berkshire Mountains (including Mount Greylock) locally project above the Taconics. Farther to the north/left, the Green Mountains are visible in Vermont. These highlands today expose low grade metamorphic rocks (e.g., slates and phyllites in the Taconics; some schist in the higher Berkshires and Greens). However, those rocks would have been deeply buried under younger rocks during the Devonian Acadian orogeny.

At Sunset Rock, up the trail beyond Artists Rock, a 23 m-thick outcrop of the Twilight Park Conglomerate (member of the Oneonta Formation) is very well exposed. The unit marks the first major progradation of clastics into the Catskill front from the Acadian orogen. Bridge and Nickelsen (1985) hypothesized that the progradation was due to increased slope due to tectonic changes, stating that climate did not appear to vary through the interval. In this case, a base-level drop associated with a significant sea level fall was not discussed. Bridge and students generally dismiss eustatic sea level controls over processes active in the terrestrial settings of the Catskill magnafacies. However, Devonian workers (e.g., House and Kirchgasser, 1993; Brett and Baird, 1996; Bartholomew, 2006; Ver Straeten, 2007a, in press) have now established significant eustatic control over sea level changes in New York's marine succession. And that tectonic patterns of flexure are superposed over the record of third order sea level cycles/stratigraphic sequences (Figure 8).

The author proposes a counter hypothesis, that Twilight Park gravels may have prograded basinward during one of the major sea level falls near the Middle-Upper Devonian boundary. As there is no tight control over where the Middle-Upper Devonian boundary actually occurs in the Catskill front, that sea level drop could be one of a few near the boundary.

Petrologically, clastic rocks of the Catskill front indicate that rocks exposed in the Acadian orogen consisted of dominantly low grade (up to greenschist) metamorphic and sedimentary rocks. Rare igneous- or high grade metamorphic-derived sediments indicate only very minor exposure of such rocks. A hidden source of more igneous than previously thought seems to be indicated by common vertic paleosols, which form in smectite clay-rich sediments (smectite clays are derived from the weathering of igneous rocks, including volcanic ash).

A number of previous authors hypothesized that Cambrian and Ordovician rocks in the Taconics region were the source of the Devonian Catskill delta sediments. However, beginning in the Late Silurian, the Acadian orogen and the associated foreland basin developed on the far margin of eastern North America. Through the Lower Devonian, massive volumes of Acadian-derived sediments were deposited in the foredeep basin in New England, while the Hudson Valley was positioned in the back bulge basin to cratonward edge of the foredeep of the foreland basin system. As the orogen progressively migrated cratonward toward eastern New York, Devonian sediments in that early Acadian foredeep basin (e.g., Littleton Formation of New England) were caught up in the deformational front of the orogen (e.g., wedge-top of the foreland basin system). Some of the Lower Devonian foredeep sediments were metamorphosed; metamorphosed or not, the foreland basin strata were thrust up, eroded and transported into the Catskill front. The clastic wedge, of which 2.7 kilometers thickness is preserved in the Catskills, by the latest Devonian ap-

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parently distributed muds as far west as northern Iowa (B. Witzke, pers. commun., 1999). Possibly by the Late Devonian, folding and thrusting of the foreland basin wedge-top may have migrated west of the Hudson River.

- At end of Stop 7, return to cars.
- 96.1 Return to Rte. 23a in Haines Falls.
- 100.2 Intersection with Rte. 23a in Haines Falls. Turn left, unless driving west through the Catskills to get home.
- 105.1 Stop light in Palenville. To proceed to the NYS Thruway southbound at Saugerties, fork right onto Rte. 32a, then Rte. 32 south. To proceed to the NYS Thruway northbound at Catskill, fork left and remain on Rte. 23a to Rte 9W, then Rte. 23 west, and then Rte. 23b to the Thruway entrance.

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Fluvial Geomorphology of the Upper Esopus Creek Watershed and Implications for Stream Management

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INTRODUCTION

The New York City water supply system depends on the maintenance of natural high water quality in the watersheds that drain into the reservoirs that store water for the system. Accordingly, concentrations of suspended sediment in both streams and reservoirs are major concerns, and the identification and remediation of sites that are concentrated sources of high suspended load/turbidity are essential to water quality management.

The concern about water quality by New York City's Department of Environmental Protection (DEP) is further motivated by a Filtration Avoidance Determination (FAD), which was renewed by the U.S. Environmental Protection Agency in 2007 and currently runs through 2017. This determination allows New York City to avoid the high capital cost that would be associated with building and maintaining a filtration system to treat the New York City water supply originating west of the Hudson River (~90% of the total supply). The FAD requires DEP to maintain high water quality through, among other practices, the development of stream management protocols and funding additional stream restoration projects (US EPA, 2007).

The DEP established a Stream Management Program to oversee the assessment of stream corridor condition and the development of management strategies for the watersheds that comprise the West-of-Hudson water supply system (Figure 1), including the Ashokan Reservoir/Upper Esopus Creek watershed (Figure 2). The DEP has worked with municipal, county, state and federal partners to develop Stream Management Plans (SMP) for many of the principal streams supplying the Ashokan, Schoharie, Cannonsville, and Pepacton Reservoirs. The work in the Upper Esopus Creek watershed that will be summarized in this field trip has been part of this overall effort (see www.catskill-streams.org).

The purpose of this field trip, then, is to explore the Upper Esopus Creek watershed, highlighting historic changes to the creek channel (Miller, 2009) and strategies that have been implemented or are being considered to alleviate or manage suspended sediment yield. We will examine sites where lateral migration or change in base elevation of the channel is degrading water quality or threatening property damage, where bank erosion has created a source of turbidity, and where stream restoration/remediation strategies have been applied and will be applied in the future in efforts to remediate "hot-spots" of erosion and sources of high suspended sediment loading.

BACKGROUND

The watershed of Upper Esopus Creek occupies an area of approximately 500 km² (Figure 2) from headwaters on the south side of Panther Mountain to the upper end of Ashokan Reservoir downstream of Boiceville. The watershed streams drain the most rugged part of the Catskills, including the north slope of 4142 ft (1255 m) high Slide Mountain, and accordingly the main basin and its tributary basins have high relief. The base level controlled by Ashokan Reservoir is at an elevation of 633 ft (193 m). Nine main tributaries enter the Upper Esopus, including Broadstreet Hollow (Stops 5a and 5b), Stony Clove (Stops 6 and 7), and Beaver Kill (Optional Stops Bk1 and Bk2), which will be part of the field trip.

The watershed is subject to high precipitation levels, in large part due to the orographic effect of the surrounding high peaks. Annual precipitation increases from 36-42 inches (915-1065 mm) in the northwest to a high of 45-60 inches (1145-1525 mm) annually in the vicinity of the high peaks, based on regional maps from Thaler (1996) and records from the Slide Mountain, Phoenicia, and Shokan weather stations. Some 20% of the precipitation falls as snow, at least at the higher elevations, with an average of 100 inches (~2500 mm) of snow recorded at Slide Mountain from 1971



Figure 1. DEM of Catskill watersheds of the New York City water supply system. From Czekanski, 2005.

to 2000 (NYCDEP et al., 2007). However, often the precipitation falls as short-lived, high-intensity rainfall events that can produce significant flash flooding. For example, a widespread heavy storm June 26-28, 2006, dropped substantial rainfall (up to 14 inches—355 mm) across much of the western Catskills and produced extensive flooding in the upper Delaware and Susquehanna watersheds. Although much less rain fell over most of the Upper Esopus watershed, a persistent cell centered over the Beaver Kill on June 26, 2006, dropped an estimated 8-10 inches (200-255 mm) of rain, triggering extensive flooding in this tributary. And a flash flood was triggered in a tributary of the East Branch Delaware River on June 17, 2007, by a severe thunderstorm that dropped as much as 11 inches (280 mm) of rain mostly within a 3-hour period (Schaffner et al., 2008).

Historical land-use changes also have been significant in the Upper Esopus watershed. The watershed today is largely forested, but widespread logging and agricultural development in the middle to late 19th century likely had a strong impact on sediment yield in the basin. Tanneries and furniture factories (including one at Chichester, Stop 7) were developed in the watershed, and there were a number of sawmills that employed selective logging (Kudish, 2000). Clear-cut logging to supply charcoal kilns that supported the tanning and manufacturing industries (Evers, 1972) probably had an even greater impact. By the late 1800s, much of the watershed was deforested; high sediment yields certainly followed. The extent of legacy sediments that might be associated with these disturbances is largely unknown. The 20th century was marked by forest recovery, and slopes today are densely forested. Agriculture and housing occupies many of the larger stream terraces within the trunk valley and its tributaries. Streams likely were adjusting to changing sediment loads through at least part of the last century.

Water quality and sedimentation in Catskill streams is largely controlled by the underlying geology (Schneiderman, 2000). Most of the Upper Esopus watershed is underlain by Devonian sandstones and shales of the Oneonta and Walton formations (Fisher et al., 1970). However, the valley bottoms and many of the slopes are underlain by weakly to moderately consolidated glacial sediments that date from the Last Glacial Maximum. Ice from the LGM covered the Catskills, scouring bedrock and deepening many of the valleys. The ice retreat phase likely was characterized by lobes of the ice sheet from the Hudson Valley that remained in main stream valleys like the Esopus, even after most of the Catskills had been deglaciated (Cadwell, 1986; Dineen, 1986). Some of these lobes likely impounded lakes in the trunk and tributary valleys as ice blocked drainage of meltwater and surface runoff (Figure 3a; Rich, 1935). These pro-glacial lakes deposited laminated silt and clay units that locally exceed 30 m in thickness (NYCDEP et al., 2007). Local readvances of glacial lobes led to interbedding of till and other diamicts with glaciolacustrine sediments, resulting in very complex relationships like those exposed at stop Bk1. These interbedded tills and glaciolacustrine sediments can strongly affect subsurface water flow, leading to slope instabilities (particularly

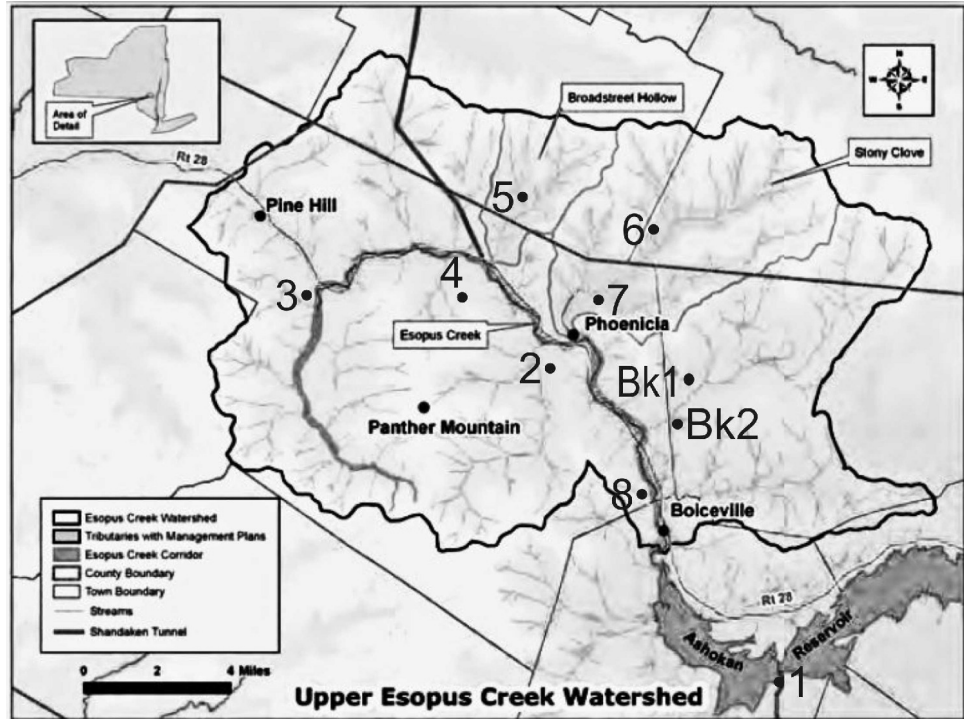


Figure 2. Map of Upper Esopus Creek watershed showing approximate locations of field trip stops. Modified from Cornell Cooperative Extension Ulster County et al., 2007.

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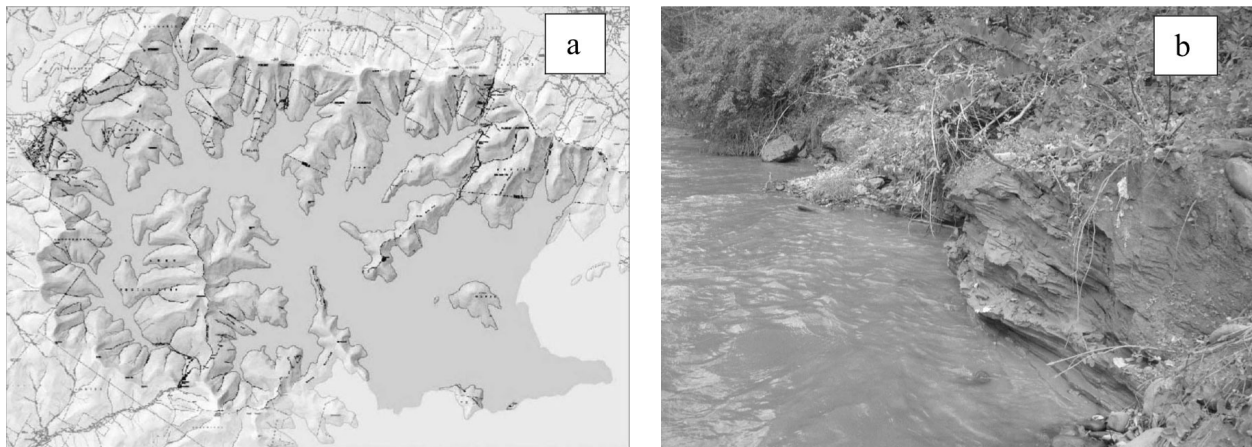


Figure 3. Glaciolacustrine conditions in Esopus Creek. a. Map of hypothetical Lake Peekamoose, from NYCDEP et al. (2007) based on Rich (1935). b. Exposure of laminated glaciolacustrine silt and clay in Esopus Creek.

where slopes are undermined by stream undercutting). Many of these glaciolacustrine deposits are overlain by post-glacial fluvial sediments, which can act as conduits for groundwater flow that can then pond on the finer deposits, as has apparently occurred at Stop 3. Bank erosion into the glaciolacustrine deposits produces the most significant sources of turbidity into Esopus Creek and its tributaries (Figure 3b).

HISTORIC CHANNEL CHANGES

Miller (2009) has investigated the extent of historic channel migration and changes in the main-stem Upper Esopus Creek using repeat aerial photographs and historic maps. He georeferenced aerial photographs and early topographic maps published in 1903 in order to quantify changes in channel position with time (Table 1). The 1903 maps provide some generalized indicators of changes in the center-line of the Esopus Creek channel, but because of a large RMS error, these were not used for quantitative comparisons with the later aerial photographs. However, the old maps are still useful in identifying locations along the valley where large changes have occurred in the channel configuration (Figure 4). These and other changes along the main-stem Esopus Creek are referenced to study reaches that were identified in Erwin et al. (2005) (Figure 5).

Channel changes have been imposed historically by construction of the railroad along the valley, the initial construction of Hwy. 28, and the more recent realignment of the state highway. The channel has been migrating along many reaches—in some cases due to meander migration (e.g. Reaches 16 and 2, Stops 3 and 8), in others due to Channel changes have been imposed historically by construction of the railroad along the valley, the initial construction of Hwy. 28, and the more recent realignment of the state highway. The channel has been migrating along many reaches—in some cases due to meander migration (e.g. Reaches 16 and 2, Stops 3 and 8), in others due to abandonment and re-occupation of secondary channels and/or avulsions of the primary channel (e.g. Reach 7, Stop 2). Miller and Knuepfer (2009) reported that channel changes such as bank erosion, bar deposition, vegetation encroachment on bar surfaces and the establishment and abandonment of secondary channels are correlated with the timing, frequency and duration of peak and bankfull discharge events. Interestingly, however, the impact of individual large flows isn't necessarily as dramatic as the impact of repeated moderate to large flow events. The largest flow of record at the

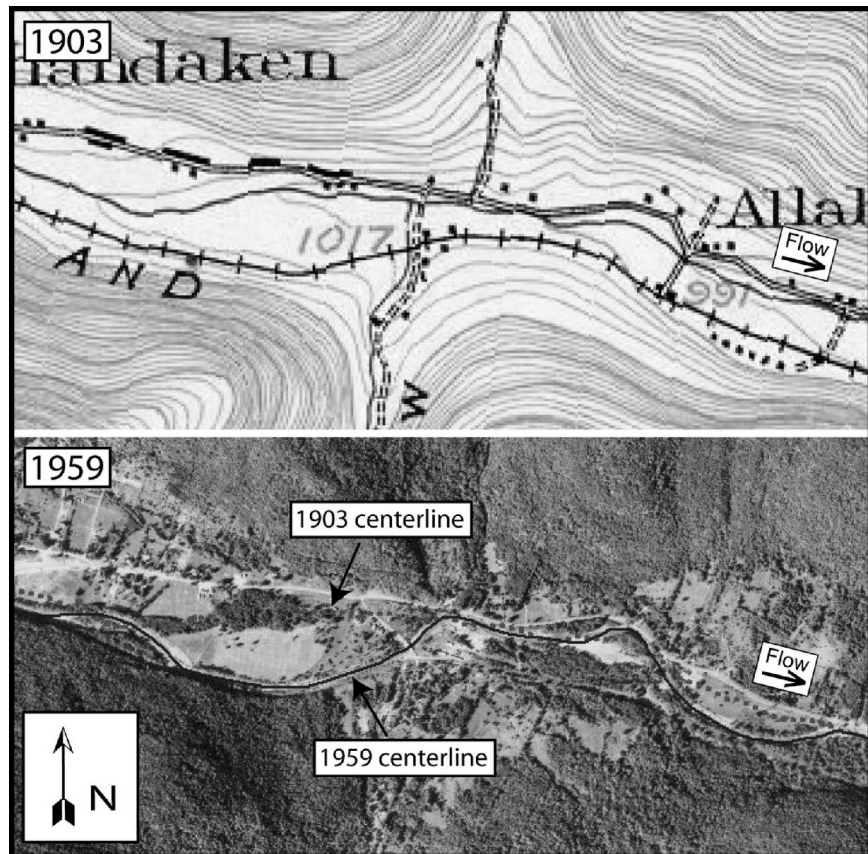


Figure 4. Comparison of 1903 map and 1959 aerial photograph of Esopus Creek near Allaben, NY, from Miller (2009). This is the vicinity of Stop 4 on the field trip. Note the abandonment of the 1903 channel where it is marked on the lower image. Reoccupation of this channel during the April 2005 flood led to washout of Fox Hollow Road and the destruction of several homes. The 1959 meanders right of the 1959 centerline label were cut off during realignment of Hwy. 28.

Format	Series	Date acquired	Scale	Number of images rectified	Maximum RMS error*(m)
Topographic map	1903	Aug. 1903	1:62500	2	26.8
Aerial photo	1959	9/5/1959	1:15840	9	2.3
Aerial photo	1967	4/30/1967	1:20000	7	1.3
Aerial photo	1968	6/5/1968	1:24000	2	1.0
Aerial photo	1980	9/11/1980	1:40000	2	0.5
Aerial photo	1980	10/23/1980	1:40000	5	2.3
DOQQ	1994-99	4/30/1997	NA	mosaic	NA
DOQ	2001	4/29/2001	NA	mosaic	NA

* Max RMS error refers to the highest RMS error among rectified images utilized in this study.

Table 1. Source imagery used in this study (from Miller, 2009).

Coldbrook gage just upstream of Ashokan Reservoir occurred on March 21, 1980, after a period of relatively low flows (Figure 6). The 1980 aerial photographs, taken later that year, show relatively little channel adjustment. However, by the time the 2001 imagery was taken, considerably greater changes had occurred at many locations along the channel, following several large flow events in the 1980s and 1990s (Figure 7). Miller(2009) argues that it is the *repetition* of large flow events over short recurrence periods that has the most significant impact on channel changes. Given that climate changes in upcoming decades are predicted to increase the frequency and magnitude of high-intensity storms in the Catskills (Frumhoff et al., 2007), this historic pattern implies that channel migration and bank erosion may be continuing and even increasing problems in the future.

CHALLENGES FOR STREAM MANAGEMENT

Geologic Controls on Erosion

Channel migration, realignment, avulsion, and cutoff have combined with the underlying surficial geology to produce areas of enhanced bank erosion and landslide development at numerous locations along Esopus Creek and its tributaries (NYCDEP et al., 2007). The Watershed and Stream Characterization section of the Upper Esopus Creek Management Plan (NYCDEP et al., 2007) describes the controls on bank erosion exerted by different types of unconsolidated sediments:

- Once exposed, glaciolacustrine sediment typically erodes readily during storm events. However, where the silt and clay unit is overlain by coarser fluvial sediment, exposure is typically short-lived, and the bank tends to get armored by the collapse and draping of the coarser sediment (Figure 8a). Some glaciolacustrine deposits are more resistant to erosion where clays are stiffer or where they have not been disturbed by older hillslope failures.
- The till tends to erode either as (a) mass slumping from saturated conditions or by (b) translational fracture-bound failures forming high steep banks (Figure 8b). In general, till exposures yield coarser bedload (such as at optional stop Bk2).
- The coarse-grained, non-cohesive fluvial sediment can erode easily if not protected by dense roots or revetment (Figure 8c).

Thus, mapping bank materials in detail is an important tool in identifying existing and potential sources of suspended sediment/turbidity. This work is ongoing in the main-stem Esopus Creek and in tributary watersheds

Stream Restoration

Water quality management can best be achieved by remediating sites where bank erosion and/or channel migration and avulsion has created a “hot spot” of suspended sediment input. While eroding banks are often “solved” by hard engineering structures (often out of short-term necessity where infrastructure is at risk), the DEP Stream Management Program team has focused on using natural channel design approaches of Rosgen (1996) as modified for the flashy Catskills streams. The underlying principle is to use some approximation of the “natural” fluvial system—including stream hydraulic geometry, planform, and riparian vegetation—to create channel geometries that are more appropriate for the stream type than hard engineering approaches tend to produce. The advantages of using natural channel design approaches include flood hazard mitigation (particularly by reconnecting the stream with its floodplain), improvement of instream and riparian habitat conditions, improvement in water quality and fisheries, and property protection (Baldigo et al., 2008). Such an approach to channel restoration ideally will yield self-sustaining

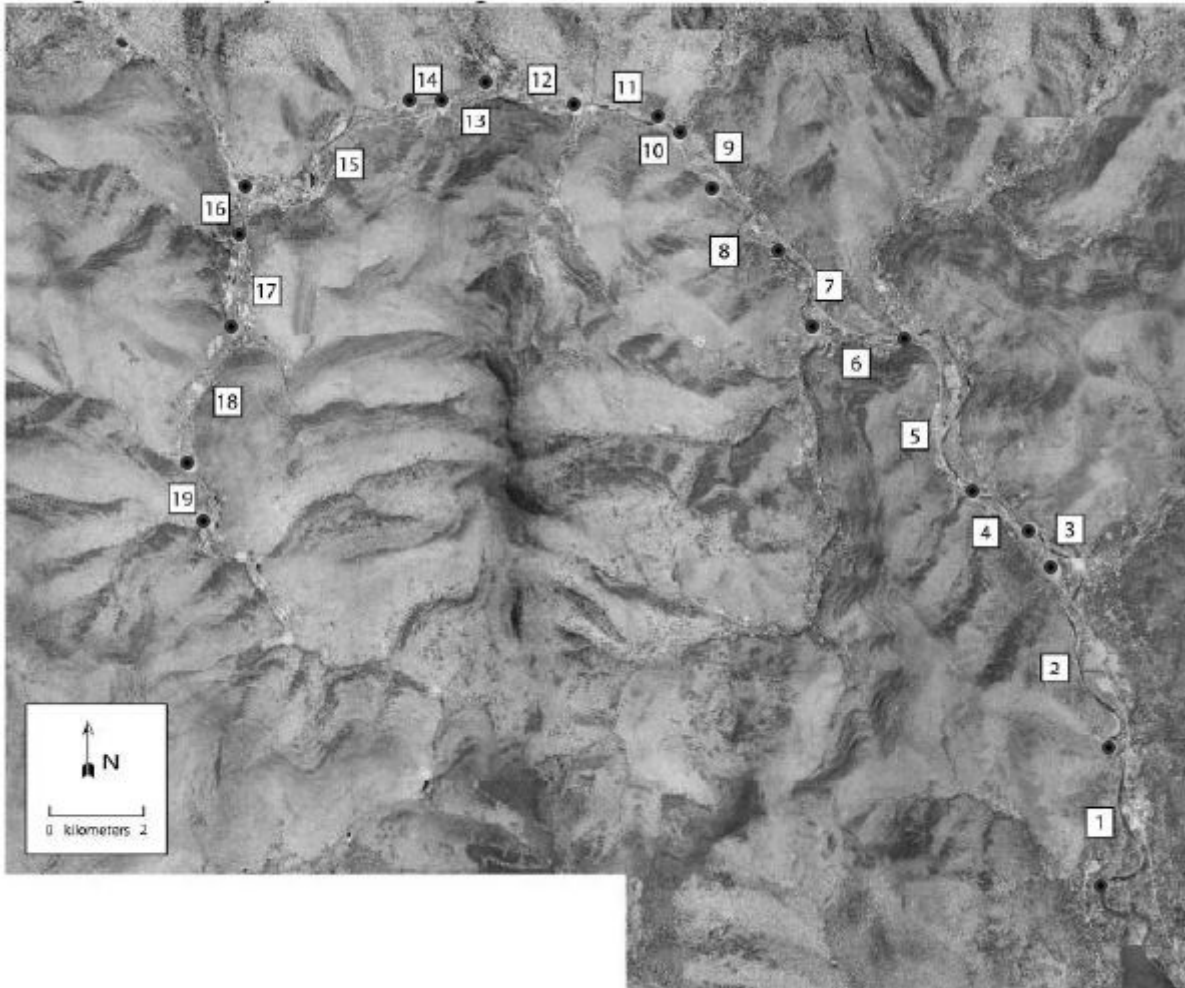


Figure 5. Reach breaks along Esopus Creek as defined in Erwin et al. (2005). Breaks based on changes in valley width, channel slope and channel pattern, and location of tributary confluences.

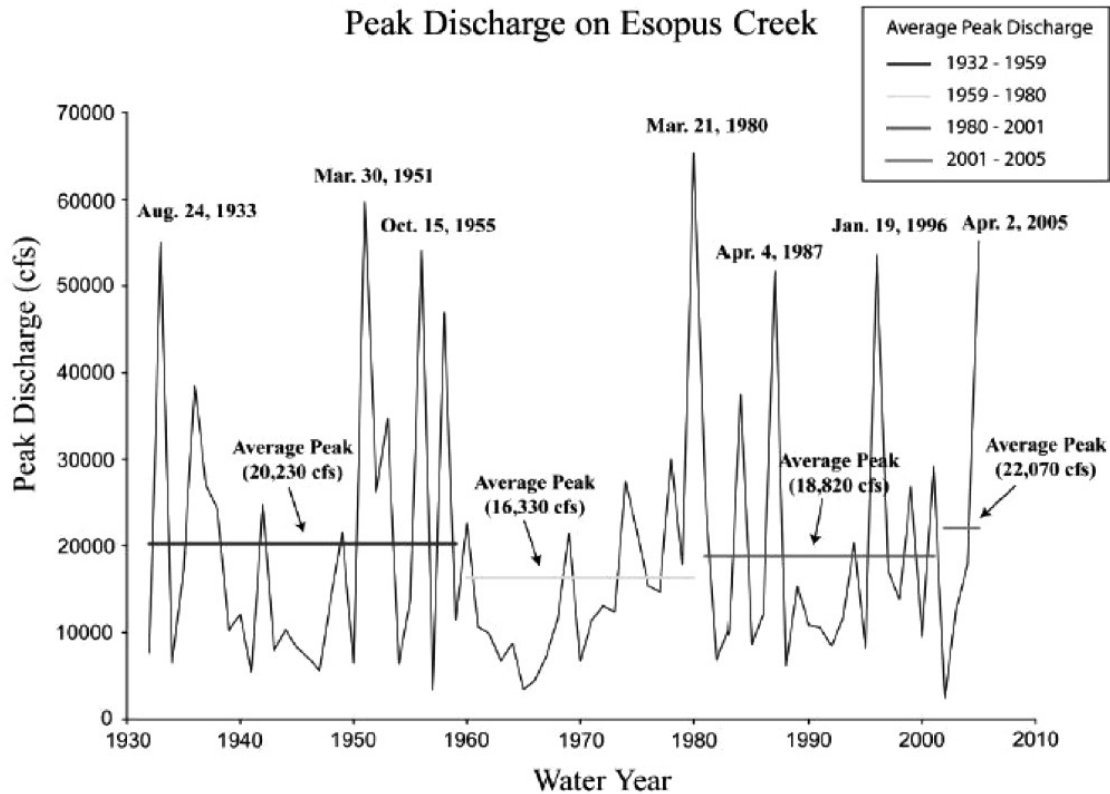


Figure 6. Historic peak discharge on Esopus Creek at the Coldbrook gage. Average peak discharge calculated for periods between aerial photographs. Especially low discharges characterize the 1959-1980 period (despite the flood of record in 1980). From Miller (2009).

stream reaches because of the use of the form (geomorphology) and function (hydraulics and sediment transport) of minimally impacted stream reaches.

The DEP Stream Management Program and partners have undertaken a number of demonstration projects in the Catskills with three occurring in the Upper Esopus watershed (Stops 2, 5, and 6). Two of these projects used the “Natural Channel Design” methodology taught by hydrologist Dave Rosgen (Rosgen, 1996). This methodology uses analog, or reference, morphology to determine channel and floodplain grades, width-to-depth ratios, planform and meso-feature spacing and sizing appropriate to the valley settings, based on Rosgen’s stream-classification system (Rosgen, 1996). One of the challenges in this approach is identifying bankfull stage and discharge—the benchmark discharge from which design parameters are drawn in the Rosgen method—particularly in the absence of gage records and detailed studies. Dunne and Leopold (1978) argued that the relationship between bankfull geometry and drainage basin area is reasonably consistent within individual hydrophysiographic provinces. Miller and Davis (2003) developed regional curves for the Catskill Mountains in order to provide a more appropriate basis for applying Rosgen’s classification system. The restoration projects themselves have used a variety of techniques, including the placement of instream structures to provide grade control, reduce shear stress at the channel margins and allow vegetation to mature sufficiently to perform these functions; reestablishment of meander patterns consistent with the “stable” reference reaches; and re-planting of riparian buffers using native riparian tree, shrub, and herb species (e.g., Greene County Soil and Water Conservation District, 2008).

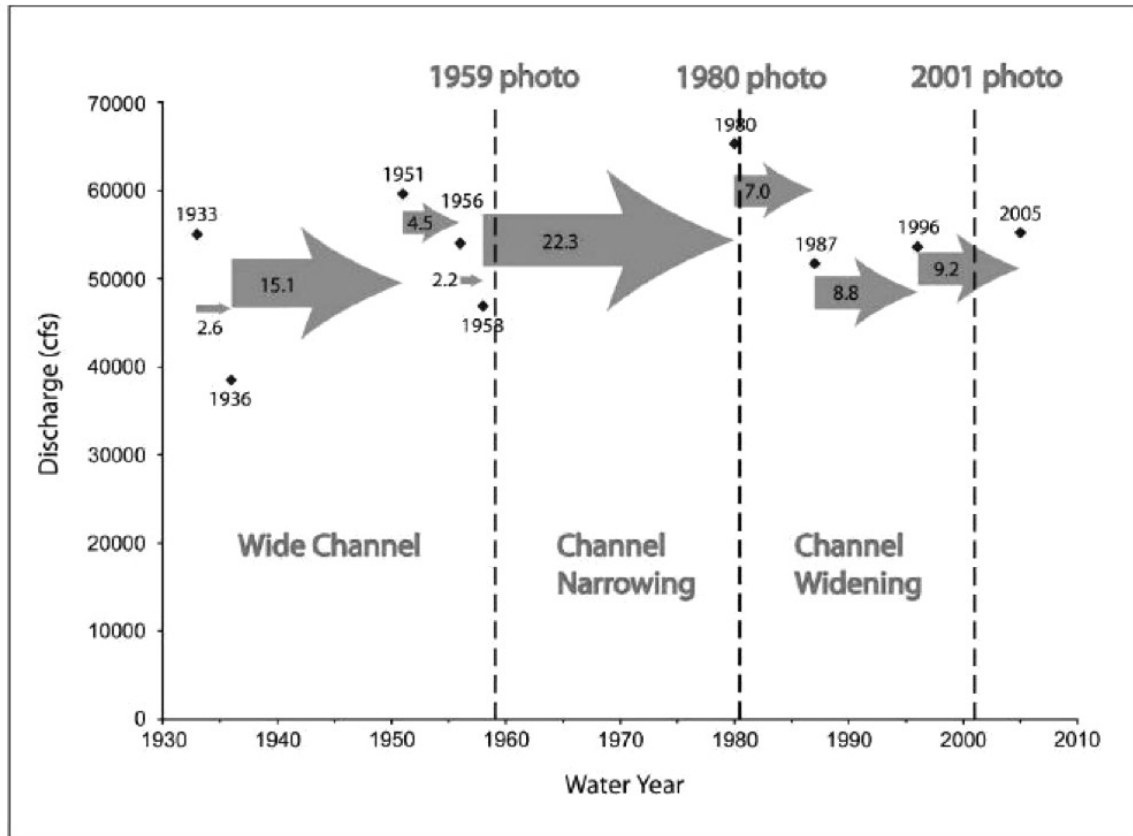


Figure 7. Intervals between large discharge events at Coldbrook gage and interpreted stream response. Note that bank erosion appears to occur mostly during intervals of frequent large flows. From Miller (2009).

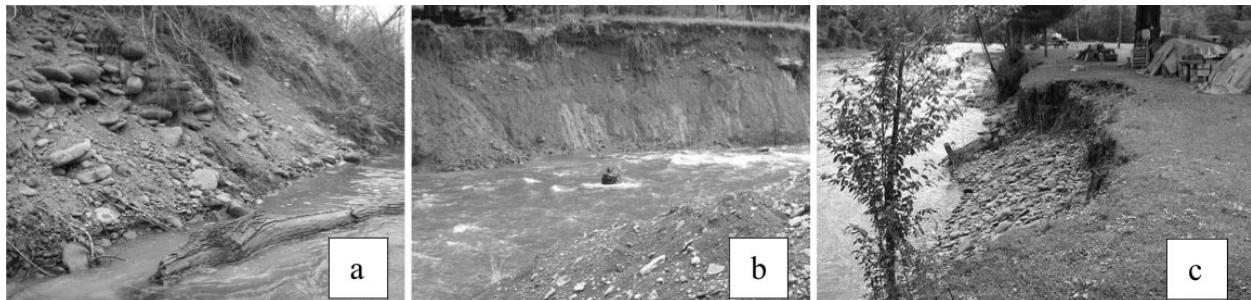


Figure 8. Bank erosion in different materials. a. Exposure of glaciolacustrine clay and silt has been armored by raveling of overlying coarse fluvial gravels. Such a process can decrease sediment loading compared to an un-armored exposure of glaciolacustrine sediments. b. Exposure of till (lighter color) overlain by post-glacial fluvial sediments. Note the steep exposure. c. Failure of a bank in unconsolidated fluvial deposits. All photos from NYCDEP et al., 2007.

SCOPE OF THE FIELD TRIP

The accompanying road log and site description provide a number of details about the Upper Esopus watershed, historic channel changes and the restoration demonstration projects. In addition, we will visit sites that have not been remediated, but are critical remediation targets, in order to better consider the application of natural channel design to the complex channel, bed, and bank geometries and geology that are found in the Esopus system.

ROAD LOG

Cumulative Mileage	Miles from Previous Site	Description
0.0	0.0	Leave Wooster Hall Parking Lot, SUNY New Paltz campus
0.1	0.1	Left turn onto South Mannheim Blvd (Hwy. 32)
0.4	0.3	Right turn onto Main St (Hwy. 299)
1.3	0.9	Right turn onto NY State Thruway onramp (Interstate 87)
1.9	0.6	Merge onto Thruway north towards Albany
7.5	5.6	Cross Wallkill River, incised into Ordovician shales and siltstones
10.0	2.5	Cross Rondout Creek, also incised into Ordovician shales. Lefever Falls on the left.
16.8	6.8	Cross Esopus Creek
17.0	0.2	Exit 19, NY State Thruway, to Kingston and Hwy. 28
17.5	0.5	Traffic circle at jct. Hwy. 28; exit to Hwy 28 W. Note that there is a Park and Ride just past the 3rd exit off the traffic circle (to Hwy. 587 and 28 E) that serves as an alternative starting and ending point for the trip.
20.2	2.7	Kings Town Stone Quarry on right into Middle Devonian Mt. Marion Formation.
21.8	1.6	Catskill Mountain Coffee on right.
27.3	5.5	Kenozia Lake on right. Drained in 2008 for repairs, the lake has refilled quickly in 2009 due to the heavy spring meltoff and rains.
29.8	2.5	Reservoir Road, Shokan; turn left (south) toward Ashokan Reservoir.
31.2	0.4	Weir that divides West and East basins of Ashokan Reservoir. Drive across to Stop 1.
31.6	0.4	Turn right into parking area (if you're an official vehicle); otherwise, turn left, find location to pull over, and walk back to Reservoir Road junction. Note that Monument Road here is on the Ashokan Dam.

STOP 1. ASHOKAN RESERVOIR AND ESOPUS CREEK WATERSHED. (15 MINUTES) (UTM location 18 T 565712E 4644450N). Ashokan Reservoir was the first of the New York City water-supply reservoirs to be constructed in the Catskills. The dam across Esopus Creek was completed in 1913 and the reservoir was impounded by 1914 (Schneiderman, 2000; New York City Department of Environmental Protection, 2009). Water is carried to New York City through the Catskill Aqueduct, completed in 1915. Because of concerns about the influx of suspended sediment from Esopus Creek, the reservoir is managed in two basins. The West Basin receives input from Esopus Creek (and the Shandaken Tunnel from Schoharie Reservoir), and is designed to act as a settling basin. It is separated from the East Basin, from which water is withdrawn into the Catskill Aqueduct, by the weir across which we have just driven (Figure 9). This overview provides a dramatic view into the mountainous upper Esopus Creek watershed. Also, an automated suspended sediment sampling site is located just east of the weir, visible from this overview site.

33.4	1.8	Retrace route back to Hwy. 28 and turn left (west).
37.6	4.2	Descend into Esopus Creek valley bottom. Jct. Hwy. 28A (on left) in Boiceville.
40.1	2.5	Junction Hwy. 212 (into Beaver Kill valley, location of optional field trip stops).
40.3	0.2	Cross Esopus Creek
41.5	1.2	Catskill Mountain Railroad on right
43.6	2.1	Turnoff to Phoenicia on right

- 43.8 0.2 Bridge across Esopus Creek; Stony Clove enters from left just downstream (right) of bridge. Stony Clove is a significant source of suspended sediment load/turbidity during high flow events and will be the location of Stops 6 and 7.
- 44.0 0.2 Junction Hwy. 214 (to Phoenicia and Stony Clove)
- 44.5 0.5 Junction Woodland Valley Road; turn left.
- 44.8 0.3 Bridge across Esopus Creek. Cross bridge and turn into unpaved parking area on left.

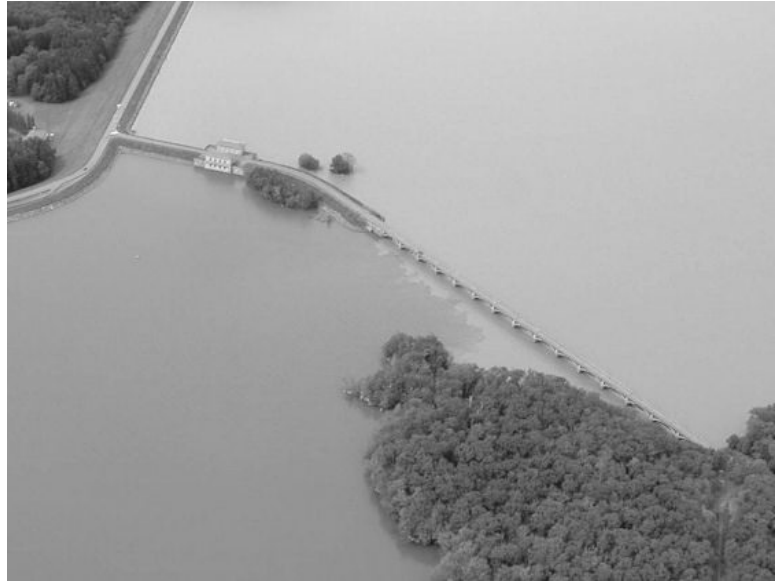


Figure 9. Weir across Ashokan Reservoir; view looking west after the June 2006 flood. Light colored water at top of photo (upstream side of weir) is actually red due to high suspended sediment load; note “leakage” across weir into East Basin of Ashokan Reservoir. Photo from NYCDEP et al., 2007.

STOP 2. ESOPUS CREEK AT WOODLAND VALLEY RESTORATION DEMONSTRATION SITE. (30

MINUTES) (Location UTM 18 T 555458E 4659007N). This stop in Esopus Creek Reach 7 will provide an introduction to the challenges in stream management of the Esopus Creek watershed and an illustration of the application of channel stabilization/stream restoration techniques. There is an informational kiosk at the site documenting the project history. The site is at the confluence of Woodland Creek with Esopus Creek—a setting that is inherently dynamic given the variability in magnitude of water and sediment discharge from the two mountain streams (Figure 10). Anecdotal and aerial photo evidence suggests that the channel and bar formation have undergone dramatic shifts in response to large flood events. The setting’s inherent “instability” is compounded by a double-span bridge located just downstream of the confluence that produces a clear constriction in channel dimensions. Following the January 1996 flood, a headcut into underlying glacial lake clay propagated through a secondary channel along the left bank and eventually captured most of the Esopus Creek flow (Figure 10). The altered alignment of the channel, the stratified composition of the 32-foot (9.8-meter) high terrace (from bottom to top: glacial lake clay, glacial till, pre-historic stream deposits; Figure 11) and subsequent floods resulted in a rapidly retreating eroding bank (approximately 3 feet or 0.9 m per year). Risks associated with continuing erosion



Figure 10. Aerial photo of the Woodland Creek/Esopus Creek confluence. The project reach extends from the head of the bifurcated reach down to the Woodland bridge (2001, DOQQ)



Figure 11. Esopus Creek at Woodward Valley restoration site before (left) and after (right) restoration project. Post-project photo taken July 13, 2004, approximately 9 months after construction, but before major flood in April 2005. From Barnet, 2004.

included increasing potential for exhuming several residential septic systems, causing additional property damage, producing a continuing source of turbidity and creating a hazard to recreational users of the stream (Figures 10 and 11).

In 2000, DEP hired FIScH Engineering (principal engineer Dr. Craig Fischenich) to complete an assessment and conceptual design for remediating this reach (Fischenich, 2001). The hydraulic and stability analyses, sediment transport calculations, and geomorphic assessment resulted in recommendations for an approach that incorporated several technical techniques including a channel relocation based upon natural channel design principles, bioengineering and traditional bank revetment, and habitat and recreational enhancement features. DEP selected this project to be the demonstration stream restoration project for Esopus Creek required by the US EPA as part of the FAD schedule of compliance (US EPA, 2007). The project was constructed in 2003 in two stages: the channel work, bank revetment, and flood plain reconstruction were completed by Oct. 1, 2003 and the vegetation (trees and willow fascines) and bioengineering (VRSS) were completed in early Dec., 2003. Figure 12 is an aerial view of the site taken approximately one year after construction (Aug., 2004).

- 45.1 0.3 Retrace route back to Hwy. 28 and turn left (west).
- 46.7 1.6 Office of Ashokan Watershed Stream Management Program on left (which will be lunch stop)
- 47.6 0.9 Bridge across Broadstreet Hollow (location of Stops 5a and 5b)
- 47.9 0.3 Shandaken Tunnel (location of Stop 4b) emerges on right side of highway



Figure 12. Esopus Creek Restoration Demonstration Project - one year after completion (August, 2004).

48.9	1.0	Fox Hollow Road on left (location of Stop 4a)
49.7	0.8	Esopus Creek bridge
52.6	2.9	Big Indian; junction County Route 47 to Olivera and Frost Valley, up the uppermost Esopus Creek.
52.8	0.2	Start of bridge across Birch Creek
53.1	0.3	Junction Lasher Road. Turn left.
53.5	0.4	Pull off at barn and walk down to Esopus Creek.

STOP 3. ESOPUS CREEK BANK FAILURE ON REACH 16. (40 MINUTES) (Location UTM 18 T 545312E 4661394N). Note: This is Private Property. Get permission from owner across road before crossing property to creek. Esopus Creek has been actively changing its channel configuration through an anabranching reach from near Hatchery Hollow about 3 km upstream from this site. This has included a number of avulsions and meander-bend migrations (Miller, 2009). At this site, a very tight meander bend has been expanding westward at least since 1959 (Figure 13). This expansion has resulted in undermining and rapid headward growth of propagating failures in an elevated terrace that forms the left descending stream bank. Fluvial gravels overlie interbedded glaciolacustrine clays and silts and glacial till. Stormwater runoff and groundwater flow off the adjacent mountainside has limited infiltration potential due to the impermeable glacial deposits and emerges as springs on the creek bank. Water does infiltrate into rotational slip surfaces in the underlying lacustrine unit resulting in very active collapse of the bank (Figures 14 and 15). This reach is often the upstream source of suspended sediment into Esopus Creek following high water events. This reach is part of DEP’s long-term geomorphic monitoring program with 6 monumented cross sections, repeated longitudinal profile surveys, sediment sampling, and GPS-based mapping of channel feature alignment.

53.9	0.4	Return to Hwy. 28 and turn right, back down the Esopus Valley
57.5	3.6	Junction Hwy. 42 on left at Shandaken. The highway climbs up Bushnellsville Creek through Deep Notch and into the Schoharie Creek watershed.
58.1	0.6	Junction Fox Hollow Road. Turn right.

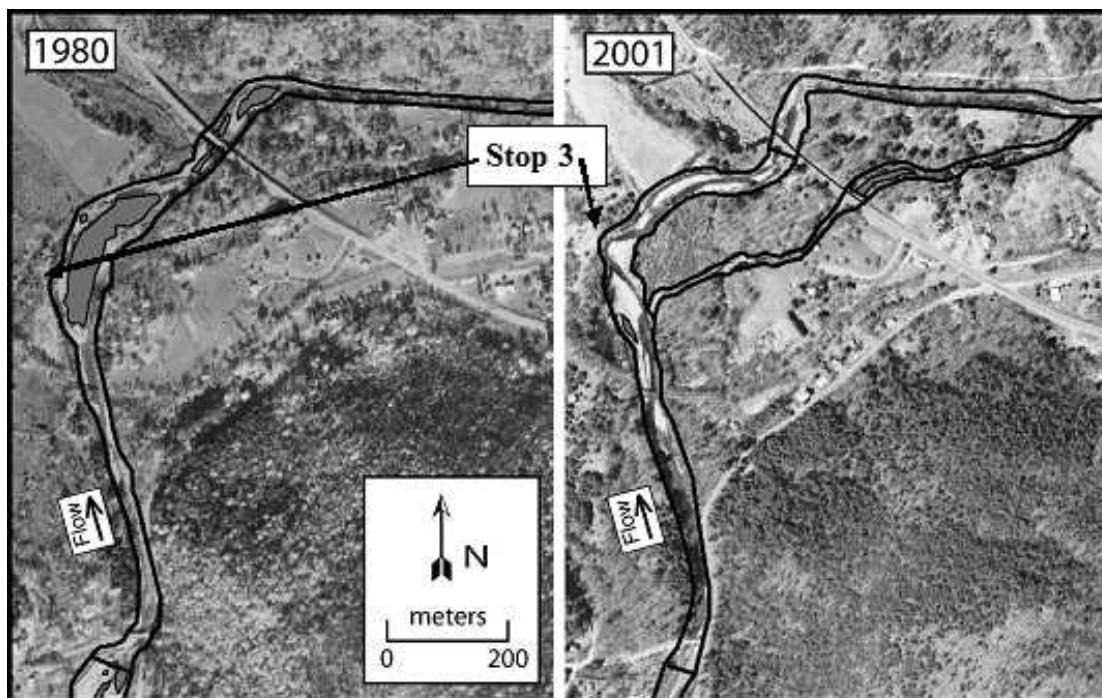


Figure 13. Changes in channel complexity and meander migration, vicinity of Stop 3. From Miller, 2009.

58.3 0.2 Cross Esopus Creek and pull off.

STOP 4A. ESOPUS CREEK GAGE AT ALLABEN (REACH 11/12). (20 MINUTES) (Location UTM 18 T 551243E 4662925N). A USGS gaging station, maintained in cooperation with the DEP, has been operating at this site since 1988; from 1963 to 1988 a gage was located about 0.8 km upstream. Bankfull discharge is ~3000 cfs (85 cumecs) has been exceeded numerous times during the period of record, and the site has recorded peak discharges of at least 15,000 cfs (425 cumecs) 4 times. The estimated 100-year flood at this site is around 26000 cfs (~740 cumecs) (NYCDEP et al., 2007), whereas the largest flood of record—April 2, 2005—was 21,700 cfs (~615 cumecs). This site also has been used as a location to measure the hydraulic geometry of the bankfull channel. Miller and Davis (2003) used this as one of their calibration sites in developing regional bankfull discharge and hydraulic geometry curves for the Catskills region.



Figure 14. Aerial view of meander bend along Esopus Creek undergoing failure at Stop 3. Photo taken April 10, 2005. From NYC DEP et al., 2007.

Realignment of Hwy. 28 in the 1960s was accompanied by cutoff of a natural meander downstream of the bridge (Figure 16; Miller, 2009). Miller (2009) describes the changes in this way: In 1959 the stream bifurcated around a large island (I11-1) upon entering the meander bend. But with road construction and cutoff of this meander, a new island (I11-3) formed just upstream of the old meander bend, as seen on the 1980 aerial photograph (Figure 16). The stream impinged on the road 200 meters farther upstream (20.2 km). By 2001, this island was attached to the left bank and no longer part of the channel. A new island (I11-5) has formed 140 meters farther downstream due to a channel avulsion across the floodplain leaving the old thalweg between the newly formed island I11-5 and the road.

58.4 0.1 Return to Hwy. 28 and turn right

59.4 1.0 Pull off on left side of highway at Shandaken Tunnel portal.

STOP 4B. SHANDAKEN TUNNEL PORTAL. (10 MINUTES) (Location UTM 18 T 552615E 4662710N). The Shandaken Tunnel brings water from Schoharie Reservoir to Esopus Creek to contribute to the New York City water supply system. This 18-mile (29-km) long underground tunnel, built by hand between 1917 and 1924, was the longest handmade aqueduct in the world when it was completed. Both minimum and max-



Figure 15. View upstream toward actively retreating streambank with rotational failure scarps exposed in adjacent terrace (photo courtesy DEP, 2006).

imum discharge from the tunnel as well as turbidity limits are regulated by the New York State Department of Environmental Conservation in order to balance recreational (and ecological) uses of Esopus Creek downstream of the tunnel with water-supply needs of New York City.

- 60.6 1.2 Pull into Ashokan Watershed Stream Management Program office parking lot. **LUNCH STOP** (40 MINUTES).
- 61.5 0.9 Turn left (west) on Hwy. 28 to retrace steps to Broadstreet Hollow Road; turn right.
- 63.8 2.3 Pull off to side of road to observe reference reach used in the Broadstreet Hollow restoration project.

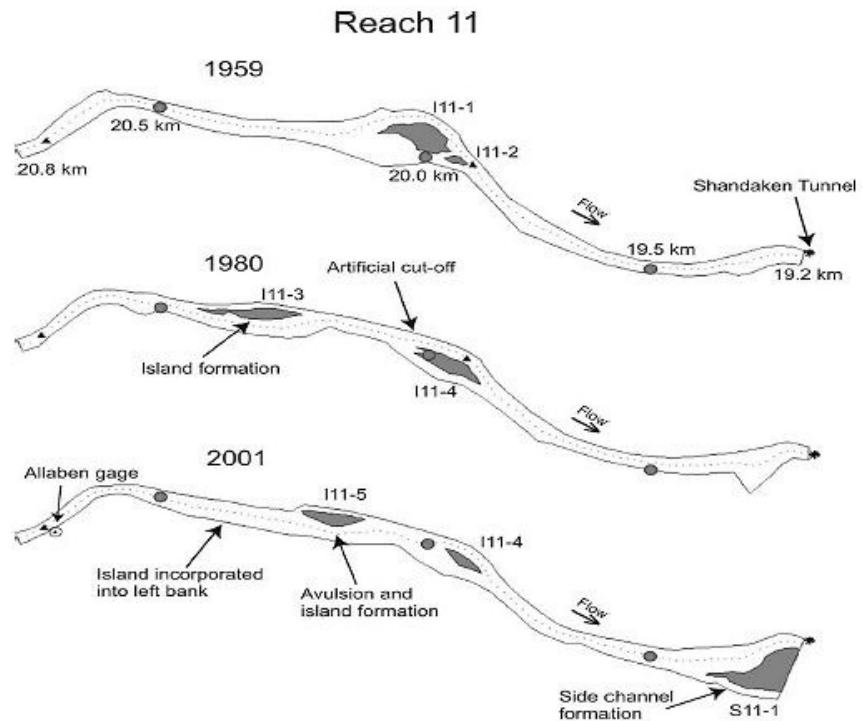


Figure 16. Overlays of georeferenced aerial photographs downstream of Allaben gage showing channel changes in response to artificial meander cutoff during construction of Hwy. 28. From Miller, 2009.

STOP 5. BROADSTREET HOLLOW RESTORATION PROJECT. Severe flooding in 1996 resulted in destabilization of a section of the right bank of the Broadstreet Hollow stream, with bank widening that threatened some houses on the left bank. An emergency response team relocated the channel and used extensive rock riprap to harden the banks and stream bed, but in doing so further destabilized the channel (Greene County Soil and Water Conservation District, 2008). Rapid incision of the channel into underlying glacial lake sediment was exacerbated by a high flow in 1999, resulting in destabilization of the left bank and development of a migrating artesian mud boil within the channel. This produced high turbidity levels, with Broadstreet Hollow becoming a significant source of turbidity for Esopus Creek. By 2000, the Broadstreet Hollow Restoration Project was initiated cooperatively among the Greene County and Ulster County Soil and Water Conservation Districts and the DEP Stream Management Program. The project was designed according to Natural Channel Design guidelines, using a reference reach upstream of the project area to define “natural” step-pool conditions and a construction program that attempted to recreate such conditions in the project reach.

STOP 5A. BROADSTREET HOLLOW RESTORATION SITE: REFERENCE REACH. (30 MINUTES) (Approx. location UTM 18 T 554947E 4665246N). The Broadstreet Hollow creek here displays a typical step-pool pattern with large cobble and boulder bedload. This reach is classified as a Rosgen B3 channel (Rosgen, 1996). Sediment size and distribution as well as channel morphology (hydraulic geometry and slope) provide the targets for design of the restoration reach using Rosgen’s natural channel design methodology. This site also has been monitored since project initiation; repeated cross-section measurements have been coupled with monitoring transport of large marked boulders to document changes in this “natural” channel reach.

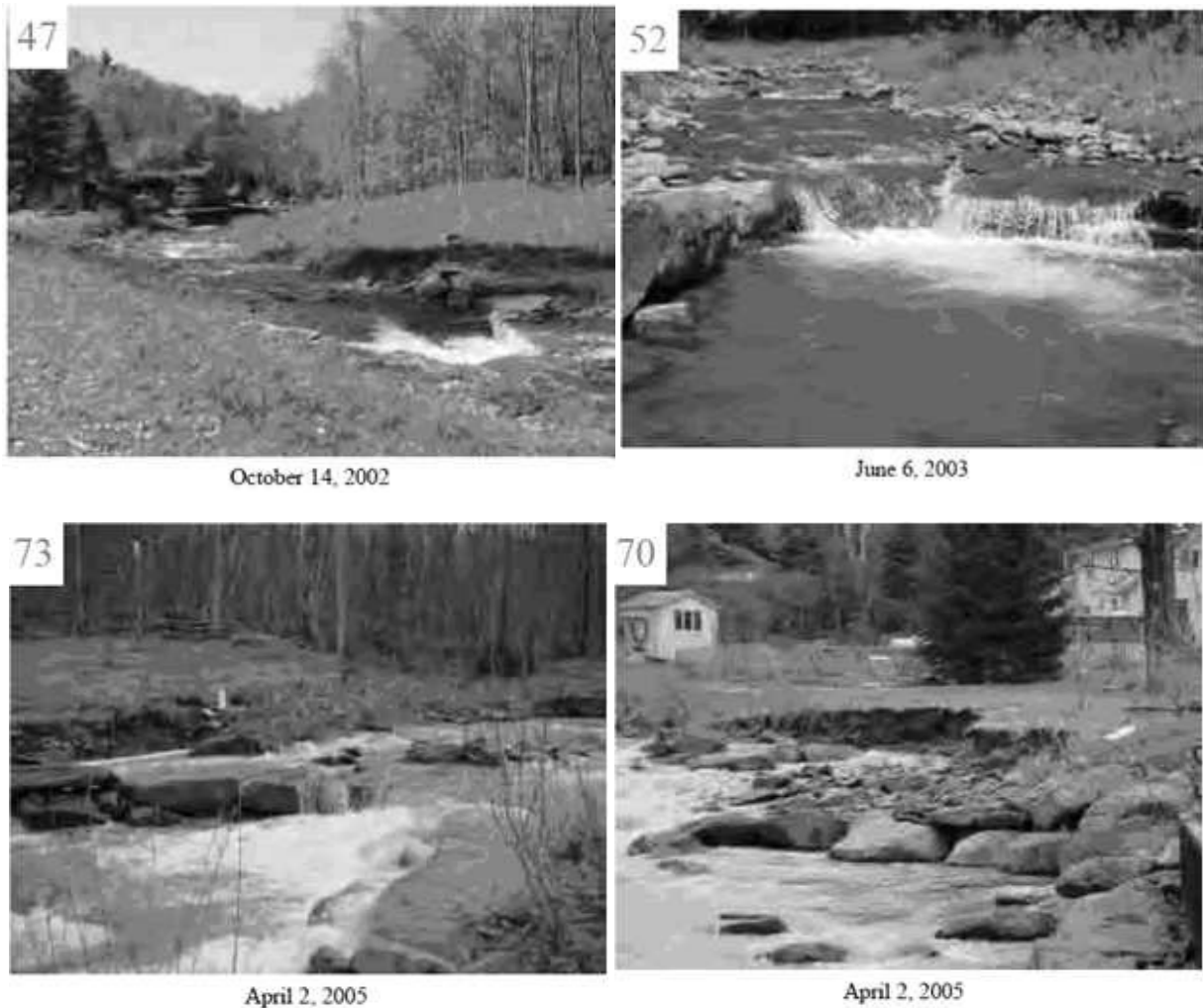


Figure 17. Broadstreet Hollow restoration site. Image 47 was taken prior to construction of the remediation project. Image 52 shows one of the rock vanes constructed across the creek. Images 70 and 73 show bank erosion resulting from high flow in 2005. Photos from GCSWCD, 2008.

64.2 0.4 Turn around and drive back down Broadstreet Hollow Road. Pull off on right at house to observe restoration site.

STOP 5B. BROADSTREET HOLLOW RESTORATION SITE: RESTORATION REACH. (30 MINUTES) (Location UTM 18 T 554624E 4664764N). Complex glaciolacustrine and post-glacial geology along with residential development have contributed to the challenges in restoring and stabilizing this reach of the Broadstreet Hollow stream. As reported by GCSWCD (2008), slumping of glaciolacustrine clays on the right bank has been fostered by artesian pressure in an interlayered sand deposit that has resulted in an artesian boil within the stream channel. Dewatering wells were installed; they have since been decommissioned. However, mud boils have continued to propagate upstream within the channel, still present in June 2009. Channel restoration was designed using the step-pool bedform and hydraulic geometry of the reference reach. Large quarried rocks were used to construct “cross vanes” to function as boulder-dominated steps in a constructed B3 channel (Figure 17). Several high flow events through this reach have caused some of the rock structures to be flanked or to need repair. Also, some of the struc-

tures have caused excess scour in the pool below the constructed step resulting in impaired fish passage at low flow. Willows and other riparian trees also were planted along the bank to add to bank stabilization.

- 66.1 1.9 Return to Hwy. 28 and turn left to continue downstream.
- 68.4 2.3 Junction Hwy. 214. Turn left into Phoenicia.
- 68.6 0.2 Left on Hwy. 214 into Stony Clove.
- 69.8 1.2 Left side of road undergoes routine repairs due to collapse of underlying glaciolacustrine clays. Note bedrock exposure across creek.
- 70.3 0.5 Bridge over Stony Clove; Stony Clove gaging station.
- 70.8 0.5 Junction Silver Hollow Road on right.
- 72.2 1.4 Greene County line
- 72.9 0.7 Pull off into field on right just before Lanesville hamlet sign.



Figure 18. Pre-restoration landsliding along right bank of Stony Clove at Lanesville site. Photo from DuBois, 2003.

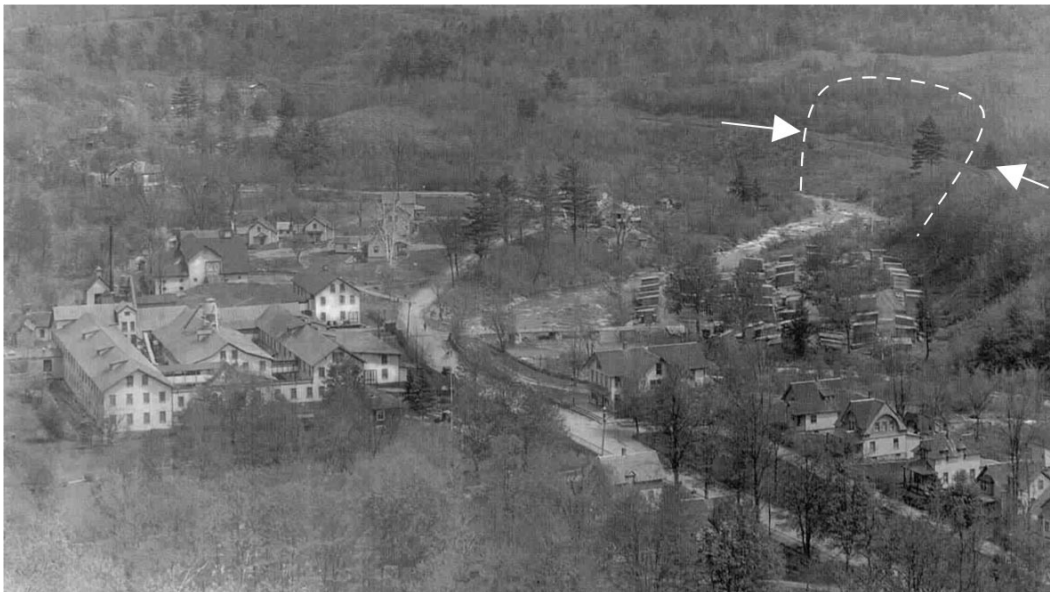


Figure 19. Photograph of Chichester, NY, ca. 1900 looking upstream on Stony Clove (channel on right side of photo). Note railroad grade above stream and alignment of channel away from bank and near the road. Area of modern landslide indicated.

STOP 6. LANESVILLE DEMONSTRATION RESTORATION PROJECT SITE ON STONY CLOVE CREEK. (30 MINUTES) (Location UTM 18 T 566227E 4663260N). A long history of channel modification and response has led to severe bank erosion and undermining of the right bank of Stony Clove just downstream of the hamlet of Lanesville. This reach was chosen for remediation because the development of a landslide on the left bank (Figure 18) was a significant source of fine sediment and turbidity into Stony Clove, Esopus Creek, and Ashokan Reservoir. This reach also had become increasingly confined by channel modifications, exacerbating the tendency to undermine the unstable left bank. Bank stabilization was promoted through construction of a bankfull bench at the toe of the slope, designed to re-establish a floodplain and minimize the interaction of most flows with the unstable bank (DuBois, 2003). Natural channel design principles were followed, including the construction of rock vanes and cross vanes within the channel and planting of a riparian buffer along the right bank and on the constructed bench (DuBois, 2003). The project was begun in 2003 but not completed until 2005 because of high water levels in the creek. Since construction this project has received several flow events around or greater than the bankfull discharge and has functioned quite well in stabilizing the reach and reducing turbidity downstream.

- 73.2 0.3 Turn around and drive back down Stony Clove. Meteorological station on right.
- 75.3 2.1 Pull off on left side of road to walk down to creek.

STOP 7. LANDSLIDE ALONG STONY CLOVE CREEK. (45 MINUTES) (Pulloff location UTM 18 T 567436E 4661227N). Get permission from landowner before walking down to creek. The Ashokan Watershed Stream Management Program Action Plan (http://www.catskillstreams.org/majorstreams_ec.html) identifies this reach of the stream between the Silver Hollow Bridge and the Route 214 crossing 1 km downstream as the highest priority restoration target for improving water quality in the Esopus Creek watershed. This area has a long history of modification, including development and construction of a railroad line (Figure 19) and various efforts to control the creek channel alignment. Today there are three distinct streambank/hillslope failures into glacial lake sediment and glacial till that are quite active. During spring snow melt and after rainfall runoff events this reach of stream is the most significant “point source” of turbidity for days and sometimes weeks after high water events (Figures 20 and 21). This stop is located at the largest and most problematic of the failures. It is a large active landslide initiated sometime prior to 1995. Upslope propagation of the slope failure resulted in collapse of a railroad right-of-way as well as increased groundwater and sediment discharge. The scale of this landslide presents especially great challenges for channel restoration and suspended sediment management. We will discuss restoration and management strategies being considered here. Given what we have seen elsewhere in the basin, how might the stream be managed through this reach to preserve/restore water quality?

- 77.3 2.0 Return to Phoenicia and turn left onto old Hwy. 28.
- 79.3 2.0 Mt. Tremper trailhead on left.
- 81.2 1.9 Entrance to Zen Monastery on left.
- 81.3 0.1 Junction State Hwy. 212. Turn right on Hwy. 212 to continue trip or turn left for optional side trip.



Figure 20. Landslide along Stony Clove upstream of Chichester Post Office. Photo from NYCDEP et al., 2007.

OPTIONAL SIDE TRIP UP BEAVER KILL. This optional trip gives access to a spectacular exposure of interbedded till and glaciolacustrine sediments as well as another extensive landslide that presents a stream management challenge. Mileage readings are for this side trip.

- | | | |
|-----|-----|---|
| 1.0 | 1.0 | Bridge over Beaver Kill |
| 3.1 | 2.1 | Bridge over Beaver Kill |
| 3.3 | 0.2 | Exposure of "Willow moraine". Pull over on left side of road. |

STOP BK 1. EXPOSED WILLOW MORAINE. (40 MINUTES) (Location UTM 18 T 0563899E 4657667N) This cutbank exposes a complex sequence of interbedded till and lacustrine sediments currently being studied by Andrew Kozlowski and colleagues from the New York State Museum as part of surficial geologic mapping of the quadrangles in this area. A lower, consolidated till is overlain by some 18 till units interbedded with lacustrine sediments. This exposure illustrates the complexity of glacial deposits in the Upper Esopus watershed and highlights the difficulties in mapping stratigraphy—and identifying possible sources of suspended sediment turbidity into Esopus Creek and its tributaries.

- | | | |
|-----|-----|--|
| 5.6 | 2.3 | Turn around and drive back down Beaver Kill (Hwy. 212). Pull off on right side of road at second bridge. |
|-----|-----|--|

STOP BK 2. LANDSLIDE IN GLACIAL DEPOSITS. (15 MINUTES) (Location UTM 18 T 561108E 4656095N) A landslide along the right bank of Beaver Kill has propagated as much as 70 m upslope and at least 70 m along the channel downstream of the Hwy. 212 bridge (Figure 22). This site is a dramatic representation of several similar stream bank erosion/hillslope mass failures in the confined portions of the Beaver Kill watershed that seem to be significant sources of bedload sediment supply resulting in numerous gravel/cobble bars that induce further lateral erosion of the channel. An assessment of this stream corridor is planned for late summer 2009.

- | | | |
|-----|-----|--|
| 6.6 | 1.0 | Continue down Hwy. 212 to junction with Old Hwy. 28. |
|-----|-----|--|



Figure 21. Downstream-looking view of suspended sediment from landslides entering Stony Clove Creek and increasing downstream turbidity. Photo by Dan Davis, DEP, March 19, 2009.



Figure 22. Landslide along Beaver Kill. Stop BK2 at bridge. Photo from NYCDEP et al., 2007.



Figure 23. Changes in meander geometry at Beechford. From Miller, 2009.

END OF OPTIONAL SIDE TRIP

- | | | |
|------|-----|--|
| 81.3 | 0.0 | Restart main road log. |
| 81.9 | 0.6 | Continue down Hwy. 212 to junction with Hwy. 28. Turn left toward Kingston. |
| 83.1 | 1.2 | Turn right onto Lower Winnie Road and pull off. Walk over to edge of river bank. |

STOP 8. MEANDER BEND MIGRATION ON ESOPUS CREEK AT BEECHFORD. (15 MINUTES) (Location UTM 18 T 560476E 4652445N) The meander of Esopus Creek just upstream of this location has been migrating eastward and downstream, as recognized from comparison of aerial photographs from 1959, 1980, and 2001 (Miller, 2009; Figure 23). The channel here appears to have been left (east) of the island on the 1903 topographic map. By 1959, the channel had avulsed to the right (west) side of the floodplain, though the original channel was still occupied (at least in part) as a secondary channel, creating a mid-channel island. We do not know exactly when this occurred, nor whether this was a natural avulsion or was induced to protect the Hwy. 28 alignment. However, the main channel has been migrating eastward since 1959, eroding the island and creating a point bar on the right (west) bank. This continuing meander migration likely will result in reoccupation of the original channel and could threaten the Hwy. 28 alignment in the future.

- | | | |
|-------|------|---|
| 83.3 | 0.2 | Continue on Lower Winnie Road back to Hwy. 28 and turn right to continue back toward Kingston. |
| 101.1 | 17.8 | Return to Kingston traffic circle on Hwy. 28 and merge onto NY State Thruway (I-87). |
| 117.3 | 16.2 | Return to New Paltz on NY State Thruway (I-87). Take Exit 18. |
| 119.2 | 1.9 | Take Hwy. 299 west to Hwy. 32 in New Paltz. Turn left, then right at East Entrance to SUNY New Paltz and back to parking lot. |

END OF FIELD TRIP

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Pre-Ottawan (1.09 Ga) Infrastructure and Tectonics of the Hudson Highlands, New York

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INTRODUCTION

This trip focuses on some of the oldest Mesoproterozoic rocks in the Appalachians and the nature of the pre-Ottawan (pre-1.09 Ga) infrastructure of the Hudson Highlands and Manhattan Prong. Zircon geochronology by Aleinikoff (Table 1) is revised from that given in Table 1 of Ratcliffe and Aleinikoff (2008) and is still preliminary. Ongoing geologic mapping east of the Hudson River in the Peekskill, West Point, Lake Carmel, Mohegan Lake, Brewster and Peach Lake quadrangles continues to lead to a reinterpretation and clarification of our former views (Ratcliffe and Aleinikoff, 2001, 2008) as well a new synthesis for the Hudson Highlands east of the Hudson River. After publishing the detailed geology of the Poughquag quadrangle (Ratcliffe and Burton, 1990) and the Oscawana Lake quadrangle (Ratcliffe, 1992), investigations were resumed in 2000 to prepare the previously mapped but unpublished quadrangles for publication.

This trip presents, in preliminary form, much of the recent, (post-2000) information, which helps clarify the age and distribution of Mesoproterozoic basement rocks of the Hudson Highlands and Manhattan Prong and is an abbreviated version of our 2008 NEIGC trip. Please note that stops have been changed and that the route is different.

A remarkable newly recognized aspect of Hudson Highlands geology covered in this trip is the widespread presence of orthogneisses and metavolcanic rocks belonging to the time period between 1400 Ma and 1200 Ma, or during Geochrons 14 and 13 of Rivers (1997). These rocks formed before the Grenville orogeny, defined by Rivers (1997), as ranging from 1.09 Ga to 0.98 Ga, and are coeval with crustal convergence in the Elzevir belt in Ontario and Quebec, at the western margin of the Grenville orogenic belt. We know little about the nature of the deformation that may have affected rocks of the Hudson Highlands during the Elzevirian orogeny except that some were transformed into gneissic rocks prior to about 1230 Ma and multiply deformed after that.

Igneous activity and deformation in the time period of about 1170 to 1145 my ago, at or near the time of the Shawinigan pulse of the Grenville orogeny of Rivers (1997) affected many the rocks of the Hudson Highlands and presumably of the Manhattan Prong although data there are sparse. Penetrative, regionally developed, Y F2 folds of the Hudson Highlands have been mapped in northern New Jersey (Dallmeyer, 1972) and West Point area of the Hudson River (Helenek, 1971; Helenek and Mose, 1984) eastward through the Hudson Highlands (Ratcliffe and Burton, 1990; Ratcliffe, 1992). This deformation resulted in redeformation of older rocks as well as deformation of hornblende granites of the Storm King type, which have an intrusive age of 1174 ± 8 Ma (Table 1). Folds of this generation are folded by the sigmoidal deformation zones associated with the intrusion of the Canopus pluton, and crosscut by mafic dike swarms. Acceptance of an intrusive age of 1143 ± 12 Ma for the Canopus pluton (Table 1) and for the associated mafic dikes, requires that the regional Y F2 dynamothermal event be older than the Ottawan phase of the Grenville orogeny. This is important because many workers think that Ottawan or the terminal phase of the Grenville orogeny was responsible for the major deformational features in Mesoproterozoic rocks of the Appalachians. This view is being reevaluated in the Appalachian outliers of the Grenville and in the Adirondacks, as the recognition of rocks and deformational events at about 1230 Ma and 1150-1134 Ma that are older than the Ottawan are discovered (Ratcliffe and others, 1991; Ratcliffe and Aleinikoff, 2001, 2008). In the Adirondack Highlands evidence for Shawinigan migmatite in the age range of 1180-1160 Ma is widespread (Heumann and others, 2006; Bickford and others, 2007) Throughout the Hudson Highlands and in northern Appalachians, there is a relative scarcity of Ottawan igneous rocks, although ages of metamorphic overgrowths on zircon and of migmatization of older rocks attest to widespread Ottawan remetamorphism.

Rocks intruded between 1134 ± 8 Ma and 1045 Ma, are foliated, as in adjacent areas, such as the New Milford massif (Walsh and others, 2004) and as seen on this trip. However there are abundant data which suggest that the deformational effects in the Hudson Highlands and elsewhere in the northern Appalachians are the result of multiple orogenic events, some predating the Ottawan. On this trip we will visit and summarize information concerning recently dated rocks that are critical to defining the pre-Ottawan history of the Grenville orogen here and elsewhere in the Appalachians of New York and New England.

In addition newly determined SHRIMP U-Pb zircon data, and Ar^{40}/Ar^{39} hornblende ages support late Ordovician to Silurian Taconian overprinting of the basement and cover rocks between 450 to about 443 my ago and suggest that Acadian effects were minimal by comparison.

MESOPROTEROZOIC ROCKS OF THE GRENVILLE OROGEN OF THE NORTHERN APPALACHIANS

We think that is important to place the setting of this field trip within the context of the broader issue namely, the distribution of pre- Grenville Mesoproterozoic rocks throughout the northern Appalachians which has been developing over the past 10 years with the utilization of SHRIMP- based zircon geochronology.

Figures 1, 2, and 3 are generalized maps of the Green Mountain area in Vermont, the Berkshire massif of Massachusetts and the northern Hudson Highlands and adjacent Manhattan Prong. These show the generalized distribution of Mesoproterozoic igneous rocks based on published and, unpublished geochronology largely by Aleinikoff and mapping from 1961 to the present by Ratcliffe, for the new Bedrock Geologic Map of Vermont (in prep), the Bedrock Geologic Map of Massachusetts (Zen and others, 1983) and in the Hudson Highlands east of the Hudson River included in part, in published reports (Ratcliffe and Burton, 1990; Ratcliffe, 1992). Mapping in the Lake Carmel, Brewster, and Peach Lake quadrangles was conducted from 2002 to the present by Ratcliffe. These diagrams do not portray the many subdivisions of the paragneiss sections or the igneous rocks, but only show generalized distributions of igneous rocks or suites that have been dated by recent U-Pb zircon geochronology studies. The level of understanding among these maps is uneven as geochronologic data for the Berkshire massif are limited as compared to the other two areas.

Evidence for very old, preGrenville rocks in the Grenvillian orogen of the Appalachians has been growing in recent years but the distribution of these rocks is uneven. In the Green Mountain massif and eastern domes of Vermont, igneous rocks in the age range of 1400 to 1300 Ma (Geon 14) are common and constitute about 25% of the area of the Mesoproterozoic there (Figure 1). Much, but an unknown amount of the paragneiss sequence in the Green Mountains also is older than 1300 Ma so the percentage may be much higher than shown in Figure 1. Ratcliffe and others (1991) and Ratcliffe (1997) showed that plutonic rocks of the College Hill pluton dated at about 1230 Ma cross cut the older 1400 to 1300 Ma gneisses, indicating Elzervirian deformation in the Green Mountains as well as classical Grenvillian events formed between 1200 and 1000 Ma which affected all the Mesoproterozoic rocks.

The Oldest Rocks - Geon 14

Five new U-Pb ages (Table 1) and geochemistry provide evidence for calcalkaline igneous activity between 1349 and 1328 Ma in the Fordham Gneiss and in the Hudson Highlands east of the Canopus shear zone. Broad areas of biotite-tonalitic, trondhjemitic and biotite granite gneisses (Yrg) extend northward from the main belt of Reservoir Gneiss near Peekskill, into the Poughquag, Lake Carmel and northern part of the Brewster quadrangles (Figure 2). This belt may contain separable types of granitic gneiss but distinguishing them on maps has proven almost impossible thus far. Tonalitic gneisses at Cat Hill (Stop 7) and trondhjemitic gneiss (Stop 1) tend to occur in the western areas and more granitic rocks are to the east and to the south. Granodioritic to trondhjemitic Reservoir Gneiss at Peekskill (Stop 8) has a U-Pb zircon age of 1338 ± 9 Ma and we think that map continuity suggests that similarly old rocks are present in the Brewster area.

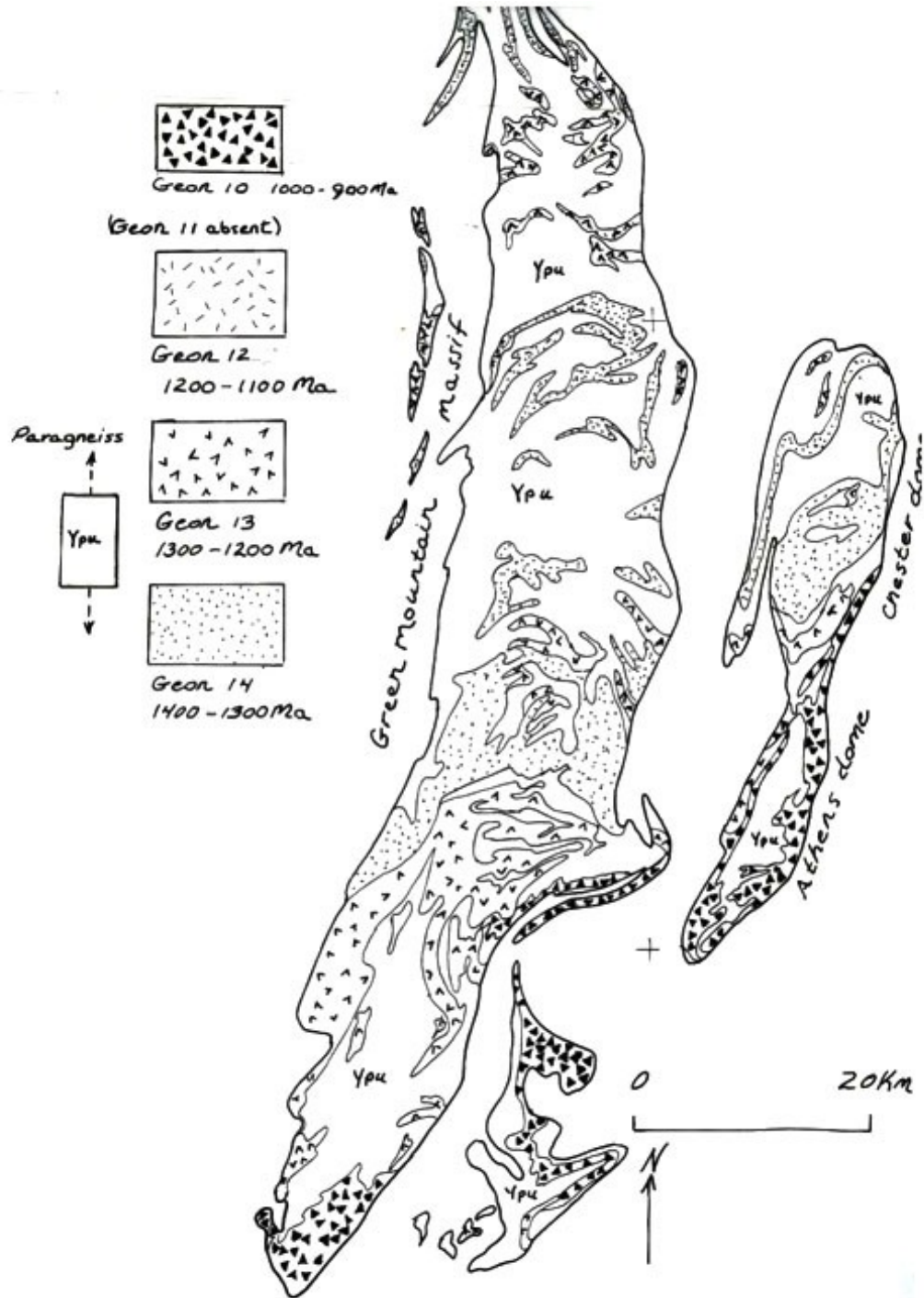


Figure 1. Simplified map of Mesoproterozoic igneous rocks of the Green Mountain massif and eastern domes of Vermont showing broad belts which are characterized by pre-Grenville intrusive rock of Geons 14, and 13, and Grenville intrusive rocks of Geons 12, 11 (minimally represented), and post Ottawan rocks of the Cardinal Brook Intrusive Suite of Geon 10. Paragneiss Ypu contains metasedimentary as well as probable metavolcanic rocks that could range from Geon 14 to Geon 13.

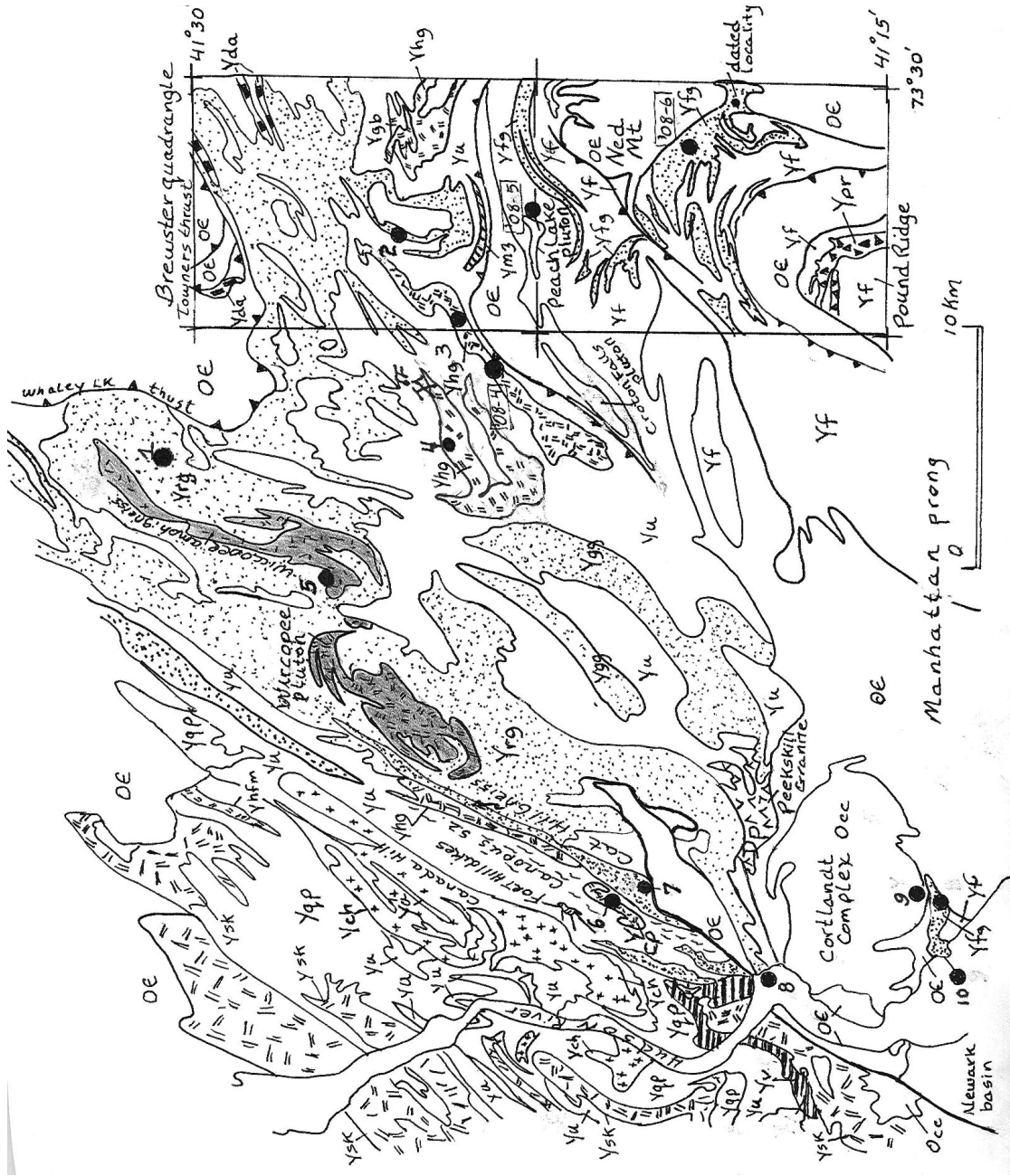


Figure 2. Generalized map of the Hudson Highlands of NY and Connecticut modified from Figure 1 of Ratcliffe and Burton (1990) showing Mesoproterozoic intrusive rocks. Many faults have been eliminated for clarity and new data from the Brewster and Peach Lake quadrangles have been added based on recent mapping by Ratcliffe. Key to unit designators: Geon 14- Yrg Reservoir Gneiss (Stops 1, 2 and 10); Ych Cat Hill Gneiss; Ygg granitic gneiss undated; Yfg granitic gneiss of the Fordham Gneiss (Stops 6 and 9); Shaded- metapyroxenite, metagabbro, metadiorite and amphibolite gneiss of the Wicopsee pluton and amphibolite belt; Yqp quartz plagioclase gneiss and leucogneiss west of the Canopus shear zone; (continues)

Figure 2. (continued) Geon 13 Ymz monzonite at Joe's Hill, Brewster, NY; Yv metavolcanic and amphibolite Camp Smith; Geon 12-Ysk hornblende granite of the Storm King type; Yhg hornblende granite and diorite at Brewster (Stop 3); Yhfm ferrodiorite and ferromonzonite, Canopus pluton (Stop 12) Fort Hill intrusion and dikes, identified on map; Ygb gabbro and Yhg southeastern part of Brewster quadrangle; Geon 11-Yda Danbury Augen Gneiss; Geon 12 or 11-Ych Canada Hill Granite age uncertain; Geon 10 -Ypd Pound Ridge Granite; Yu paragneiss and undated biotite-quartz-plagioclase gneiss undifferentiated largely Geon 14 and 13 but age uncertain; Yf Fordham Gneiss undifferentiated.



Figure 3. Simplified map, showing the distribution of granitic gneisses of Geon 12 in the Berkshire massif and the post Grenville Stamford Granite of the Cardinal Brook Intrusive Suite, of Geon 10. Ytg Tynningham Gneiss, Ygg other similar unmapped granitic gneisses from Zen and others (1983). Sample locations are those used for Rb/Sr whole rock dating of Mose, reported in Ratcliffe (1975).

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Sample No.	Unit	Age-type	Overgrowth ages
		GEON 11	
*NM100	Danbury augen granite	1045 ± 7.9 Ma (7/6)	0.965 Ga
*NM772	Layered biotite gneiss	1048 ± 11 Ma (7/6)	0.98 Ga
*NM628	Biotite granite gneiss	1050 ± 14 Ma (7/6)	1.03, 0.98 Ga
*NM576B	Migmatite	1057 ± 10 Ma (7/6)	1.0 Ga
		GEON 12	
Br2549	Hornblende granite (Stop 3)	1134 ± 8 Ma (7/6)	
CP-3	Canopus pluton (Stop 6)	1143 ± 12 Ma (CA)	~1.0 Ga
JA-PK	Storm King granite	1173 ± 7 Ma (CA)	~1.14 Ga
		GEON 13	
Br2215	Quartz monzonite Joe's Hill (Brewster)	1240 ± 7 Ma (7/6)	
PK9-04	YV aplitic gneiss	1238 ± 8 Ma (7/6)	1.22, 0.99 Ga
		GEON 14	
*NM174	Pink granite gneiss	1311 ± 7 Ma (7/6)	1, 0.98 Ga
Br1176	Fordham Gneiss (Stop 6-2008) Scott Ridge (Peach Lake)	1328 ± 8 Ma (CA)	
Br576A	Reservoir Gneiss (Luddingtonville) (Stop 1)	1338 ± 9 Ma (7/6)	1.26, 1.0 Ga
PK11, 876	Cat Hill Gneiss (Peekskill) (Stop 7)	1333 ± 9 Ma (CA)	
PK1-05	Reservoir Gneiss (Stop 8)	1338 ± 6 Ma (CA)	~1.26, 0.98 Ga
PK7A	Fordham Gneiss (Stop 10)	1349 ± 7 Ma (CA)	~1.12, 1.02 Ga

Table 1. U-Pb Zircon ages of Mesoproterozoic rocks of the northern Hudson Highlands, Manhattan Prong, and New Milford massif (NM*) (Walsh and others, 2004), CN = concordant age (7/6)=Pb/Pb age

Mafic rocks of the Wicoppee pluton (Figure 2) and associated amphibolite gneiss (Stops 5a, b) appear to be intruded by granitic components of the Reservoir Gneiss (Ratcliffe, 1992) and may be the source of some of the xenoliths or enclaves in the granitic gneiss. Biotite granitic gneiss similar to the Reservoir Gneiss but undated occur in a southern zone (labeled Ygg). Granitic gneiss from the New Milford massif in Connecticut has limited areal extent and a U-Pb zircon age of 1313 Ma (Walsh and others, 2002) and closely resembles rocks of the Reservoir Gneiss in the Brewster area. Biotite granite gneiss (Yfg) of the Fordham Gneiss is wide spread in the southern part of the Brewster area and in the core of the belt of Fordham Gneiss in the center of the Peach Lake quadrangle (Stop 6 of Ratcliffe and Aleinikoff, 2008). This rock has a U-Pb zircon age of 1328 ± 8 Ma. Granitic gneiss of the Fordham at stop 10 has a U-Pb zircon age of 1349 ± 7 Ma. In the New Jersey Highlands, Aleinikoff and Volkert (2007) and Volkert and Aleinikoff (2007) report ages of intrusive and volcanic rocks of the Losee Suite as about 1230-1300 Ma.

Areas shown as quartz-plagioclase gneiss (Yqp) in Figure 2 in the West Point and Peekskill quadrangles extend into New Jersey and are Losee-like, although they have not been dated with certainty and probably belong to Geon 14 or 13.

Rocks of Geon 13

The College Hill pluton and other dated but unnamed granite gneisses from the Green Mountains of Vermont have intrusive ages of about 1230 Ma (Ratcliffe and others, 1991) and produce migmatite near their borders and cross cut older folds and gneissosity (Ratcliffe, 1997). Extensive areas of biotite granitic rocks correlated with the dated localities occur throughout the central and southern Green Mountains and eastern domes and northern part of the Green Mountain massif are thought to be of the same approximate age based on geologic mapping. In the Hudson Highlands there are only two dated rocks belonging to Geon 13 (Figure 2). One is a small late cross cutting monzonite to granodioritic mass (Ymz on Figure 2) that intrudes across gneissosity in older biotite-hornblende migmatitic gneiss on Joes Hill in the southern part of the Brewster quadrangle. This rock provides evidence for gneissosity of Elzevirian age in the older gneisses. The other rock is aplitic gneiss (Yv in Figure 2) on the east side of the Hudson River and on the north flank of the Dunderberg antiform near Camp Smith. This rock is interlayered either by interbedding or by lit par lit injection with numerous continuous mapped amphibolitic gneisses. The felsic component has a zircon SHRIMP age of 1238 ± 7 Ma.

Rocks of Geon 12

Granitic rocks dated between about 1170 to 1135 Ma are present in Green Mountains, Chester and Athens domes, Berkshire massif (Figure 3), and in the Hudson Highlands of New York and New Jersey. Locally these intrusive rocks include gabbro, hypersthene ferrodiorite, ferromonzonite, ferrodiorite dikes, biotite granite augen gneisses, and hornblende granite gneisses of the Storm King Granite type in the Hudson Highlands.

The Tyringham Gneiss in the Berkshire massif and related granitic rocks (Figure 3) make up about 35 per cent of the Berkshire massif, largely in the central and southern area. The age of the Tyringham Gneiss from samples on Beartown Mountain and near Tyringham was determined by whole rock Rb/Sr dating by Doug Mose at about 1170 Ma (Mose in Ratcliffe, 1975). Ratcliffe and Zartman (1976) reported upper intercept ages of approximately 1040 to 1080 for the Tyringham on Beartown Mountain. A more recent SHRIMP zircon age from a sample of Tyringham Gneiss in the East Lee quadrangle of about 1180 Ma (Karabinos and others, 2003) and tends to support the original Rb/Sr age, as does a SHRIMP zircon age from the Tyringham on Beartown Mountain (Aleinikoff, unpublished data). If the correlations of Ratcliffe are correct as shown on the Bedrock Geologic Map of Massachusetts (Zen, 1983), then most of the granitic rocks of the Berkshire massif belong to Geon 12. Because the Tyringham Gneiss intrudes all mapped units within the Mesoproterozoic of the Berkshire massif it establishes a minimum age for that sequence (Ratcliffe and Zartman, 1976), which is generally consistent with the inferred age of very similar paragneiss sequences in the Vermont and in the Hudson Highlands and Fordham Gneiss.

In the Hudson Highlands, hornblende granite of the Storm King type (Ysk on Figure 2) was intruded at about 1174 ± 8 Ma and was deformed along with country rock before intrusion of the Canopus pluton and attendant mafic dikes (stop 6) and hornblende granite and diorite at Brewster at stop 3 (see below).

In the Hudson Highlands the 1143 ± 12 Ma Canopus pluton (Stop 6) was intruded during significant right slip deformation on vertical faults of the Canopus fault zone. Northeast trending mafic dikes and northwest trending folds document transpressional tectonics at 1143 Ma. Hornblende granite gneiss and aplite near Brewster NY (Stops 3 and 4) intrudes aluminous and calcareous paragneiss to produce magnetite-sulphide mineralization of the Brewster magnetite district and crosscuts an older gneissosity. Hornblende-pyroxene pegmatite and hornblende aplite are common as the border phase of the hornblende granite. The zircon SHRIMP age of the hornblende granite at Brewster is 1134 ± 8 Ma. Associated mafic dikes and diorite – gabbro stocks as well as the hornblende granite cross cut older gneissic structure, older isoclinal folds, and high-grade mylonitic shear zones, as the Canopus pluton and related dikes do. Similar mineralization at Lake Mahopac and at Tilley Foster suggests that this is the age of magnetite and calc silicate mineralization there as well. We think that these two dated rocks (Canopus and hornblende/diorite at Brewster (stops 3 and 4) are attributable to the same tectonic events of the Shawinagan pulse of the Grenville orogeny, i.e., not the Ottawa phase of the Grenville orogeny.

In the Green Mountains of Vermont (Figure 1) rocks of Geon 12 only are known from the northern most part of the massif where foliated coarse K-feldspar augen gneiss crosscut a gneissosity in older gneisses and has SHRIMP age of about 1172 Ma. Hornblende gabbro and diorite are rocks associated with the augen gneisses in the northern most

part of the massif as well.. These observations suggest some high-grade metamorphism and deformation occurred before 1172 Ma.

The data from the Green Mountains and Hudson Highlands support important deformation of the basement rocks after 1230 Ma and prior to 1172 Ma in the Green Mountains and after 1173 Ma but before 1143 Ma in the Hudson Highlands. This is approximately the time of the Shawinigan pulse of the Grenville orogeny, following the usage of Rivers (1997).

Rocks of Geon 11 and 10

Except for widespread pegmatitic muscovite –biotite granites and zircon overgrowth ages there is little evidence for igneous activity in the Green Mountains of Vermont during Geon 11 (Figure 1) and only one of these rocks is well dated. Post Grenvillian rapakivi plutons (Karabinos and Aleinikoff, 1990)of the Cardinal Brook Intrusive Suite (Ratcliffe, 1991) occur in the southeastern Green Mountains, in the core of the Chester and Rayponda domes and at the north end of the Berkshire massif (Ratcliffe and others, 1991). This belt of intrusives lies in a distinct NNE trend, which crosses both the individual mapped units in the Grenville orogen but also east west trending bands of Geon 13 and 14 rocks (Figure 3). In the New Milford area and in the adjacent Hudson Highlands megacrystic Danbury augen gneiss was intruded at 1045 ± 8 Ma roughly coincident with widespread migmatization (Walsh and others, 2004). Biotite granite pegmatite, and migmatization accompanied formation of the Canada Hill pluton formerly thought to be intruded at 1014 Ma. Recent analyses on zircon from the main Canada Hill pluton however, have cast some doubt on this date and the true age of the Canada Hill is uncertain at present (Aleinikoff, 2008, personal communication to Ratcliffe) and may be much older than previously thought.

Mesoproterozoic Paragneisses

Abundant quartzite, aluminous metasedimentary rocks, dolomitic and calcitic marbles, calc-silicate rocks are wide spread and interlayered with biotite- quartz-plagioclase paragneiss and amphibolite through out the Green Mountain Mesoproterozoic sequence. Amphibolites of both igneous and metasedimentary origin occur. These biotite- quartz-plagioclase gneisses appear to host metasedimentary rocks of different age. A similar sequence occurs in the Berkshire massif. Because of the uncertainty of origin and the abundant interbedded metasedimentary rock little geochronology has been attempted on these rocks.

Paragneisses in the Hudson Highlands appear to be more localized because of the greater abundance of plutonic rocks and or thicker metavolcanic units, principally massive garnet-plagioclase quartz gneisses. Calc-silicate and sulfidic metapelitic rocks, quartzite and prominent phlogopite -diopside dolomite or calcite diopside marble, white coarse- grained dolomite marble and diopside-scapolite quartzite are distinctive but thin units. Paragneisses are intruded by igneous rocks of Geon 12 and 13, but there are no clear cross cutting examples of the oldest rocks, those of Geon 14, with paragneiss. As we will see, all of the old gneisses always contain mafic biotite-hornblende amphibolite, biotite leptite and dioritic hornblende gneiss inclusions. Some are clearly dikes, although boudinage commonly makes contact relations indistinct. Pyroxene and hornblende form at the contact between mafic masses in the old felsic rock suggesting either intrusive contact reactions or later reaction with younger more felsic sweat outs.

GENERAL GEOLOGY OF THE TRIP AREA

Figure 4 shows the general geology of the northern Hudson Highlands as presented in 1990 following extensive detailed mapping by Ratcliffe, Henry Helenek and William Burton following several years of mapping in the late 1980's. Figure 4 from the Poughquagh Quadangle map (Ratcliffe and Burton, 1990) shows the interpretation of the eastern Hudson Highlands at that time. They recognized that a large area east of the Canopus fault consisted of leucocratic gneisses of the Reservoir Gneiss, ranging from tonalite to granite gneiss in a belt previously identified as the Reservoir Granite by Berkey and Rice (1921).

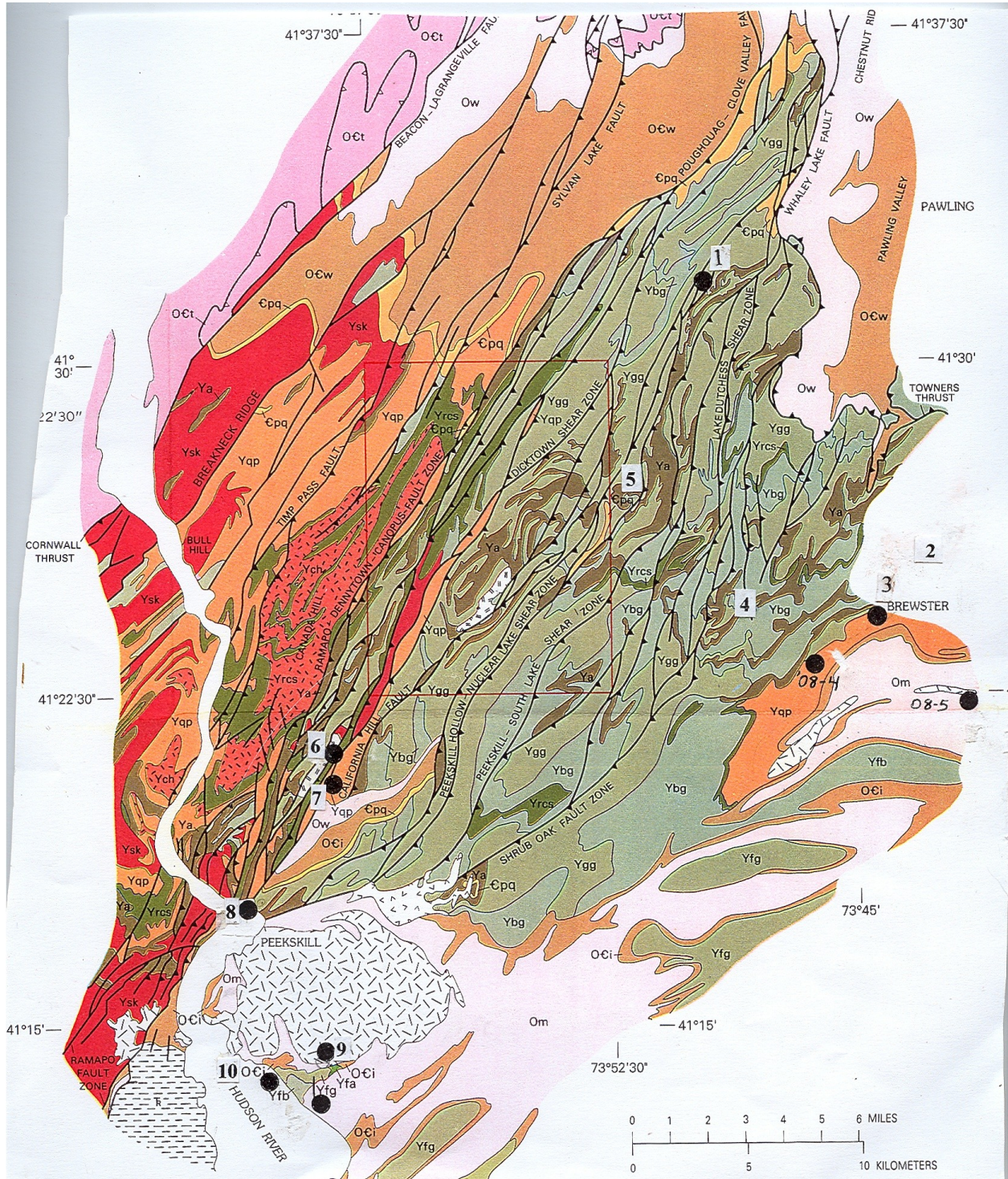


Figure 4. (continues)

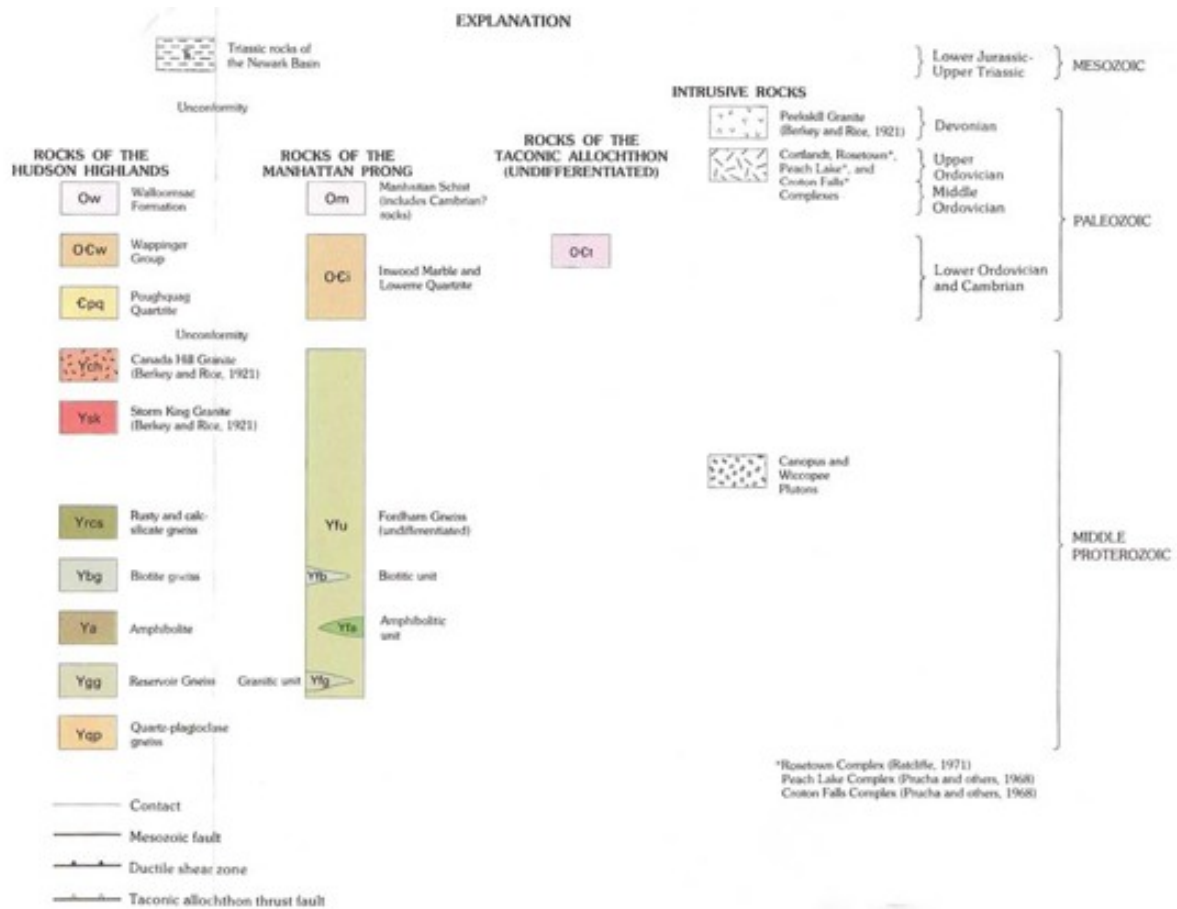


Figure 4. (continued) Geology of field trip area, showing stop locations in the Hudson Highlands from Ratcliffe and Burton (1990). Additional stop locations (see Figure 2).

Rocks of the Hudson Highlands and Manhattan Prong consist of a basement complex of Mesoproterozoic age making up the Highlands gneisses and Fordham Gneiss. These older rocks are now known to range in age from approximately 1350 Ma to about 1000 Ma. In the Hudson Highlands and northwestern part of the Manhattan prong, the Poughquag Quartzite (or Lowerre Quartzite) unconformably overlies the basement. In eastern exposures near Towners and in much of the Manhattan prong, the base of the quartzite becomes increasingly micaceous and feldspathic and contains beds typical of the Dalton Formation which regionally has been interpreted to range down in to the Neoproterozoic; for example see a description of the type Dalton in the Pittsfield East quadrangle of Massachusetts (Ratcliffe, 1984). One important aspect of the Dalton Formation as it is mapped throughout Vermont, Massachusetts, New York and Connecticut, is the dominance of quartzofeldspathic schists and granofels and the absence of volcanic rocks (Zen and others, 1983, Rodgers, 1985). At the south end of the Green Mountain massif and on the northern end of the Berkshire massif the Dalton Formation passes laterally into more albitic and aluminous rocks of the Hoosac Formation which contains significant amounts of mafic volcanic rocks having chemical signatures of rift basalts as well as oceanic MORB basalts (Ratcliffe and Armstrong, 1999) In the Hoosac Formation and in the facies of Dalton passing into the Hoosac, quartzites of the Poughquag and Lowerre types are minor and feather out eastward as the entire sequence becomes older, and more oceanic. This transition between Hoosac and Dalton may be expressed in parts of the Manhattan prong by rocks not marked by quartzite beds like those of the Poughquag and hence rocks mapped as Lowerre Quartzite are now referred to as Dalton Formation in much of the Manhattan prong (Rodgers, 1985) although the rocks by and large are not typical of the Dalton Formation. Regionally rocks of the

Hoosac Formation or correlatives, such as the Cannan Mountain Schist, and or Manhattan schist (units B and C of Hall, 1968) occupy thrust sheets overlying rocks as young as late Ordovician. As a result, rocks resembling the combined Dalton- Hoosac Formation can occur in two structural positions, either resting on the older basement unconformably in which case they are referred to as Dalton or if in thrust sheets as Manhattan Schist. This points to a basic flaw in the use of the term Manhattan Schist to refer only to the allochthonous part of the Manhattan Schist of Hall (1968). This approach produces both a forced stratigraphy and a forced structural solution.

Shelf carbonate rocks of Cambrian through Medial Ordovician age comprise the Wappinger Group of the Hudson River valley and some inliers in the Hudson Highlands and are comparable to carbonate rocks of the Inwood Marble of the Manhattan Prong. Overlying the Inwood or Wappinger are rocks now referred to as the Walloomsac Formation which locally in the Manhattan prong, is referred to as Manhattan A (of Hall, 1968). In the Hudson River valley, Massachusetts and Vermont, rocks of the Walloomsac range from upper Medial Ordovician to Late Ordovician. The depositional age range of the Walloomsac, where controlled by fossils, is about 465 to 445 Ma and marks the pre-Taconian flysch deposits formed just prior to the collisional phase of the Taconian orogeny. The basal Walloomsac Formation contains thin limestones and limestone boulder conglomerate in western exposures but regionally the facies change going eastward. Schistose marbles, phlogopitic feldspathic dolostone, quartzites replace limestone eastward as the formation bites deeper into underlying carbonates and locally rests on Mesoproterozoic rocks. In general the quartzitic, dolomitic and feldspathic rocks are developed regionally in the eastern outcrop belts from western Massachusetts and Connecticut in a line extending from the southwest corner of Massachusetts through the west flank of the Housatonic massif to the west side of the Pawling valley and from there southeastward to the western margin of the Manhattan prong at Crugers, NY, on the Hudson River near Stop 7.

Distinction Between the Manhattan Prong and Hudson Highlands

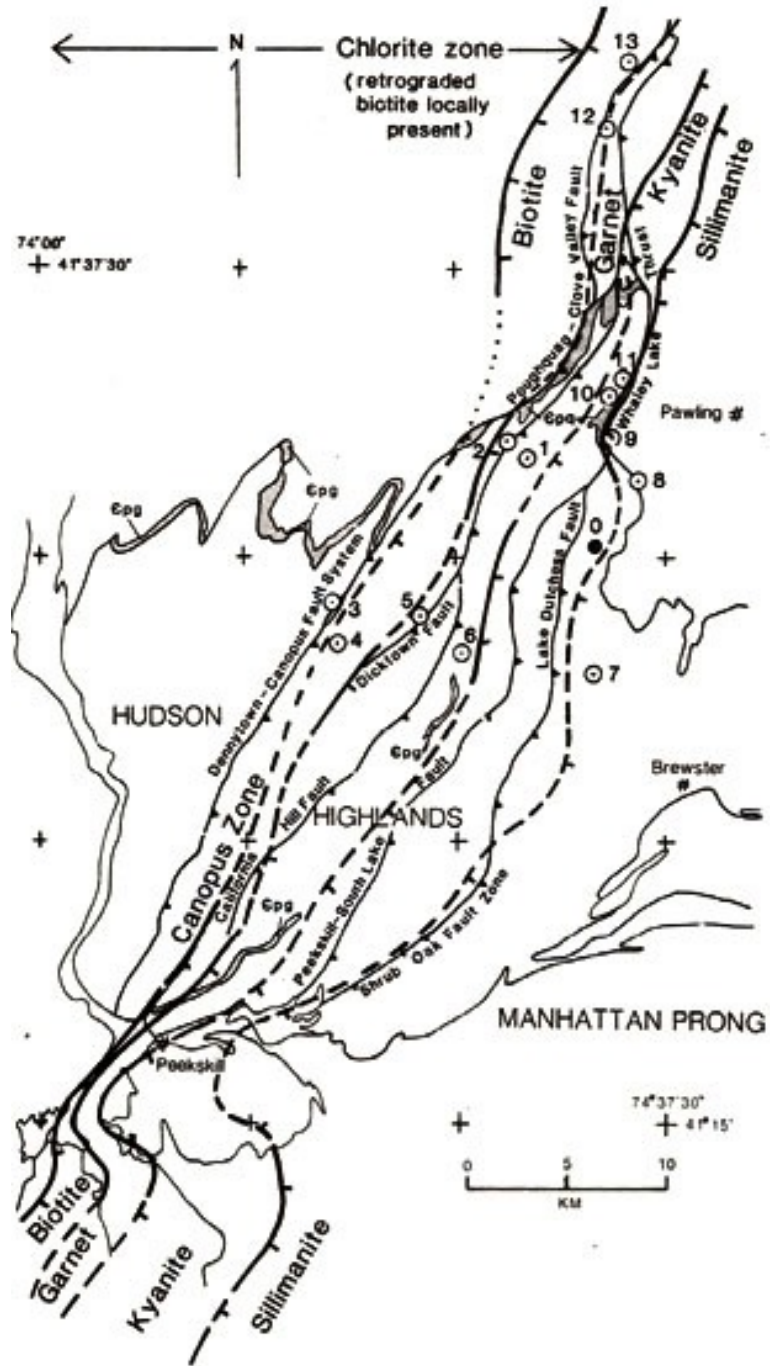
These two terms are largely geographic and signify no real differences in the ages of rocks or tectonic history. Conventionally the boundary between the Manhattan Prong and the Hudson Highlands extends from Peekskill to Brewster NY where the contact alternately varies from faulted segments to segments having normal sedimentary contacts from basement up into quartzite or other rocks of the Paleozoic cover sequence (see Figure) and more recent modifications in Figure 4. Semiductile and brittle steeply northwest dipping and strike slip faults characterize segments between Somers and Brewster and eastward.

Deformational History and Structure

This is really not the focus of this trip so this guide is brief on the subject. The area from Brewster lies within either sillimanite- K-spar or sillimanite- muscovite grade of regional metamorphism that is Taconian and the grade decreases westward toward the Hudson River to staurolite grade and to biotite grade near stop 11 and 12. Regional distribution of isograds (Figure 5) coincides with the trend and spacing of a ductile right- oblique thrust and shear zone complex which was responsible for much of the basement response to Taconian deformation (Ratcliffe and Burton, 1990). Igneous rocks of the Cortlandt complex intrude this deformational zone and Barrovian metamorphic field gradient and cross cut folds and foliations in the country rocks as well. The zircon SHRIMP age from the monzonite of the Cortlandt complex at Stop 7 of 447 ± 2 Ma is in our judgment the most accurate age obtained for the Cortlandt thus far and we accept this as the age of intrusion. This age is in good agreement with the age of the highest grade sillimanite-K-spar grade metamorphism of 444 - 443 Ma obtained from the New Milford area for formation of migmatite and intrusion of the Candlewood Granite (Walsh and others, 2004), as well as regional Ar/Ar chronometry of the Taconic metamorphism at about 450 Ma as summarized by Hames and others (1991) and by Ratcliffe and others (Trip B4, this volume). Dietsch and Aleinikoff (2006) report U-Pb zircon ages of 437 ± 4 Ma for migmatitization in the core of the Waterbury dome, Connecticut, supporting the concept of extended Taconian metamorphism. Ar/Ar data on hornblende from the region support relict Taconian slow cooling ages from near the Cortlandt complex and a broad area of Devonian cooling ages in the rest of the Manhattan prong as far east as Long Island Sound (Dietsch and others, 2006). Two muscovite cooling ages, one from the Peekskill granite in garnet zone (Kunk, unpublished data, 2007) and one from a granite dike near Purdys in the sillimanite zone (Dietsch and others, 2006), support late Devonian cooling through muscovite blocking temperature of 350 C at about 350 Ma. These data suggest that the dominant metamorphism was Taconian but that elevated temperatures occurred in the Acadian over

much of the area east of the sillimanite zone of metamorphism until the Medial to Late Devonian.

Intrusion of the Devonian Peekskill Granite (Ratcliffe and others, 1982) has been confirmed by a concordant SHRIMP age of 380 ± 4 Ma. A slightly younger age of 374 Ma was determined by Douglas Mose (Ratcliffe and others, 1982) using the Rb/Sr whole rock technique. Reexamination of the contact relations of the Peekskill granite and its relation to movement on the Peekskill fault confirms the earlier conclusion of Ratcliffe and others (1982) that the intrusion overlapped movement on the fault and that Devonian activation or reactivation of this fault is probable. The muscovite cooling age above indicates the cooling from granite intrusive temperatures of 650° to 700° C took about 30 million years.



ISOGRADS IN RELATION TO SHEAR ZONES

Figure 5. Taconian isograds (Ratcliffe and others, 1985) Stop numbers refer to 1985 field trip.

ROAD LOG

Road log mileages and directions start at Stop 1 driving time from New Paltz about 55 mins. We will leave parking lot at New Paltz at 7:30 and start at Stop 1 at 8:45.

Directions to Stop 1

From parking lot turn N on Manheim Blvd. toward Pond Rd., 0.5 mi turn right on Main St., 0.9 mi. merge onto NY Thruway South. Follow 16 mi. take Exit 17 toward I-84. Take I-84 east to toward Newberg/ Beacon. Cross Hudson Newburgh Bridge. Continue on I-84 to Exit 17 at Luddingtonville Rd. and park opposite the west bound entrance ramp.

General descriptions of the geology we are traversing from stop to stop are shown in italics.

A generalized description of the pertinent geology from the Newburgh Bridge to Stop 1

At the crossing of the Hudson on your right is Storm King Mountain held up by Mesoproterozoic hornblende granite gneiss of the Storm King Granite. Although not dated here at the type locality this rock is identical to that dated (1174 ± 8 Ma) at Dunderberg Mountain to the south. Here it occupies the core of a large isoclinal fold of the regional YF2 generation (Dodd, 1965). These folds are characterized by penetrative high grade mineral fabric and strong northeast plunging fold axes and hornblende lineation. This fold involves hypersthene- quartz- plagioclase leucogneiss that is very similar to the 1337 Ma trondhjemite gneisses at Cat Hill at Stop 7. This old gneiss or its equivalent is similar to some of the Losee Gneiss of New Jersey. Quartz- plagioclase gneiss like this cores YF1 folds through out the Hudson Highlands east of the River and at Bear Mountain.

The Storm King Granite is thrust northwestward over Cambrian and Ordovician cover sequence rocks on the Cornwall thrust that dips at least 45 degrees to the south east at the Hudson River crossing of the Delaware aqueduct (Berkey and Rice, 1921, plate 48). On the east side of the Hudson River from a point just north of the bridge graywackes of the Austin Glen Formation and the upper parts of the Taconic allochthon form the foot wall of a faulted klippe of Storm King and other Mesoproterozoic gneisses and quartzite cover. The klippe is truncated by high angle normal faults (east side down) of probable Mesozoic age on both sides. This complex of faults is termed the Beacon-Lagrangeville fault. (Ratcliffe, 1985). The sole of the klippe likely is the Cornwall thrust and the eastern carbonates part of the normal cover sequence. The Poughquag Quartzite rests unconformably on the basement above the Cornwall thrust in this klippe.

The route of I-84 travels east of the klippe on Wappinger Group carbonate rocks. Several exposures of variegated dolostones of the Pine Plains Formation can be seen along the thruway before the crossing of route 9 at Fishkill. At this point a large fault- controlled valley extends southeastward into the Highlands to the Hudson River north of Cold Spring and crosses the River along the southeast side of Storm King Mountain. I originally placed a southeast dipping thrust fault along this contact in 1980 (Figure). Subsequent studies indicate that the dominate fault structures in this part of the West Point quadrangle are typical of Mesozoic normal and oblique-normal high angle faults. Chlorite and hematite coated, slickensided surfaces are prevalent, as is microbreccia and zeolite mineralization. In addition, the Cambrian Poughquag Quartzite and lamprophyre dikes of the Cortlandt intrusive event are faulted. The mineralization and fractures styles are characteristic of the Ramapo and other fault rocks found in coring of Mesozoic faults (Ratcliffe, 1980). Distinctive mylonitic rocks of high grade associated with Mesoproterozoic faulting or the retrogressive phyllonites typical of the Taconic and or later thrust events as illustrated in Ratcliffe (1985) are lacking or not well developed here. This leads me to conclude that the dominant faulting here is Mesozoic although some Paleozoic thrust faulting may be present as well.

The trend of these faults project southwestward but die out before Lake Tiorati in the southwestern corner of the Popolopen Lake Quadrangle mapped by Dodd (1965) and more recently examined by Gates (1999). Gates places a major ductile shear zone along the western side of Lake Tiorati that he extends northeastward along the contact of the major rock units mapped by Dodd (1965) without any offset of rock units. In support of the age of dextral shear-

ing and the age of this event Gates and others (2004) cite a 1008 ± 4 Ma zircon age from the Lake Tiorati "diorite". The origin of the Tiorati "diorite" has long been in dispute (Lowe 1950; Dodd, 1965). Dodd mapped this rock as his amphibolite II and was uncertain as to its origin.

Mapping in the area of Lake Tiorati and adjacent Thiells quadrangle in 1980 by Bill Burton and myself confirmed the map patterns and fault locations of Dodd, and we do not agree with the mapping of Gates (1999). Instead of being a definable plutonic rock the Lake Tiorati "diorite" occurs as widely separated pods of hornblende- rich rock developed where layered amphibolitic gneiss is intruded by pegmatite. Rather than being a definable intrusive rock the Tiorati "diorite" occurs as inclusions in pegmatite and does not intrude other rocks. Excellent xenoliths in pegmatites show variably changing textures from foliated amphibolite to granite-saturated hornblende gneiss to hornblende pegmatite and pyroxene- hornblende pegmatitic dioritic rock. Tiorati-like coarse pyroxene- hornblende- plagioclase pegmatite occurs as pods and lenses nowhere greater than 50 m wide and commonly less the 10 m and at several localities as xenoliths in pegmatite less than a meter wide and is always in contact with pegmatite, aplite or granite gneiss. This rock occurs at as many as 8 different structural levels west of Lake Tiorati as well as in gneisses east of the proposed Lake Tiorati shear zone in the core of the YF3 synform mapped by Dodd. For these reasons I do not accept Gates and others (2004) interpretation that the zircon age dates the time of dextral strike slip faulting but interpret the 1008 ± 4 Ma zircon age as the time of pegmatite generation and regard it as a minimum age for the tectonic fabrics in the gneisses placing this in relatively good agreement with other dated pegmatites of the area. (Volkert, Zartman and Moore, 2005).

East of route 9, I-84 crosses well layered paragneisses that are in contact with Storm King Granite on the slopes above the cuts. The Poughquag Quartzite rests directly on the basement rock with only a thin gritty pebble conglomerate locally developed and a normal succession of Wappinger Group carbonate rocks plunges northward on the noses of basement cored antiforms and broad synforms. It is important to note that throughout all of the Hudson Highlands and immediately adjacent Manhattan prong where the contact beneath the Cambrian quartzite can be located closely, there is no intermediate stratigraphic unit of possible Neoproterozoic age except for some sections that contain feldspathic quartzite attributable to the Dalton Formation. (see Ratcliffe and Burton, 1992 for example).

At the crossing of the Taconic Parkway and from that point northeastward numerous thrust faults marked by low grade phyllonites displace Mesoproterozoic rocks and carbonate rocks along the various splays of the Ordovician-reactivated Canopus and related shear zone (Figure 4). Within this belt evidence of Mesoproterozoic motion synchronous with intrusion of the composite ferrodiorite-monzonite Canopus pluton and dike swarm at about 1143 Ma (note the revision in age as reported by Ratcliffe and Aleinikoff (2001, 2008)) and which corrects the older estimates based on Rb/ Sr and early U/Pb zircon ages reported in Ratcliffe and others (1972) and in Ratcliffe (1992).

The new data Ratcliffe and Aleinikoff (2001, 2008) and this guide book, do not agree in timing of right-lateral deformation proposed for the Reservoir and other faults of about 1008 ± 4 Ma and presented in a series of papers (Gates, 1999; Gates and others, 2004). We now regard much of the deformation in the Hudson Highlands as older than the Ottawan and to be Shawinigan. That does not preclude renewed activity in the Ottawan but does urge caution in accepting the Ottawan escape tectonic model of Gates and others (2004). A similar caution is warranted from consideration of the geology of the Green Mountains of Vt and the Berkshires of Massachusetts as shown in Figures 1 and 3, where continuity of structure precludes an extension of the Piseco Lake shear zone and related structures to the east as required by the model of Gates and others (2004). Of some importance is the distribution of the post-Ottawan 955 Ma rapakivi granites of the Cardinal Brook Intrusive Suite. Notable also is E-W tectonic grain of the Berkshire massif that continues southward into the Hudson Highlands. The consistent northeast plunges of the YF3 folds in the Hudson Highlands is attributable to refolding of overturned, but not recumbent, early YF2 folds.

Road log starts at Stop 1 Exit 17 from I-84 Luddingtonville Rd.

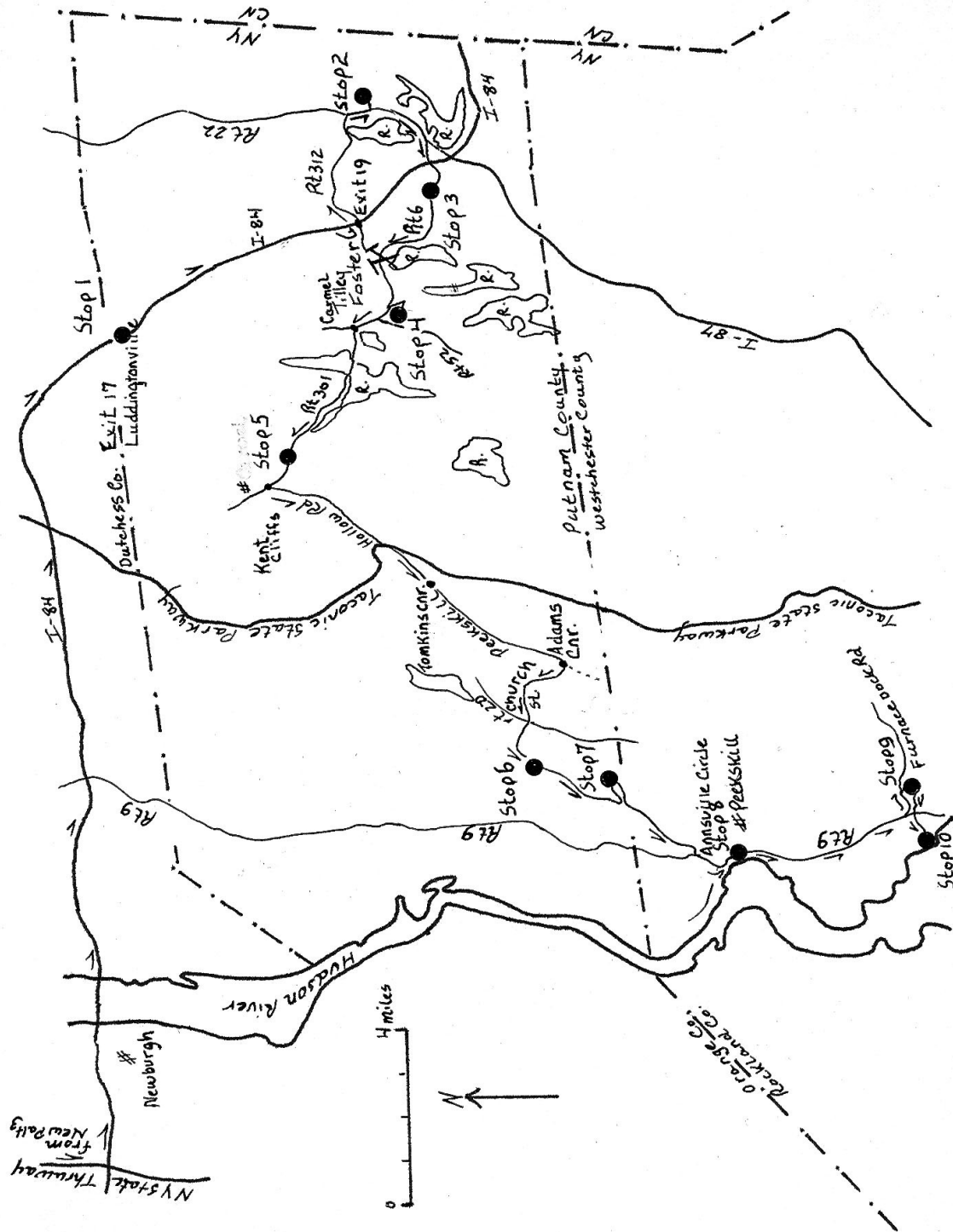


Figure 6. Map showing route and stop locations and major highways.

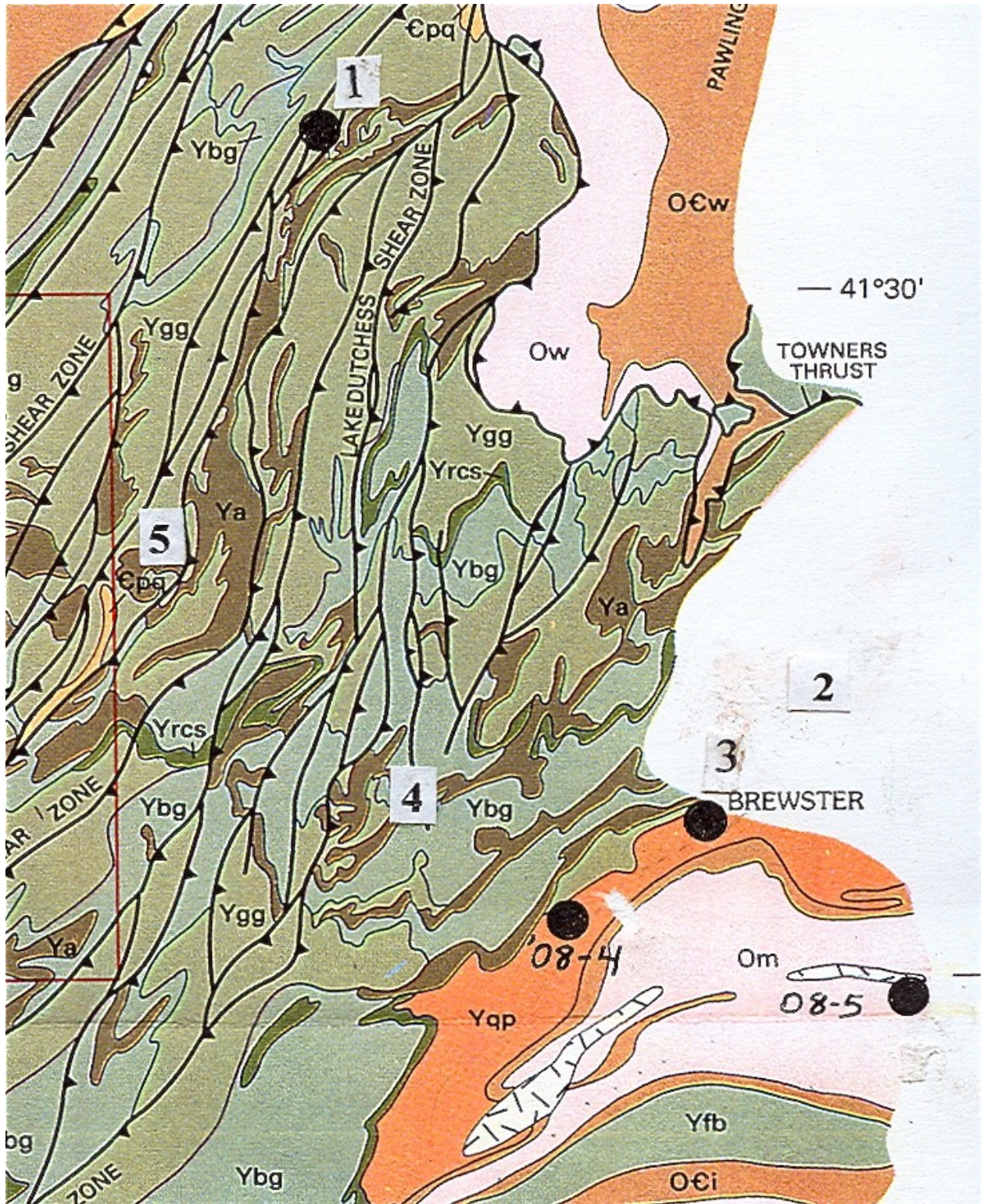


Figure 7. Blow up of the eastern part of Figure 4 showing location of Stops 1, 2, 3, 4 and 5.

Stop 1. Reservoir Gneiss--amphibolite inclusions I-84 Exit 17 Luddingtonville Rd. Poughquag quadrangle
 STAY OFF THE INTERSTATE PLEASE. Park by the overpass and walk along west entrance lane.

This belt of trondhjemitic gneiss forms the eastern end of a wide belt of Reservoir Gneiss (of Ratcliffe and Burton, 1990) that extends from near Cat Hill Stop 7 through alternate Stop 5b into the Poughquag quadrangle and the contiguous Brewster quadrangle (Figure). A SHRIMP zircon age of 1333 ± 6 Ma has been obtained from this outcrop. Ages of overgrowths on the zircon are 1.2 and 1.0 Ga indicative of Elzeviran and Ottawan zircon growth. Sweat outs of more granitic composition are common and are shown by coarser grain size in indefinite irregular pockets in which and older gneissosity is overprinted. Abundant mafic inclusions choke this rock in most exposures. Reaction rims around the mafic inclusions contain pyroxene, hornblende, and biotite and are gradational with the host rock. It is likely that these reaction zones and the pegmatitic sweat outs are the result of Ottawa metamorphism.

Normative An- Ab-Or compositions of some the felsic rocks belonging to Geon 14 and those of Geon 13 are plotted in Figure 8 and range from trondhjemitic to granite. Fields for the Reservoir Gneiss at Peekskill (stop 8), at Cat Hill (Stop 8), in the Oscawana Lake area (Stop 5b) and here at Luddingtonville (Stop 1) show the range of compositions in the Reservoir Gneiss in the westernmost belt closest to the Canopus deformation zone.

These trondhjemites and granites all have distinctively high La/ Ybcn values, steep middle to light rare earth slopes, and flat, negative to positive HREE slopes comparable to that shown for the Cat Hill Gneiss (Figure 10). The samples of the dated Fordham Gneiss have very similar patterns. Plots of Rb vs Nb+Y (Figure 9) show the similarities among the dated rocks as well. The filled dots are samples of Yrg in the Brewster area that are correlated with the main Reservoir Gneiss belt on the basis of continuity on the ground between the two areas. We can not be certain that all rocks shown as Reservoir Gneiss in the Brewster area are in fact the same as these rocks are not dated.

Mafic inclusions in this outcrop and in all stops in the rocks of Geon 14 may be inclusions and or dikes as will be discuss further at Stop 2. The amphibolites here show no clear indications of having been dikes but that cannot be ruled out. Note the felsic sweat outs and pegmatites near the wall of the “inclusions” as well as the pyroxene- hornblende reaction zones, which grade into the bordering rocks.

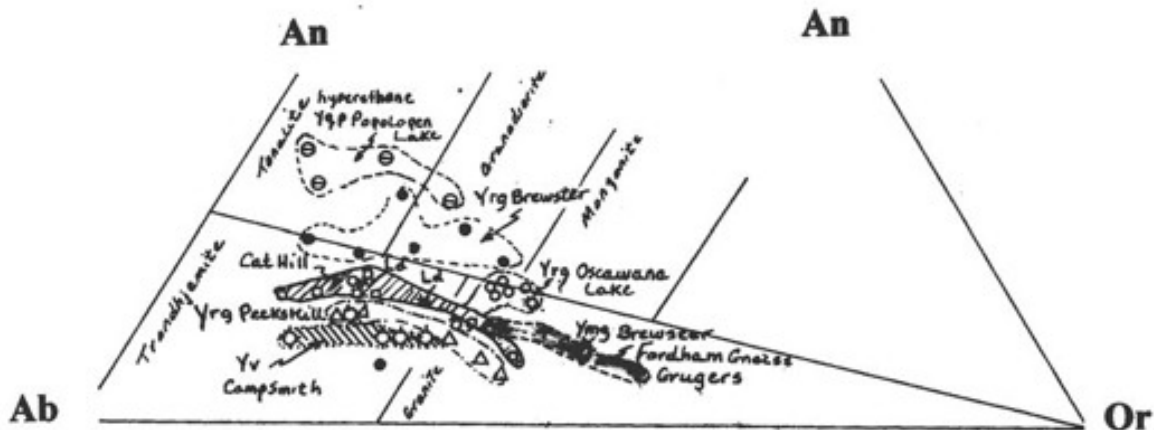


Figure 8. Normative An-Ab-Or plots of dated rocks of Geon 14 and 13. Solid dots are samples of Yrg from the Brewster area; squares Cat Hill Gneiss (Stop 7); triangles Reservoir Gneiss Peekskill (Stop 8); stars Reservoir Gneiss (Stop 1); pluses Fordham Gneiss at Stop 10 and from Scott Ridge in the Peach Lake quadrangle; open circles Reservoir Gneiss in Oscawana Lake quadrangle (Stop 5b). Yv – aplitic gneiss at Camp Smith, Peekskill quadrangle zircon SHRIMP age of 1238 ± 7 Ma; Ymz- monzonitic dike Joes Hill Brewster Zircon SHRIMP age of 1240 ± 7 Ma and Yqp- hypersthene- quartz-plagioclase gneiss Popolopen Lake quadrangle.

Enter I-84 southbound toward Brewster; turn at exit 19 onto 312 heading east—large road cut in dark gray, biotite-rich quartz plagioclase gneiss and hornblende aplite follow 312 to Rt. 22 about 3.3 mi, turn south (right) on Rt. 22, go 0.7 mi to left, entrance to small professional office complex park by far eastern end of lot. Walk up hill to north 250ft to exposure.

Stop 2. Snake Hill - Granitic gneiss typical of Brewster belt and mafic dikes

Steeply dipping biotite granite gneiss typical of the vast majority of granitic gneiss and associated migmatitic lit par lit contact zone on Joes Hill. Gneissic granite like this is crosscut at the west end of Joe Hill by a monzonite dike that has a zircon Pb207/206 SHRIMP age of 1240 ± 7 Ma. The granite gneiss here is tentatively assigned to the old gneiss belt and is judged to be 1330 Ma or older. Fine -grained foliated mafic dikes cross cut the gneissic structure and are fairly abundant on this hill and in areas to the east. Locally coarser grained diorite to gabbroic masses also cross cut the older gneisses and these rocks pass into finer grained rocks like the dikes seen here.

Figure 12 shows one of a series of meter-wide dikes petrographically like the small ones above. On Joes Hill these dikes cross cut high-grade gneissosity and intrude across mylonite zones. Although apparently straight walled, they are reformed by isoclinal folding. Both the dikes and the older gneissosity are cut by foliated granitic pegmatite of probable Mesoproterozoic age. Based on these observations, the mafic dikes, like that figured below must be Grenvillian and perhaps older than the Ottawa phase of the Grenville orogeny. In this regard these dikes resemble the dikes of the Canopus pluton (stop 6), which also cross cut an older gneissosity.

Chemically, the dikes in question are somewhat like those of the Canopus pluton as illustrated in Figure 13, and to

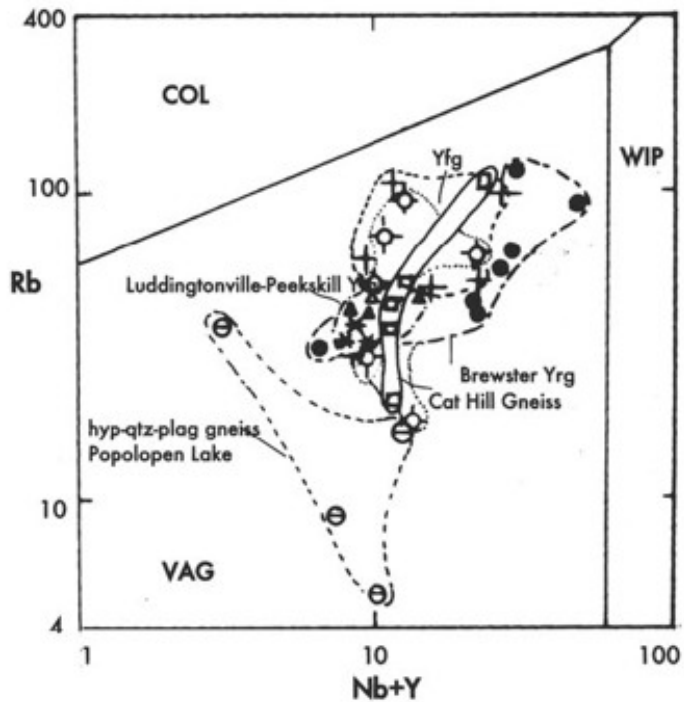


Figure 9. Rb vs Nb+Y plots of rocks of Geon 14 and 13 symbols as in Figure 8.

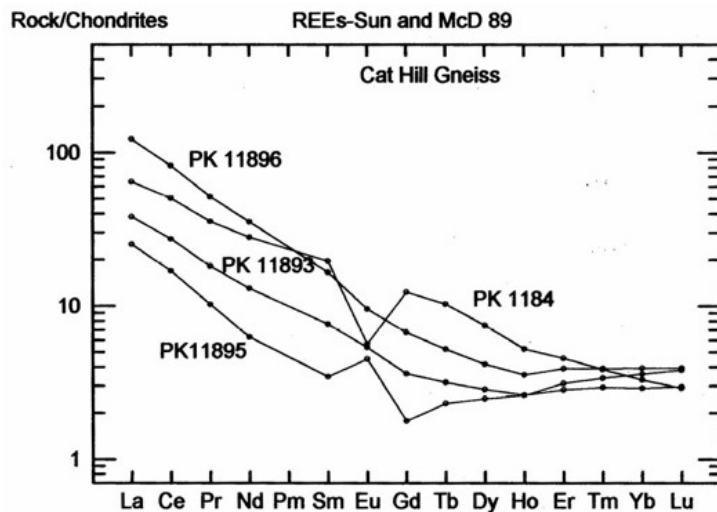


Figure 10. Chondrite-normalized Rare Earth Element patterns of samples of the Cat Hill Gneiss (Stop 7). These are representative of most samples of the Reservoir Gneiss.

dioritic rocks at stop 3. The diagram shows a similarity in chondrite LREE enrichment trends (but not in absolute values) and similar range in Ta -Yb ratios. The plots do not establish identity of the various rocks only suggest that possibility.

The dike in Figure 12 is a typical example. It cross cuts migmatitic gneiss that at this locality is intruded by a monzonitic granite that has a SHRIMP zircon age of 1240 ± 7 Ma (Table 1). The age of these dikes is younger than 1240 Ma. They could be the same age as the diorites associated with the dated hornblende granite (1134 ± 8 Ma) at stop 3 and with dikes of the Canopus pluton (Stop 6A, B) that are probably approximated by the Canopus monzonite zircon SHRIMP age of 1143 ± 12 Ma. The mafic dikes of the Canopus and elsewhere in the Brewster area cross cut coarse gneissosity in the 1340-1330 Ma rocks; both are intruded by foliated pegmatite.

Alternately these dikes could be Neoproterozoic. However major and trace element characteristic differ from Neoproterozoic dikes found in the Hudson Highlands (Ratcliffe, 1978), which are typical transitional rift basalts.

The clearly cross cutting mafic dikes and mafic and isoclinally folded “inclusions” in the granitic gneisses of the Brewster area are indistinguishable chemically and petrographically from one another. This observation plus the fact that individual dikes like that in Figure 12, which cross cut a gneissosity can be traced within outcrops into isoclinally and cross foliated zones suggests that many of the inclusions or indefinite dikes like those at stop1 may all be the same age. If so, there is a widespread mafic dike event at about 1143 to 1134 Ma. that post dates an earlier Shawinagan deformation.

Leave parking lot head south, turn left on Rt. 22 toward Brewster, then exit 22 to right for Rt. 202 to Brewster, immediately at “T” intersection bear left following 202. In one mile turn right and cross over East Branch of Croton River, proceed through Brewster parallel to RR tracks. In 0.5 mi turn left on Rt. 6, cross over RR tracks and immediately branch right onto side street go up hill 0.3 mi



Figure 11. Two fine grained mafic dikes cross the coarse ropy gneissosity of granitic gneiss typical of the granitic gneisses in the Brewster quadrangle and judged to be correlative with the Reservoir Gneiss.



Figure 12. Mafic dike cross cutting gneissosity in migmatite gneiss At Joes Hill, Brewster.

and park in parking lot opposite funeral home.

Stop 3. Unnamed 1134 ± 8 Ma hornblende granite gneiss, dioritic gneiss and sulfidic schist screen, Brewster, N.Y.

Hornblende-biotite granite gneiss 140 feet wide, on the southeast end of this NW-SE vertical cut becomes very hornblende rich and passes into a 15-ft-wide pyroxene hornblende pegmatite near the small screen of sulfidic paragneiss. The hornblende granite has a 1134 ± 8 Ma SHRIMP age. North of that dioritic gneiss about 35 ft thick passes northward into a hornblende-rich margin that grades into a second 40 ft-thick zone of granite gneiss. That granite gneiss grades upwards into diorite gneiss at northwest end of the exposure. The felsic and mafic components grade into each other at the margins and appear to be coeval. In this characteristic these rocks resemble rocks of the Canopus pluton (Stop 6A, B) and have essentially the same age as the felsic component of the Canopus which has a concordant zircon SHRIMP age of 1143 ± 12 Ma.

We regard the felsic and more mafic rocks as part of the same igneous suite. Near contacts with sulfidic schists and calc-silicate rocks hornblende-pyroxene pegmatite and a white hornblende-spotted aplite carrying magnetite is common. This is the common association for the Brewster magnetic deposits, which are developed on strike to the southwest at the borders of xenoliths included in the hornblende granite and along the wall of this intrusion. All the rocks are foliated but tend not to be gneissic like the older rocks. Similar dioritic intrusives occur to the north in Brewster and preserve coarse diabasic texture and relict coarse - plagioclase flow structure. The mafic dikes we have seen may also be the same age.

Return to Route 6, turn right (N), enter Lake Carmel quadrangle, in 2 miles pass Tilley Foster Mines at Rt. 311 intersection.

Calc-silicate rock, marble and magnetite are developed here at the contact with the same hornblende granite gneiss dated at Stop 3.

Continue on Rt. 6 past Tilley Foster 1.4 miles to Intersection of Rt. 35 Stoneleigh Rd Hannerford Market on right. Turn left on Stoneleigh. In 0.1 mile turn right into Retreat Condominiums and proceed 0.9 mi. up hill to left to construction site.

Stop 4. Hornblende Granite gneiss like that at Stop 3

NOTE: on Figure 4 this location does not correctly show the hornblende granite which was discovered after the figure was published. See Figure 2 for a more accurate map. This illustrates an important point namely that telling the

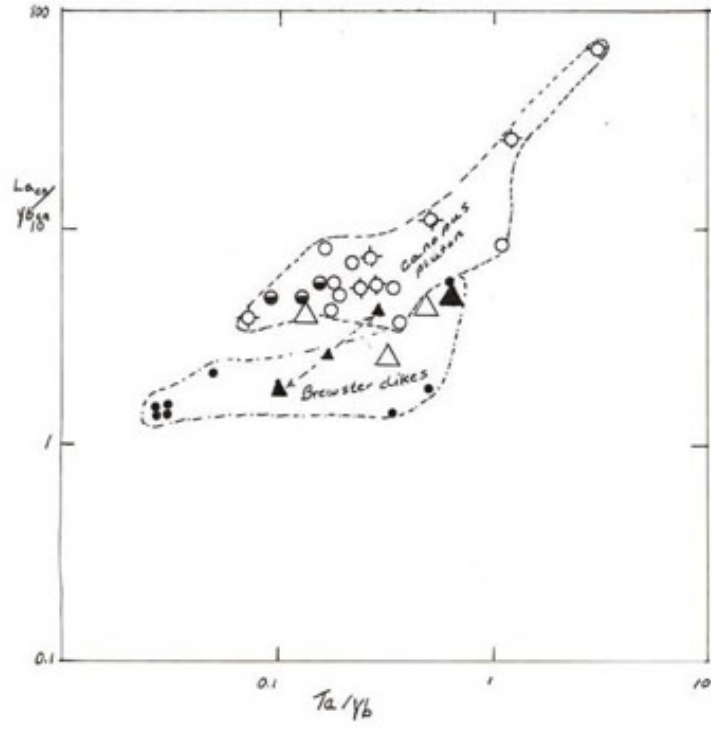


Figure 13. Chondrite normalized La/Yb vs Ta/Yb plot for mafic rocks and dikes of the Canopus pluton compared to mafic dikes at stop 2 and to amphibolite inclusions at stop 1 shown by small joined triangles. Large unfilled triangles- gabbro-diorite stocks in Brewster, half filled dots mafic diorites at stop 3 associated with dated hornblende granite.

old 1340-1300 Ma granite gneisses from the 1134 Ma ones is not easy. I expect to complete mapping in the Lake Carmel quadrangle this Fall and undoubtedly contacts will shift.

Well linedated and foliated hornblende granite gneiss and aplite cuts gneissosity in older well layered biotite and hornblende-quartz-plagioclase gneisses. Both rocks are folded into tight folds containing axial surfaces of N 65 W 45 NE and a strong lineation plunging to the east and northeast. These relationships support the idea that the 1135 Ma hornblende granites cross cut older structure just as the Canopus pluton Stop 6A, B does.

Excellent exposures in the Croton Reservoir (Stop '08-4 of Ratcliffe and Aleinikoff, 2008) are no longer available as the reservoir has now been refilled and the exposures covered. Observations there showed that:

1. The hornblende granite is foliated but does not have the same coarse ropy- textured gneissosity of the older 1340 Ma gneisses seen at Stop 2.
2. Xenoliths of foliated gneisses occur within the younger hornblende granite gneisses and aplite
3. The coarse and fine grained variants of the hornblende granite and hornblende pegmatite occur at contacts with the country rocks and an older gneissosity is truncated by the younger granites.
4. Contacts with rusty sulfidic paragneisses developed abundant magnetite and sulfides just as the wall rocks of the larger bodies of hornblende granite do in the Brewster magnetite district do.

Return to Stoneleigh Rd. turn left to light at at Route 6 and turn left toward Carmel in 08 mi. intersection with Rt. 52 bear right, in 0.3 mi. light and intersection with Rt. 301 turn left and follow across West Brach Reservoir 1.3 mi. Marker west end of Reservoir.

General geology. We are in the center of the Lake Carmel quadrangle headed west across a particularly well developed sequence of Mesoproterozoic paragneiss, calc-silicate rock, marble, thin quartzites and schist. We will cross two of the thickest belts of amphibolite in the Hudson Highlands that are mappable into the Poughquagh quadrangle to the north. These amphibolites are true amphibolites, unlike so many of the rocks mapped as amphibolites by myself and others in the Highlands. These are composed of >70% hornblende rather than the more common type of "dioritic" amphibolites which commonly contain equal amounts of plagioclase and hornblende. This belt of coarse to fine amphibolite is associated with pods of serpentinite, anthophyllite schist and gabbroic to dioritic igneous rocks. The serpentinites are in contact with paragneisses, diopside bearing quartzites, and carbonate containing ovoidal pods of what probably are serpentinitized fosterite knots.

In 3.6 mi from the causeway pass Dixon Rd where we pass into the western and thickest of the amphibolites. I mile past Dixon Rd. dam of the Boyds Corner Reservoir and pull off on right.

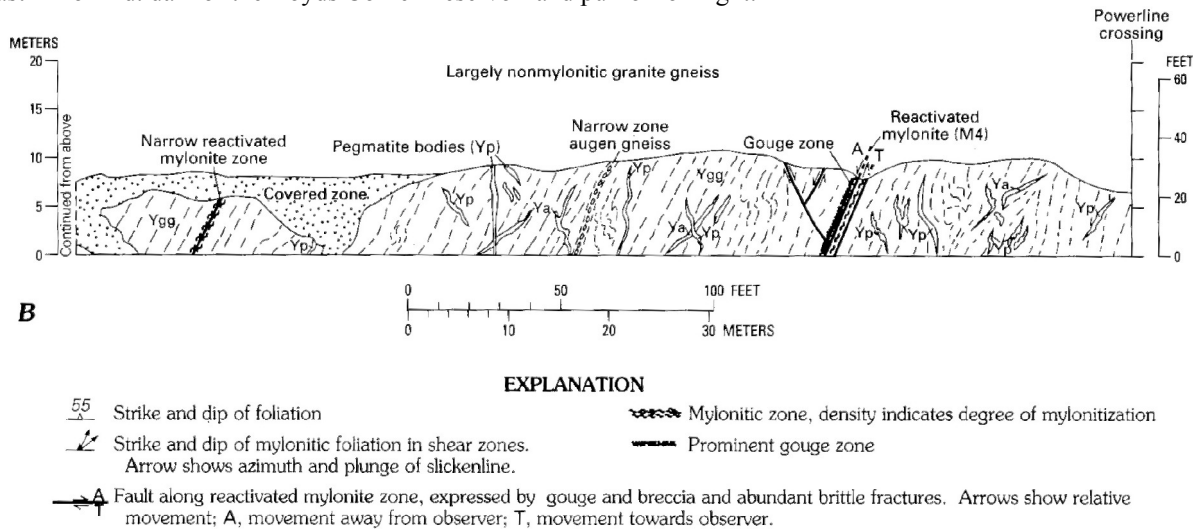


Figure 14. Cross section of the northern end of a long road cut in Reservoir Gneiss along Rt. 301 opposite Boyds Corner Reservoir, Mt Carmel quadrangle, NY. See Ratcliffe (1992) for more complete discussion.

Optional Stop 5A. Coarsest amphibolite perhaps related to the Wiccopee pluton of the Oscawana Lake quadrangle (Ratcliffe, 1992) discussion of amphibolite chemistry

Leave parking area and proceed 0.5 mile to pull off on the right opposite the reservoir.

Optional Stop 5B. Reservoir Gneiss at its type locality discussion of regional tectonic of the Oscawana Lake- right lateral deformation zone (see Stops 1 and 7 for chemistry discussion)

These road cuts in the Reservoir Gneiss represent the type locality of the gneiss and were visited on Stop 6 of Ratcliffe and others (1985) and is in part reproduced from Figure 23 of Ratcliffe (1992). Aside from illustrating the

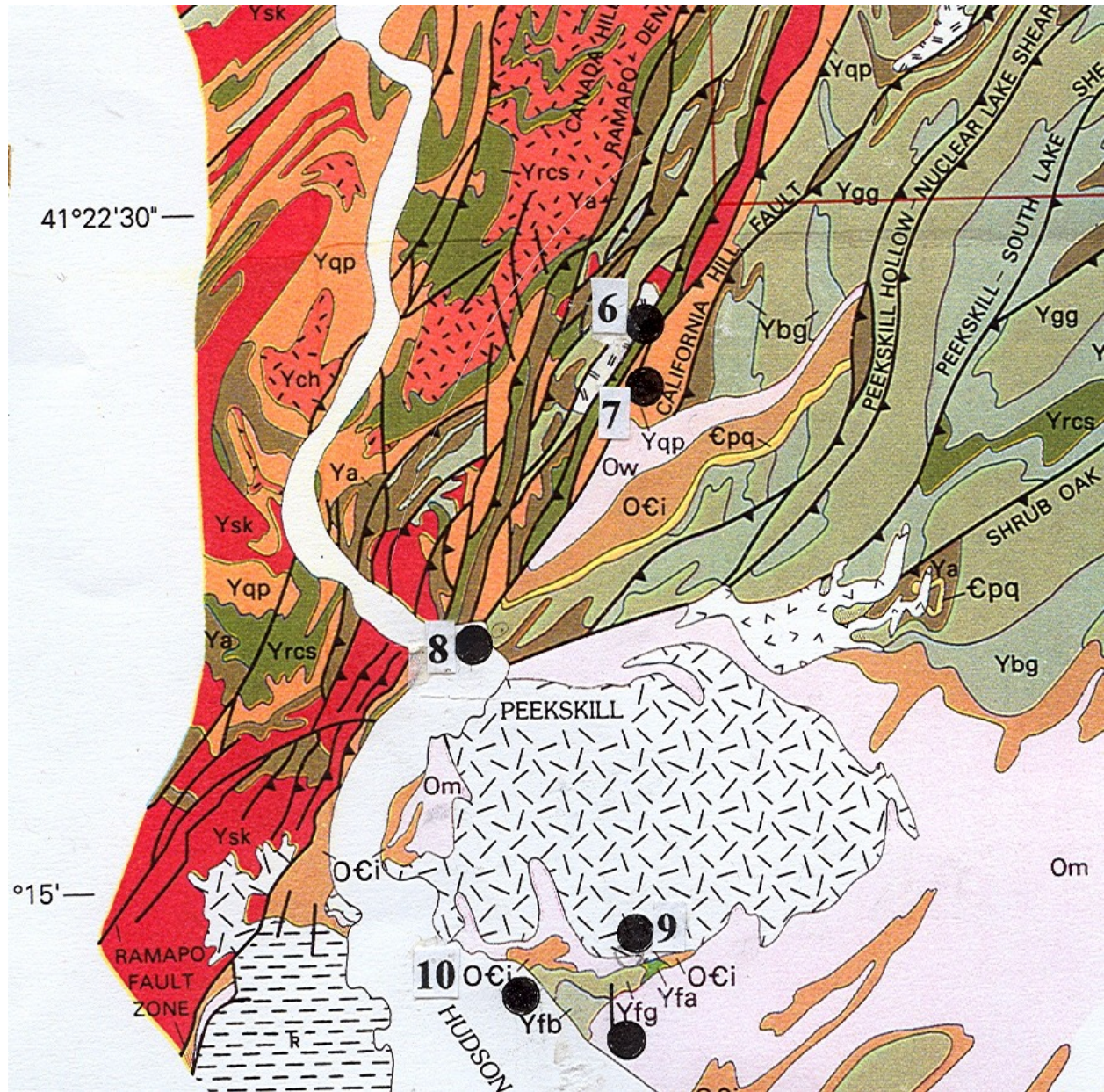


Figure 15. Blow up of a part of figure 4 showing the location of Stops 5, 6, 7, 8, 9 and 10.

type Reservoir Granite of Berkey and Rice (1921), these crops contain mylonite zones typical of the Paleozoic fabrics found to mark the larger shear zones at biotite and higher grades. Here evidence for reactivation as Mesozoic normal faults may be seen. A similar observation will be made at Stop 8.

Continue west on Rt. 301, enter the Oscawana Lake quadrangle.

0.3 mi. intersection Peekskill Hollow Rd on left- turn left, immediately take the right "Y" up hill (not the left) staying on Peekskill Hollow Rd.

We are traveling down the Peekskill Hollow- Nuclear Lake-shear zone that marks the eastern side of the large right-lateral sigmoidal deformation zone in the center of the Oscawana Lake quadrangle. Although the sigmoid is Mesoproterozoic (Ratcliffe, 1992), the California Hill and this shear zone are Paleozoic based on the retrogressive phylionite there and offsets of Paleozoic rocks along their lengths and from dating of syntectonic biotite in the mylonite of the California Hill shear zone that has a 436 ± 3 Ma Ar/ Ar age (see Figure 0 in Ratcliffe, 1992). The sigmoidal deformation of the gneisses however is Mesoproterozoic based on the mapping of non rotated metadiabase dikes that have the chemistry of Neoproterozoic rift basalt. (Ratcliffe, 1992). Based on this reasoning, the California Hill shear zone has both a Mesoproterozoic and a Paleozoic history, as do the faults bordering the Canopus Valley to the west.

In 4 miles cross under the Taconic Parkway and go 5 miles to Adams Corner light and turn right on to Rt. 22 west.

At Adams Corner we will turn west and cross the end of the Annsville syncline which probably has a Mesozoic fault long its northwestern flank. See discussion Stop 7.

Follow County Rt. 22 for 2.2 miles to Crofts Corner and Rt. 20 intersection.

Two miles south on Rt. 20, in front of the Putnam Valley Elementary School, there are superb outcrops of Mesoproterozoic shear zone rocks invaded by aplites that are also highly deformed and crossed by Mesoproterozoic pegmatites supporting intense Mesoproterozoic deformation east of and parallel to the main Canopus shear zones. This is approximately on the California Hill fault zone and supports Mesoproterozoic inheritance here.

Cross over Rt. 20 straight onto Cimmarron Rd..

Continue on Cimmarron Rd. down the hill into the Canopus Valley that is underlain by coarse phlogopitic- feldspathic- scapolite dolomite marble, quartzite and retrograde diopside calc- silicate rock. This is the thickest belt of Mesoproterozoic marble and calc-silicate rocks in the Hudson Highlands of NY.

2.2 miles from the Rt. 20 intersection on Cimmeron Rd and on the west side of the Canopus Valley is Indian Hill Rd on the west side of the Canopus Valley.

Immediately above and on the south side of Indian Hill Road is an old magnetite mine developed in marble and calc-silicate rock just off the north end of the Canopus pluton. Here aplite and pegmatite that makes the border phase of the Canopus , contain large 1-2 cm clots of magnetite where it intrudes hornblende granite gneiss like the Storm King Granite. This lit par lit border zone will be seen at Stop 6b where it is caught in the retrogressive shear zone along the west side of the Canopus Valley at the east edge of the pluton.

Cimmarron Rd. has now changed name to Sprout Brook Rd.

From the Indian Hill Rd. intersection proceed 0.7 miles south to abandoned asphalt drive on the right opposite mail box 484—pull into the drive and try to get all cars off the road. Limited parking here Aleinikoff, 2008.

Stop 6A. Canopus pluton diorite, dikes and screen of Mesoproterozoic gneiss

At the end of the driveway are excellent new exposures of the dioritic rocks of the pluton on the right and left sides of the cut. Note the well defined areas of monzodiorite in diorite in the block to the left (S), the diorite- a screen of isoclinally folded country rocks exhibiting the YF2 structure and the flow layered diorite on the right (N). Two mafic dikes cross cut both the gneiss inclusion and the flow layered diorite. The observations here point to composite intrusion of the dioritic phases of the pluton. Dike swarms in the vicinity of the pluton cross cut YF2 structures in the gneisses. Some dikes are as much a 0.5 km long and are present a many as 2 km west of the pluton. These dikes are chemically and petrographically like the dikes in the Brewster area seen at Stop 2.

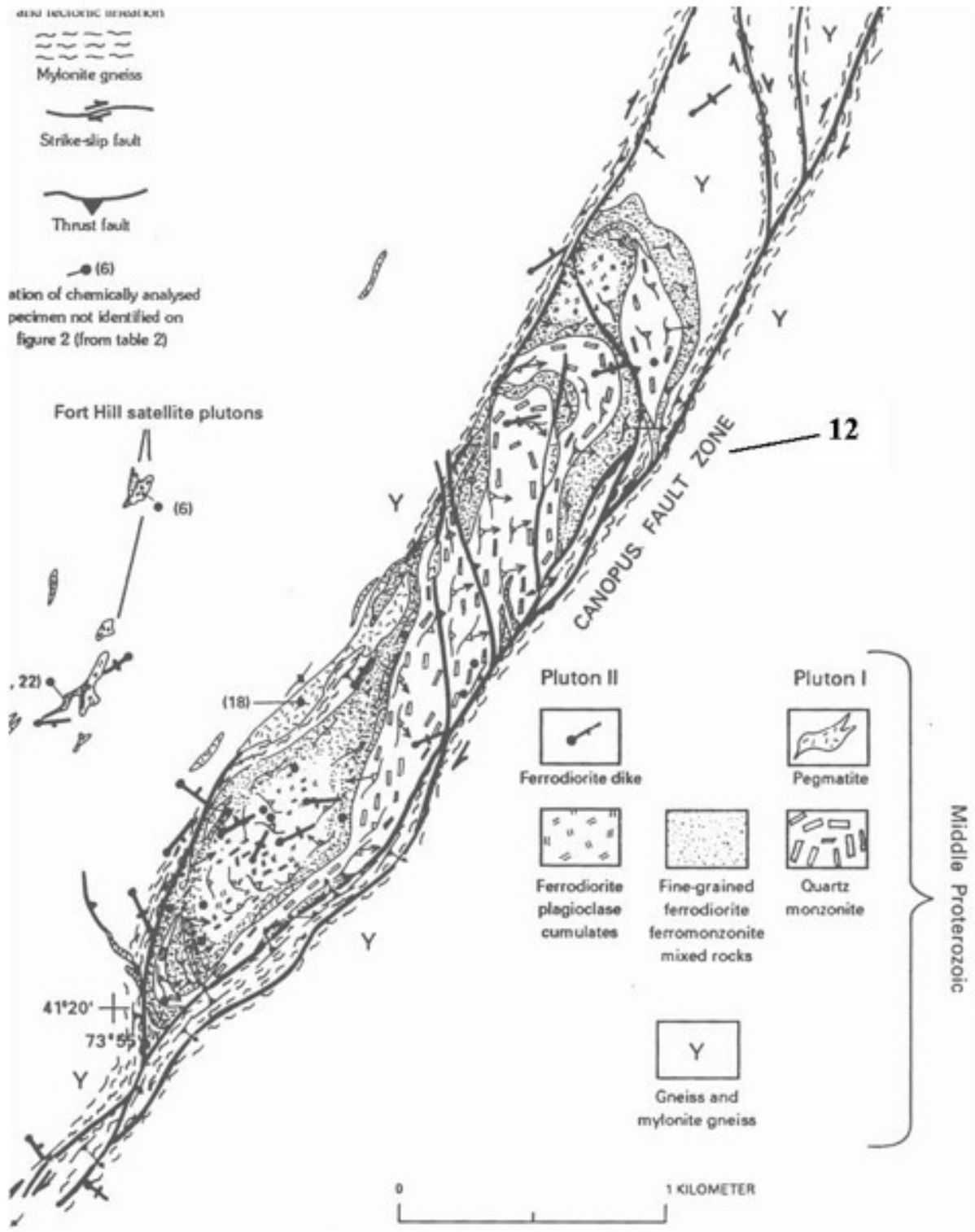


Figure 16. Map of the Canopus pluton after Ratcliffe and others (1972).

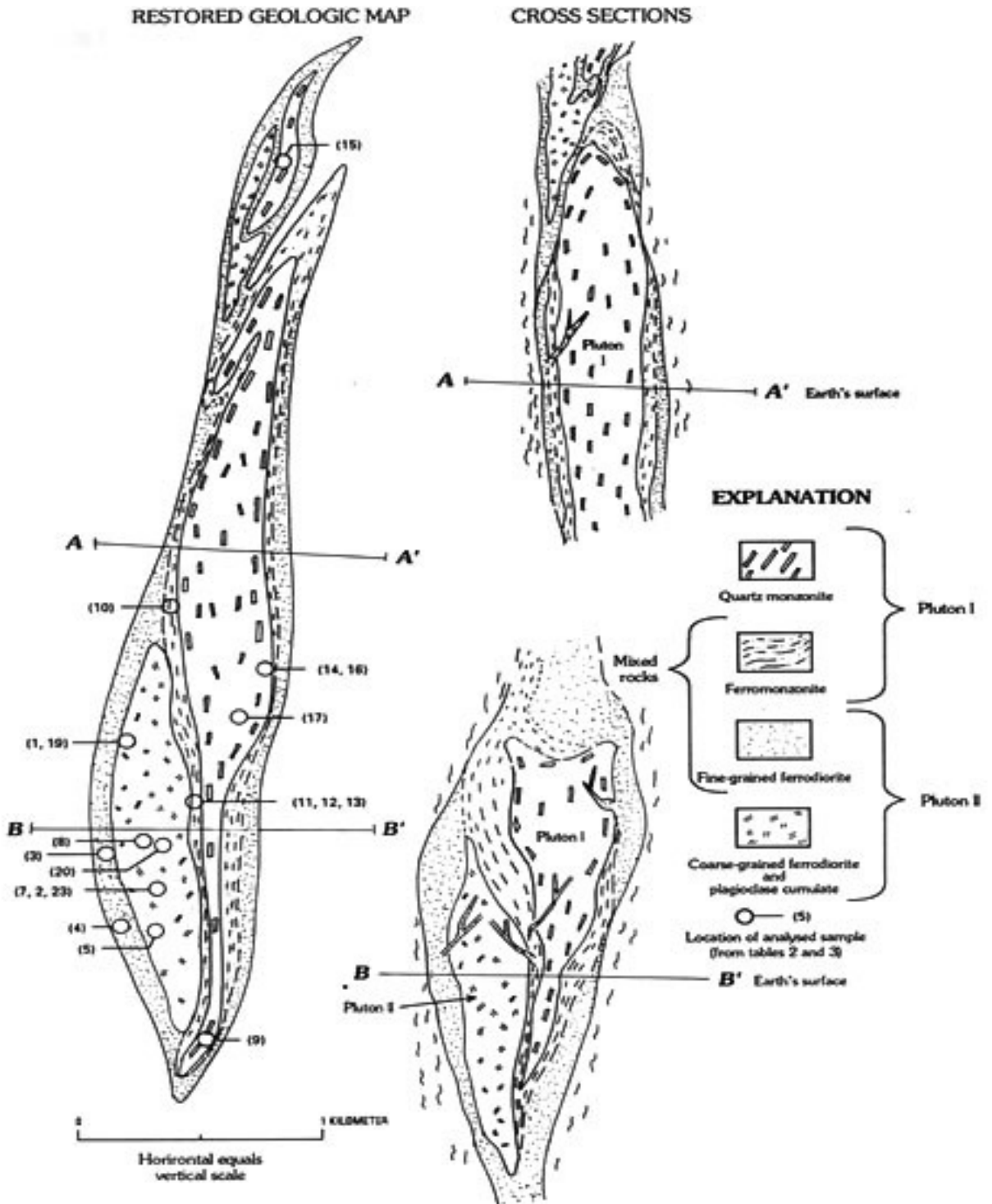


Figure 17. Restored undeformed map on left and sections across the Canopus composite dike showing the intrusive centers of pluton 1 and 2 numbers refer to location of samples for geochemistry dot presented here.

Back down and drive 0.2 mile south to new blasted driveway on right opposite mail box 462. Pull up drive far enough to allow all cars to get off of Canopus Hollow Rd.

Stop 6B. Canopus pluton 1143 ± 12 Ma, biotite monzonite and hypersthene biotite diorite composite intrusive

This spectacular and demonstrative new driveway cut exposes the transition from the Canopus shear zone into the Canopus pluton. The Canopus pluton is intruded between two Mesoproterozoic right lateral shear zones. The eastern one has been reactivated at greenschist grade and is exposed here. The western one is not reactivated and contains stringers of monzodiorite- monzonite and cross cutting composite diorite- monzonite dikes of the main pluton and abundant pegmatite, all of which crosscut the mylonitic host rocks, or as much as 2 km north of the pluton. Internal structure in the pluton consists of mutually crystallizing igneous phases and a well developed and folded igneous flow layering that defines right lateral deformation of the pluton during crystallization. At the walls igneous rocks of the pluton cross cut the bordering mylonite zones of the faults and fill internal shears within the pluton. Isoclinal folds in the mylonitic country rocks plunge nearly vertically down the mylonite structure, as do the axes of the internal and folded igneous flow structure in the pluton. This has been interpreted to mean intrusion of the igneous rocks into a subvertical active shear zone in the Mesoproterozoic. Early attempts to date the igneous rocks using Rb/Sr whole rock techniques by Dick Armstrong (Ratcliffe and others, 1972) and using multigrain zircon techniques by Aleinikoff gave ages of about 1.0 Ga and were erroneous. Subsequent SHRIMP zircon dating by Aleinikoff gives a more precise concordant age of 1143± 12 Ma with overgrowth ages of about 1 Ga. That is, during the end stages of the Shawinigan pulse of the Grenville orogeny (Rivers, 1997) as well as Ottawa overgrowths.

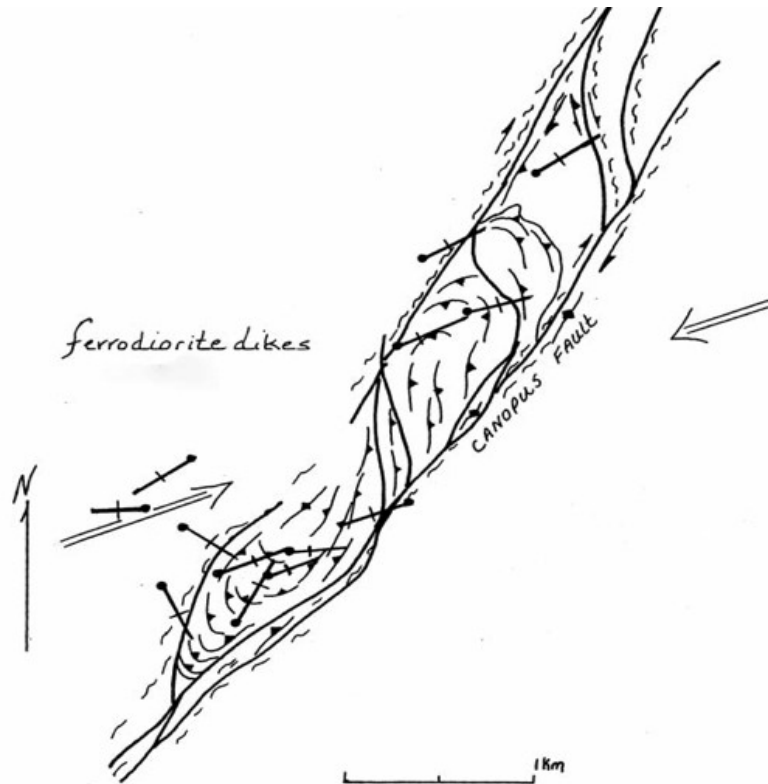


Figure 18. Schematic map showing the orientation of mafic dikes of the Canopus pluton that cross cut the sigmoidal igneous flow structure in the pluton, and the inferred stress field of the 1143 Ma strike slip event.

Starting from the road level some of the significant features to be seen are:

1. Mylonite, consisting of lit par lit veins of aplite, isoclinal steeply plunging folds in the mylonite.
2. Zones of pegmatite saturating and postdating the mylonite structure
3. The wall of the monzonite with fine grained monzonite in the mylonite
4. Tabular K-spar defining an igneous flow structure cut by aplite dikes and by intermediate diorite dikes
5. A pillowed zone consisting of monzodiorite ellipsoidal segregations in the monzonite and turbulent flow structure in the monzonite around the "pillows"
6. Ductile shear zones in the monzonite which deform the flow layering into lineated tectonites and undeformed granitic segregations which saturate the lineated zones

Some features described above are shown in Figure 19.

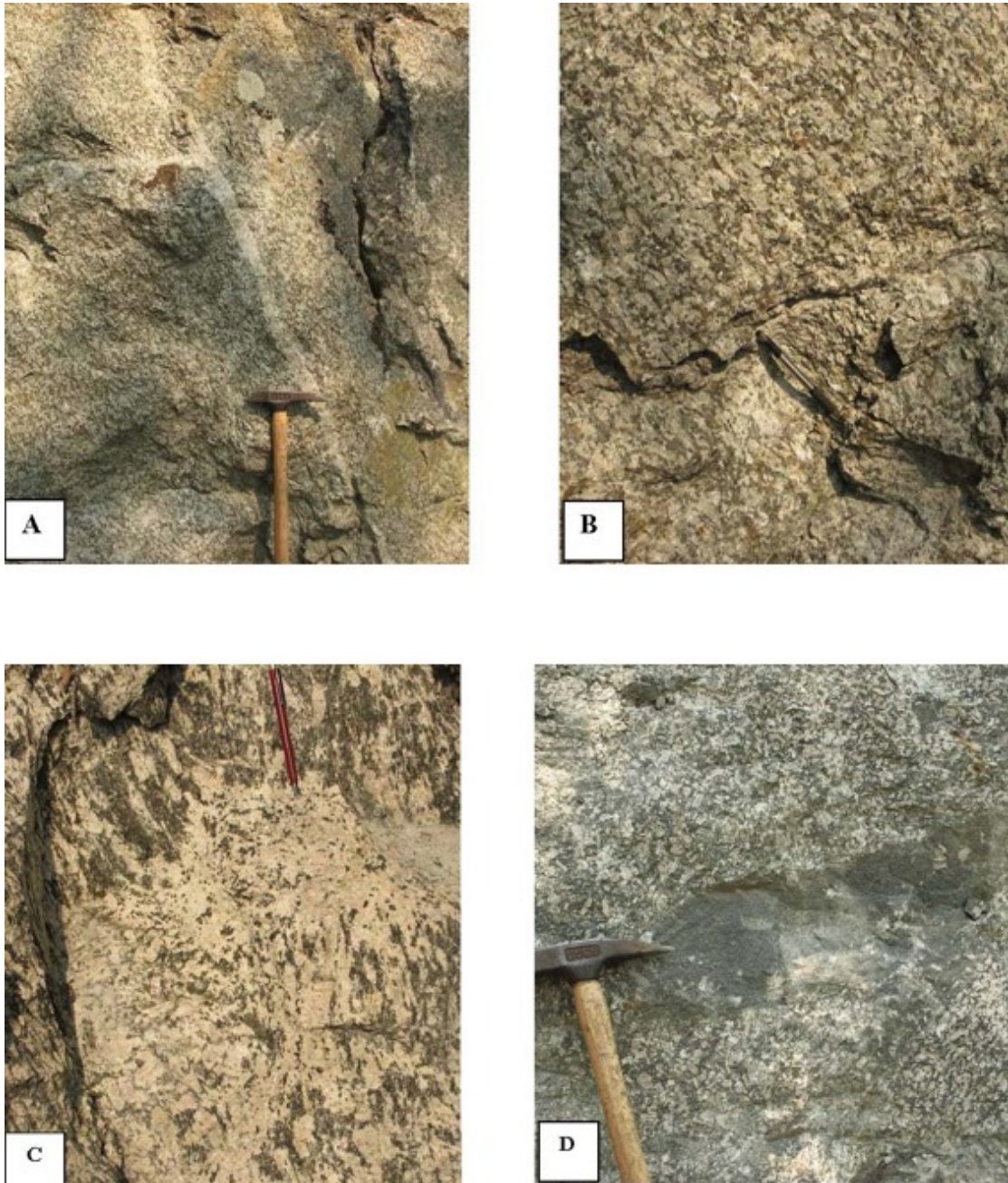


Figure 19. Igneous rocks of the Canopus pluton at Stop 6 b. A-Quartz monzonite flow structure cut by unfoliated granitic dikes. B-Close up view of tabular K-feldspar that defines igneous flow structure in the felsic parts of the intrusion. This is like the flow structure that has been mapped in the mafic and felsic parts of the pluton to define the internal geometry of the intrusive components as shown by symbols in figure 16 above. C-Deformed flow structure defined by oriented hornblende (parallel to the pencil) crosscut by pockets of late quartz monzonite showing that tectonic deformation of the pluton accompanied crystallization. D-Ferridiorite pocket in quartz monzonite in “pillowed zone”. Igneous and deformational features like these are developed throughout the pluton and show that the pluton was intruded during the right lateral deformational event during which it was intruded at about 1143 Ma.

These features and other similar observations made during mapping of the shear zones and the igneous rock (Ratcliffe, 1971; Ratcliffe and others, 1972) support the concept of syntectonic intrusion and document right-slip faulting at about 1143 Ma. To the east the right sigmoidal deformation in the area of Wiccopee diorite pluton indicates this type of tectonism in a zone as much as 10 km wide (Ratcliffe, 1992). Mafic dioritic to hornblendic dikes are common within this sheared zone cutting the fault fabrics and folds associated with the strike-slip events. This is very similar to the dikes and igneous rocks associated with stops 2 and 3 today in the Brewster area where a similar age for intrusion of 1134 ± 8 Ma has been established for comagmatic granitic and mafic rocks.

Mafic dikes commonly less than several meters thick, crosscut high grade shear zones on both sides of the Canopus shear zone and have compositions similar to the dioritic and most mafic part of the pluton. Locally ultramafic hornblendite plugs and radiating dikes are found near Fort Hill. Importantly, dioritic dikes and segregations within the pluton crosscut the igneous flow structure (Figure 6), and all rocks of the pluton lack a strong Mesoproterozoic deformational structure that is present in the country rocks. This is also true of the granites and dioritic rocks of the same age at Stops 3 and 4 and of the mafic dikes and gabbro-diorite plutons in the Brewster area. We may speculate that the deformed dikes seen in the tonalite-trondhjemite and at Stop 1 today might also be dikes of the same age.

Back down and continue south on Sprout Brook Rd.

0.5 mile entrance to Continental Village the dated monzonite came from just north of this intersection on the west side of the road.

0.6 mile stop sign at Sprout Brook and Gallows Hill Rd.

Proceed up to the left on Gallows Hill Rd.

0.2 mile turn left on Aquaduct Rd 0.4 mile small pond on left and drive pull in and park by first crops by two new houses at north end of pond.

Stop 7. Cat Hill Gneiss 1333 ± 7 Ma trondhjemite gneiss and Mesozoic cataclastic structure on the north limb of the Annsville syncline

On the south side of the road are limestone crops of the Cambrian and Ordovician Wappinger Group here barely at biotite grade. North of the road before the Cat Hill Gneiss are crops of the Lower Cambrian Poughquag Quartzite then cataclastic gneiss of the Cat Hill Gneiss. In the 1970's, I mapped an unconformity between Paleozoic rocks of the Annsville syncline and the basement as Walter Bucher had. New cuts and reexamination of the cover rocks along the Annsville syncline to the east now support high angle normal faulting and brecciation of the kind associated with Mesozoic tectonics and the Newark basin.

The Cat Hill Gneiss is less cataclastic in the outcrops behind the house but even these are not typical because of the extensive fracturing. A few hand samples from the dated locality less than 0.5 miles to the northeast demonstrate the difference. To get to the sample locality follow the directions to Stop 11 in Ratcliffe and Aleinikoff (2008)

The Cat Hill Gneiss is a white-weathering biotite-oligoclase-quartz gneiss of trondhjemitic and tonalitic composition. It is typical of a series of similar massive biotite poor garnet bearing gneisses that occur in the Reservoir Gneiss of Berkey and Rice (1923) and which closely resemble parts of the Losee gneiss of the New Jersey Highlands. It also is very similar to tonalitic gneisses of the Green Mountains of Vermont, which formed as calc-alkaline intrusives between about 1390 Ma to about 1320 Ma (Ratcliffe and others, 1991). The rock is very massive, contains almost no inclusions. It does pass into a more garnet-rich phase and also hosts abundant white aplitic granites at its contacts with amphibolites and has brown weathering spots suggestive of hypersthene but none has been found here. These bordering aplites produce a spectacular intrusive breccia with inclusions of hornblende gneisses, amphibolites. Because the Cat Hill Gneiss contains migmatitic zones of similar aplite and granite it may be that the breccia is a product of mobilization and intrusion along the walls resulting from secondary mobilization in an event younger than 1337 Ma. The purple brittle fracture zones contain minute riebeckite and are very similar to fracture zones along the Ramapo and other faults along the Newark basin in New Jersey

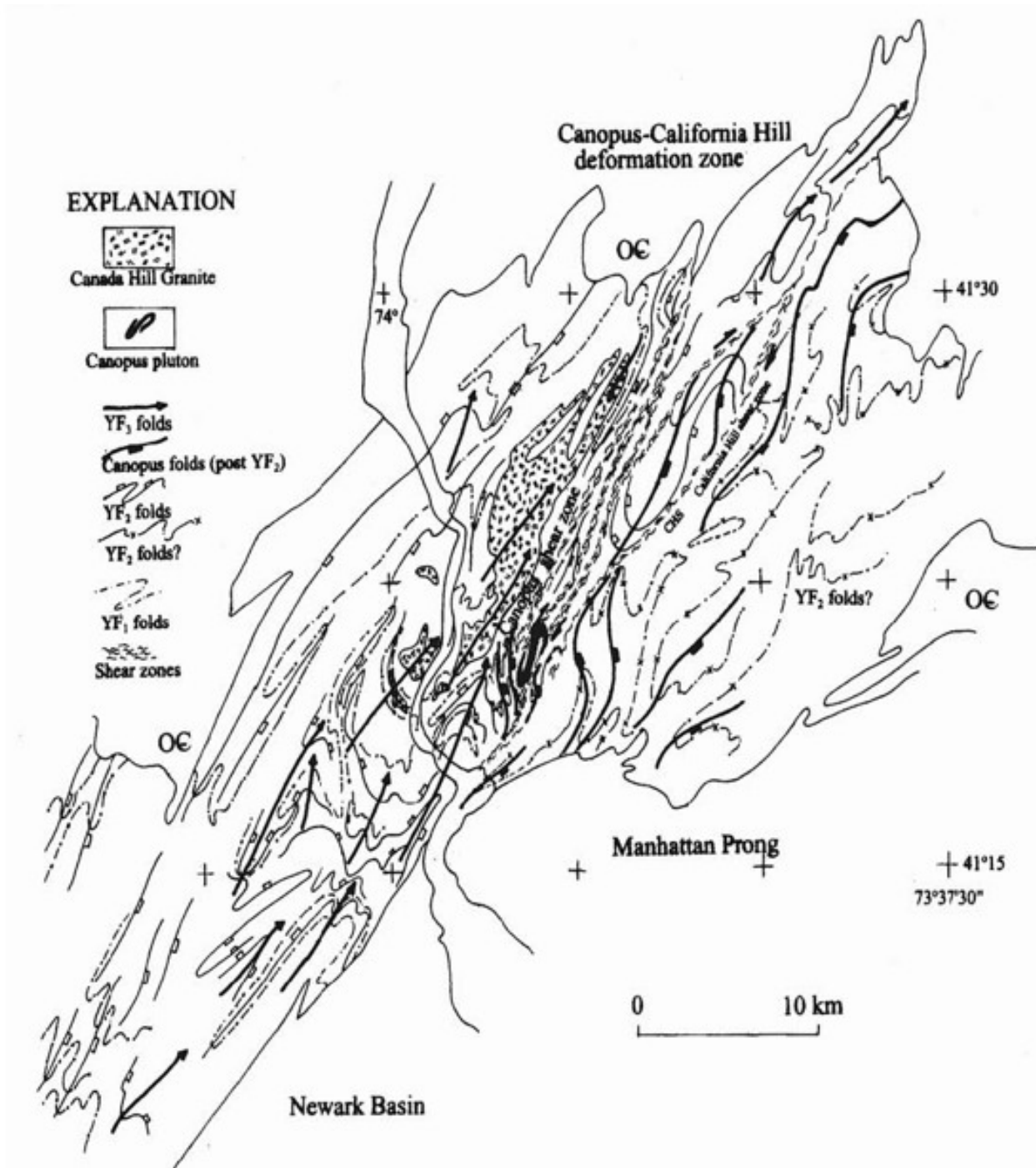


Figure 20. Regional tectonic map showing the distribution of Mesoproterozoic folds of the YF1, YF2 and YF3. Fold events YF1 and YF2 are judged to be Shawinigan and older. The YF2 folds east of the Canopus-California Hill shear zones may be Ottawa.

Figure 20 summarizes the axial traces of Mesoproterozoic folds in the Hudson Highlands of New York. Folds related to the Canopus-California hill deformation zone document a wide area of right sigmoidal folding here interpreted to be of Pre-Ottawan at approximately 1142 Ma and contemporaneous with intrusion of the Canopus pluton and its ferrodiorite dikes. Folds west of the shear zone, identified as Yf3 folds are Ottawan and folds identified as Yf2? to the east may be also be. In the Brewster quadrangle (Stop 2) and on Joes Hill similar dioritic dikes cross cut the granitic gneisses and intrude along steeply dipping high-grade shear zones. I tentatively correlate the igneous rocks of the Canopus area with the dikes in the Brewster area and with the dated granite-and diorite at stops 3 and 4. If this is correct, there may be evidence for a broad zone of right lateral shearing and igneous activity in the northern Hudson Highlands near the end of the Shawinigan phase of the Grenville orogeny. This raises the possibility that some of the amphibolite inclusions in these older rocks (Stops 1, 8 and 10) could be dikes, masking as xenoliths in the older granitic to tonalitic rocks! The more I learn about these rocks the more that interpretation seems likely to me.

Return to Aquaduct Rd turn left and left again onto Gallows Hill Rd. at the stop sign.
0.7 mi. turn left (S) onto Sprout Brook Rd.

Sprout Brook valley is a continuation of the Canopus Valley that extend south to the Hudson River and which is floored by highly mylonite silimanite containing garnet gneisses, as well as and calc-silicate rocks. Biotite grade phyllites of Ordovician age correlative with the Wallomsac Formation occur along the east side of the valley where they rest on Mesoproterozoic rocks along the east dipping Annsville fault which I originally interpreted was an Ordovician thrust fault but which I now regard as a Mesozoic normal fault traceable to branches of the main Ramapo fault to the south.

2.1 miles south to Roa Hook Rd, turn right and immediately bear right on to approach road to Rt. 9 D turn left on Rt. 9.

Exposures of Wallomsac Formation in contact with Mesoproterozoic gneiss along the extension of the Annsville fault. At these outcrops Walter Bucher documented a conglomeratic fine grained dolomite with clasts of gneiss which he interpreted as a basal units of the Wappinger Group resting unconformably on the basement. However, the large pebble- to cobble- size inclusions within the dolomite are all fragments of retrograded calc-silicate rock in an ultrafine grained mylonite or foliated dolomitic cataclasite that derived from Mesoproterozoic rock.

Follow Rt. 9 0.5 mi. into traffic circle and take Rt. 6 and 202 south towards Peekskill, cross bridge and at light turn right following Rt. 9.

Crops facing consist of Cambrian Poughquag Quartzite to the left are unconformable on Reservoir Gneiss. Large continuous crops of the Reservoir Gneiss continue for a mile to the south in high road cuts.

1 mile exit Rt. 9 to right and await instructions.

ROAD CONSTRUCTION MAY MAKE IT IMPOSSIBLE TO STOP HERE BUT ACCESS MAY BE POSSIBLE FROM THE PARKING LOT OF THE PEEKSKILL INN. An alternate site is at the Hudson River below the large outcrops.

Stop 8. Reservoir granite of Berkey and Rice (1921) at Peekskill

The exposures sampled for dating of zircon are at the east end of the large road cut on Rt. 9 and 202 below the Peekskill Inn. The exposures extend about 1000 feet to the north in high road cuts. Numerous thin to thick amphibolite gneisses and more biotitic gneisses form septae in the granodioritic gneiss. The rock is remarkably homogeneous and is interpreted as an intrusive rock. At one locality near the middle of the cut a pod of amphibolite is included in the granodiorite. A concordant SHRIMP zircon age of 1338 ± 6 Ma has been determined for the granodiorite. Zircon overgrowth ages are about 1.26 and 0.98 Ga.

Return to Rt. 9 heading south.

0.5 mi south on Rt.9 large new fresh outcrops of south dipping norite of the central basin of the Cortlandt complex here intruding the Wallomsac Formation.

3.5 mi exit Rt. 9 to Rt. 9A turn left at light.

0.5 mi. turn left on Furnace Dock Road.

0.4 mi. turn right into Furnace dock Condominiums and proceed 0.2 mi to the last building on the left pull in behind building.

Stop 9. Cortlandt complex monzonite and contact with the Walloomsac Formation

We are at the southeastern margin of the Cortlandt complex that consists of igneous norite, diorite, pyroxenite that cross cuts Taconic deformed metasediments of the Manhattan prong as well as the Laurentian basement. It was intruded following Taconic deformation and metamorphism and is one of the key observations in determining the timing of the Taconian orogeny. The igneous rocks are folded and intruded by the Devonian Peekskill Granite. A discussion of the contact relations, chemistry, and petrogenesis of the complex is beyond the scope of the present trip. Several papers and guidebooks provide that information (Bender, 1982; Ratcliffe and others, 1982, 1983) .

The central basin of the complex consists of inward dipping sheets of biotite norite, hornblende and biotite poikilitic norite and locally at the very border a monzodiorite that contains zircon. Pockets of interstitial k-feldspar- quartz and radiating biotite define slightly pinkish areas in the norite adjacent to the contact with the Walloomsac Formation. Zircon from this rock yielded a concordant SHRIMP age of 444 ± 3 Ma almost identical to the concordant SHRIMP zircon age from monzodiorite of the Peach Lake pluton (447 ± 5 Ma).

Near the contact igneous rocks and mobilized schist form an intrusive breccia In the breccia are rotated fragments of foliated and folded calc- silicate rock, and calc- schist characteristic of the basal part of the Walloomsac Formation (formerly termed Manhattan A of Hall, 1968). A breccia zone like this was previously recognized along this contact but none of the exposures are as spectacular as this one. Just west of this location and in the cuts for the exit ramp from Rt. 9W dikelets of mobilized schist containing garnet sillimanite-spinel- and corundum back intrude the chill norite and cut across the intrusive flow layering in the igneous rocks (see Stop 6 of Ratcliffe and others, 1983). Evidently the schists of the contact aureole were mobilized not as melts but as volatile-rich pneumatic injections, driven by contact-induced decarbonization and dehydration reactions along the walls.

Thin sections of the contact schist here contain garnet- biotite- staurolite- plagioclase and quartz and lack either kyanite or sillimanite both of which are present in contact rocks not far from here. Schist outside the aureole to the south on Torment Hill also contains abundant staurolite and lacks either kyanite or fibrolite.

Leave the parking lot via Furnace Dock Rd. to Rt.9A and turn left immediately, in 0.1 mi.onto Furnance Dock Rd as it continues across Rt.9A. Follow Furnance Dock 1 mile to the Hudson River and park at the entrance to Oscawana Island Park.

Stop 10 Granitic Fordham Gneiss 1351 ± 6 Ma and amphibolitic gneisses

The first exposures by the RR tracks are dark -gray biotitic gneisses, amphibolite gneisses and some sulphidic schists of the Fordham gneiss, which are in contact with the granite gneiss from the dated locality westward to this point. The biotite- hornblende gneiss resembles the inclusions we will see in the granite gneiss on Oscawana Island. The amphibolitic gneiss is layered and



Figure 21. Amphibolite screens or dikes? In the Fordham Gneiss at Oscawana Island.

grades into schists near the south end of the RR exposure. This suggests that the Fordham Gneiss contains paragneisses that are older than 1351 Ma. Hornblende from these exposures yielded a $^{40}\text{Ar}/^{39}\text{Ar}$ cooling age of 450 Ma (Dietsch and others, 2004).

Walk northwest to the island. Take bridge over tracks and follow trail to end of island. Walking south then west along the shore we will see excellent exposures of the dated gneiss and inclusions in it. Some of the tabular to irregular-shaped inclusions are layered, but the gneissosity seems to be shared by the granite. Note this is unlike the relations of the 1134 Ma intrusive rocks at Stops 3 and 4. I have not seen evidence that the 1350-1330 Ma rocks cross cut an older gneissosity.

Here we might consider how the zircon age and intrusive relationships should be interpreted. Does the zircon age date the time of the intrusion of granite, or could the intrusive relations we deduce have been produced by migmatization of a pile of felsic and mafic volcanic rocks? This is a difficult question not easily answered, either from the zircon zoning patterns or the field observations. I think we can conclude safely however that the granitic gneiss and the inclusions were present at 1351 Ma, but in what exact form is left undetermined. We have chosen to interpret the zircon age as the emplacement and crystallization age. If this is correct then the biotite hornblende gneisses in the RR cut might be older than the granite and some of the Fordham Gneiss older than 1351 Ma.

Summary diagrams of dike chemistry

Figures 22 and 23 give a summary of some of the chemical characteristics of the mafic dikes and amphibolites we have seen on this trip. The data are consistent with the mysterious dikes of the Brewster area being correlative with 1143 Ma Canopus plutonism rather than Ordovician, as the field observations indicate. The data also allow for some of the amphibolite "inclusions" in the ~ 1340 Ma granitoids to be deformed dikes, as field observations do indicate for some, perhaps in the same 1143-1134 Ma Shawinigan event. Notable are the coarse amphibolites (Stop 5) of the Nimham belt which are distinct outliers of low TiO_2 mafic rocks. Ta/Yb of the Brewster dikes overlap those of the Canopus dikes although the centroid of the former suggests a more depleted mantle source rather than more MORB-like source for the latter. The quite alkaline nature of the Cortlandt-age dike swarm suggests an enriched mantle source at about 450 Ma., and the high- TiO_2 – low MgO metadiabase dikes of probable Neoproterozoic age (Ratcliffe, 1987) suggest low degrees of partial melting in a rift event at about 570 to 600 Ma, although the age of these dikes has not been determined.

CONCLUSIONS

The recognition of pre-Shawinigan basement as old as 1350 to 1330 Ma in the Hudson Highlands and in the Fordham Gneiss of the Manhattan prong is a new finding that should permit the deciphering of the pre-Ottawan history of the very old basement rocks. Rocks of this age are now known to be widespread in other basement terranes of the Appalachians from New York north to Vermont where detailed mapping has delineated their distribution. Younger intrusive rocks 1174 to 1134 Ma of mafic and felsic composition, cross cut the older rocks and help define a Shawinigan event as well. The possibility of a widespread dike event and strike slip deformation seems likely in the Hudson Highlands but this structural event has not yet been recognized in other massifs to the north and there is uncertainty in the correlation of this event with that proposed by Gates and others (2004) to be Ottawan. Although the Ottawan deformation is known from migmatites and from intrusive rocks as young as the 1045 Ma Danbury augen granite (Walsh and Aleinikoff, 2004) and from zircon overgrowth ages (Table 1) the actual structural expression of the Ottawan in the Hudson Highlands is not as yet sorted out. As more and more workers discover the pre-Ottawan events in the Grenville of the Appalachians, the structural complexities become increasingly complex and intriguing. We hope that this trip will provide some food for thought.

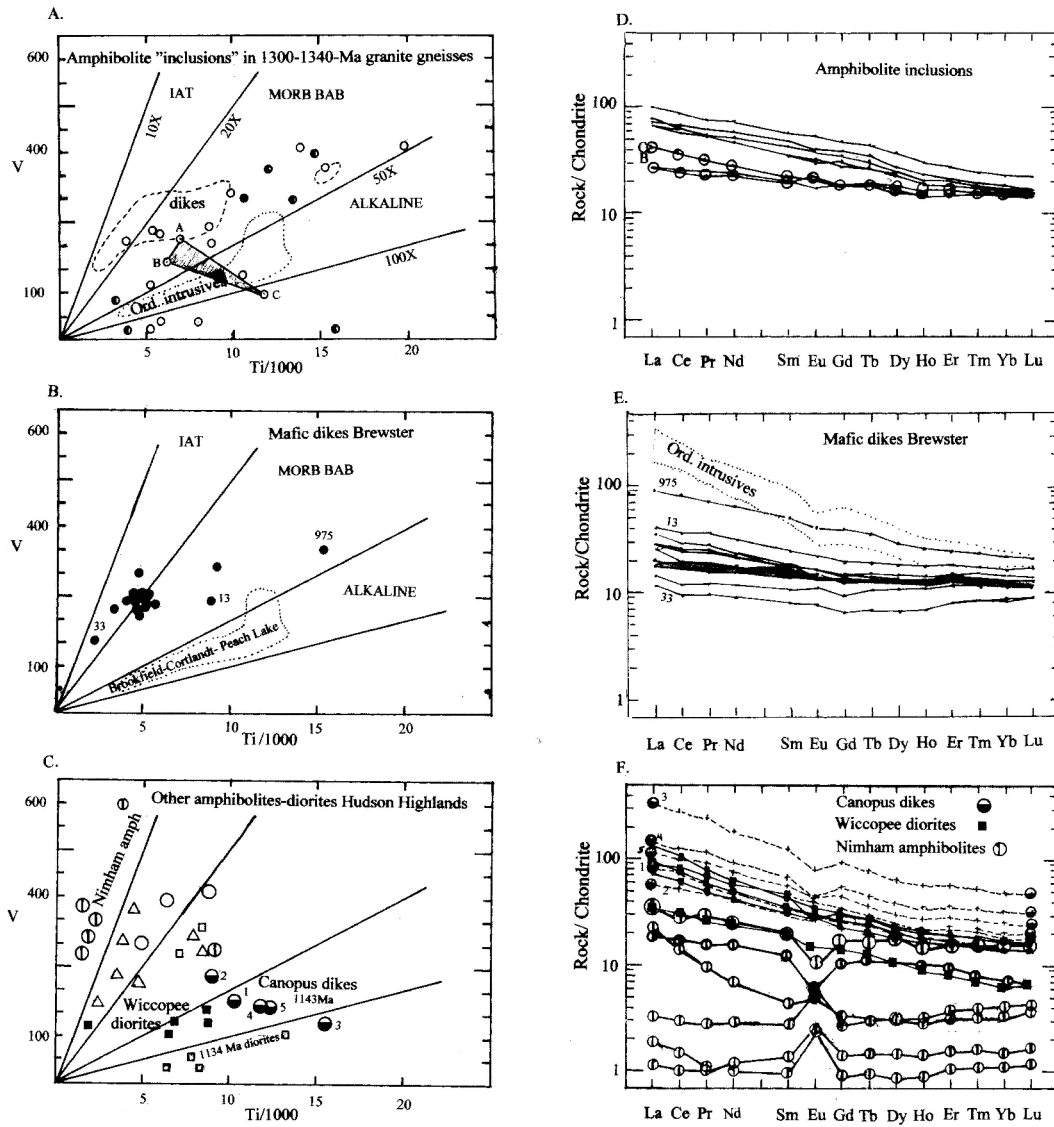


Figure 22. Chemical plots of amphibolites, diorites, and dikes of the Hudson Highlands, NY comparing “inclusions” in 1340-1330 Ma granitoids with 1143-1134 Ma diorites and dikes (Stops 3, 4 and 6) and with undated dikes (Stops 1, and 2). Ordovician dikes of the Cortlandt dike swarm and with probable Neoproterozoic metadiabase dikes (Ratcliffe, 1987). V vs Ti/1000 after Shervais (1982). REE normalization factors (Sun and McDonald, 1989) A. Amphibolite “inclusions” in 1340-1330 Ma granitoids (open circles) and Fordham Gneiss (filled circles). Dashed field shows mafic dikes in B.B. Mafic dikes that crosscut gneissosity in 1340-1330 Ma gneisses C. Other amphibolites of the Hudson Highlands. Circles with vertical line = Nimham belt (Stop5a). Triangles amphibolites Camp Smith interlayered with 1230 Ma felsic gneisses. Half-filled circles = Canopus dikes Stop 6. Open squares diorites of Wicoppee pluton. Filled squares = 1134 Ma diorites and granite Stops 3 and 4. Open circles 3 amphibolites in the Poughquag quadrangle east of Nimham belt.D.Chondrite normalized patterns of representative amphibolite inclusions. Shaded area three amphibolites (A and B) and pyroxenes- hornblende reaction zone (C) with Ottawa pegmatite Stop 1E. Chondrite normalized patterns for mafic dikes in relation to Cortlandt dike swarm F. Chondrite normalized patterns at top Canopus dikes-half filled circles (Stop 6) and dashed field). Middle fields Wicoppee diorites and at bottom amphibolites of Nimham belt (Stop 5).

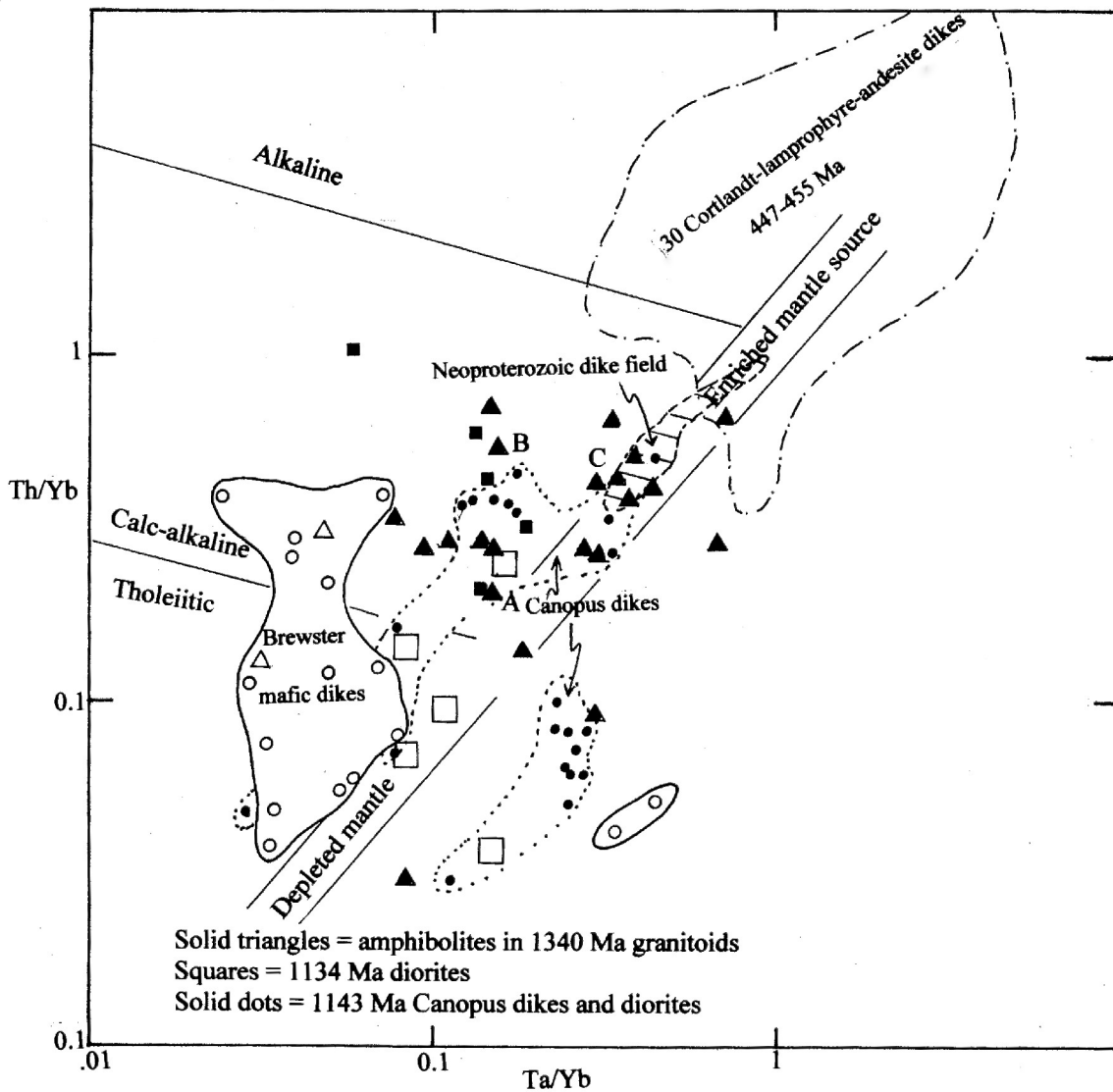


Figure 23. Th/Yb vs Ta/Yb plot (after Pierce, 1983) of mafic dikes in the Brewster area (open circles) compared to amphibolite (triangles) in 1330-1340 Ma granitoids of the Hudson Highlands and Manhattan prong, Samples A, B and C are amphibolites at Stop 1, C is pyroxene hornblende reaction zone with Ottawa pegmatite. Filled circles are mafic dikes and chill ferrodiorites of the 1143 Ma Canopus pluton (Stop 6), and squares are diorites of the Wiccopee pluton (filled squares), and 1134 Ma diorite and hornblende granite (open squares) at Stop 3 and 4. These are shown in comparison to the alkalic dikes of the Cortlandt- Rosetown and Popolopen Lake lamprophyre dike swarm of late Ordovician or early Silurian age and probable Neoproterozoic metadiabase rift basalts of the Hudson Highlands (Ratcliffe, 1987). The diagram suggests that the Brewster dikes are not Neoproterozoic or Ordovician but could be related to 1130-1140 Ma plutonism. It also suggests that many of the “inclusions” in the 1330-1340 Ma tonalitic- trondhjemite and granitic gneisses may be part of an extensive set of Shawinigan ferrogabbro-diorite- ferromonzonite intrusives that include the 1143 Ma Canopus pluton and dikes, the Wiccopee pluton and unnamed 1143 Ma diorite- hornblende granite of the Brewster area.

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**Refining the Timing of Faunal Turnover in the Middle Devonian Appalachian Basin:
Paleoecological Analysis of the Earliest Hamilton Fauna and a Revision of the
Base of the Givetian Stage in Eastern North America**

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INTRODUCTION

The Middle Devonian of the Appalachian Basin has long been a testing ground for various evolutionary and paleoecological hypotheses (Eldredge and Gould, 1972; Brett and Baird, 1995; etc.). Units such as the Onondaga Limestone and Hamilton Group contain an excellently preserved invertebrate fauna spanning a wide array of marine facies ranging from shallow, reefal environments with diverse faunas to deep-water, dysoxic settings typified by nearly monospecific assemblages. The well-documented nature of the stratigraphy of the Middle Devonian rocks of New York State in particular makes this interval the perfect area in which to investigate the spatial and temporal nature of paleoecology and faunal turnover during this critical interval in Earth history.

Workers such as Cleland (1903) recognized early on the presence of patterns of extended intervals of faunal stability within the Middle Devonian of New York State, patterns that lead Brett and Baird (1995) to the formation of the hypothesis of Coordinated Stasis. Their hypothesis states that there are extended intervals of compositional faunal stability, termed evolutionary-ecological subunits (EE subunits), within various environments throughout geologic time, that are bounded by relatively rapid pulses of faunal turnover with very low amounts of taxonomic hold-over between stable intervals.

More recent investigations on the topic of Coordinate Stasis have focused on gaining a better understanding of the type interval from which Brett and Baird formed their hypothesis within the Middle Devonian of the Appalachian Basin. Bonnelli et al. (2006) documented the nature of abundance patterns within the Hamilton EE subunit, noting that faunas from similar paleoecological settings had high levels of compositional stability but low levels of stability when taking abundance of the taxa into consideration; although the composition of the top ten taxa changed little from bed to bed, the abundance ordering between beds was quite different. A more recent study by Brett et al. (2007) focused on patterns of biofacies tracking within the Hamilton EE subunit, noting high levels of similarity of taxonomic relative abundance within biofacies through depositional cycles. While these studies have documented the nature of the intervals of faunal stability, the intervening periods of faunal turnover remain as yet poorly understood. The present study aims to shed light on the details of the faunal turnover between two of the EE subunits with the Middle Devonian of the Appalachian Basin, the Stony Hollow and Hamilton EE subunits, specifically examining how rapid was that turnover and how quickly were the previously documented suit of biofaces from the Hamilton interval established within the basin.

GEOLOGIC SETTING

Sediments of latest Eifelian-earliest Givetian age in the Hudson Valley were deposited in a retroarc foreland basin that was formed on the eastern margin of Laurentia, situated at about 20° south of the equator (Blakey, 2009, Figure 1), cratonward of the advancing Acadian Terrain. A large wedge of sediment, known as the Catskill Delta, was beginning to prograde out into the basin, affecting sedimentation and sea floor topography in the Hudson Valley at this time. The eastern margin of the current Middle Devonian Appalachian Basin sediments is undoubtedly artificial in nature, being the result of erosion during the last 300 mya, and the basin most likely extended far to the east during this time. However, as will be demonstrated below, the shoreline of the basin was most likely located near the present-day eastern boarder of New York State during the deposition of the interval under investigation.

Deposition within the Appalachian Foreland Basin during the Acadian Orogeny has been interpreted as being episodic in nature, with major pulses of clastic sedimentation, termed tectophases by Ettensohn (1985), punctuated by intervals of carbonate-dominated sedimentation during periods of tectonic quiescence. The strata of the Hamilton Group were deposited during Tectophase II of the Acadian Orogeny that began with a major deepening event in the late Eifelian. During this time there was a drowning of the shallow-water carbonate platform that had existed during the deposition of the Onondaga Formation associated with loading and downwarping of the eastern margin of the North American craton that lead to the deposition of deeper-water, dysoxic shales of the late Eifelian-early Givetian Marcellus sub-Group.

Most previous studies of the Hamilton Group strata have taken place in western and central New York State as these areas preserve relatively thin packages of fossil-rich sediment containing alternating packages of limestone and shale. Correlative sections in eastern New York State remain poorly understood due to their great thickness and the more homogeneous nature of their sediments. There is a drawback, however, to focusing only on sections in west-central New York in that these sections are much more condensed in relation to time, resulting in a loss of temporal resolution; thick sections in the east allow for a much greater understanding of the precise timing of change within the basin, however careful attention must be paid to stratigraphic detail. The stratigraphic interval containing the Stony Hollow Fauna, in west-central New York ranges from nearly nil at Lake Erie to about 10 m around Syracuse, while in eastern New York in the Hudson Valley this interval spans well over 300 m (Rickard, 1975).

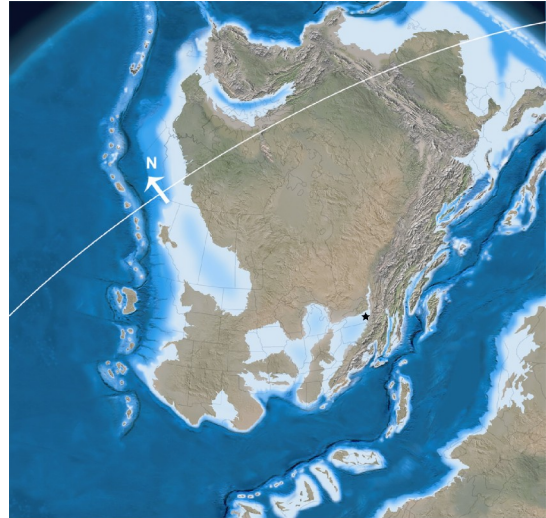


Figure 1. Paleogeographic reconstruction of the North American craton during the Middle Devonian. Star marks study area. Modified from Blakey (2009).

BACKGROUND

In their initial paper on Coordinate Stasis, Brett and Baird identified only three EE subunits within the Middle Devonian of the Appalachian Basin: the Onondaga, Hamilton, and Genesee subunits. Within the lower Hamilton EE subunit they recognized the presence of a short-lived, anomalous fauna within the basin, parsing this interval out as a sort of sub-subunit; another such interval was recognized within the upper Hamilton EE subunit as well with the incursion of the Lower Tully Fauna. Subsequent investigation of these intervals has led to the recognition of these two periods of anomalous faunas within the Hamilton EE subunit as their own unique EE subunits. The fauna from the interval between the Onondaga and Hamilton EE subunits, termed the Stony Hollow Fauna, has been well-documented as an incursion of warm-water, dysoxic-tolerant taxa from arctic Canada that invaded all across eastern North America during the latest Eifelian Stage (Koch and Day, 1996) associated with the initial deepening of tectophase II of the Acadian Orogeny (Ettensohn, 1985). Similarly, the interval of anomalous fauna within the upper portion of the Hamilton EE subunit, known as the Lower Tully EE subunit, has been identified as a dysoxic-tolerant fauna that replaces the Hamilton Fauna in many areas around the basin (Baird and Brett, 2003). Both of these EE subunits are recognized as being associated with bioevents on a global scale; the faunal turnover between the Stony Hollow and Hamilton EE subunits represents the local manifestations of the Kacak Bioevent while the turnover at the end of the Lower Tully EE subunit marks the onset of the Taganic Bioevent.

The incursion of the Stony Hollow Fauna (Figure 2) occurs during the initial deepening associated with Tectophase II of the Acadian Orogeny. The first appearance of the Stony Hollow Fauna occurs in the Bakoven Member of the Union Springs Formation, interpreted to have been deposited during the highstand of a third-order depositional sequence. The fauna recurs with increasing diversity through the Stony Hollow Member which represents falling stage portion of the sequence and continues into the transgressive phase of the next overlying third-order sequence in the Cherry Valley Member of the Mount Marion and Oatka Creek formations (Figure 3). The first appearance of the

Hamilton Fauna (Figure 4) within the Appalachian Basin was long recognized as occurring in the Halihan Hill Bed, or *Meristella*-coral bed as it was originally known (Goldring, 1935), of the Mount Marion and Oatka Creek formations (Figure 3). In west-central New York, the Halihan Hill Bed is the first abundantly fossiliferous horizon above the Cherry Valley Member, observed as lying approximately a meter above the Cherry Valley Member in outcrops just south of Syracuse.

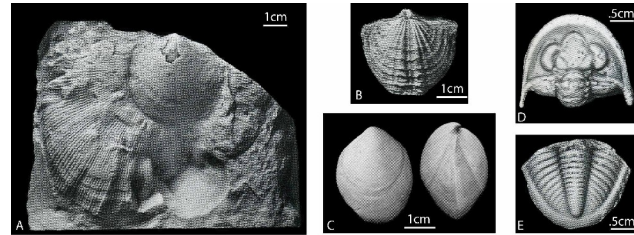


Figure 2. Stony Hollow Fauna: A) *Variatrypa*, B) *Carinatrypa*, C) *Subrennselandia*, D, E) *Dechenella* (modified from Ehlers and Kessling, 1970)

The interval between the Cherry Valley Member and the Halihan Hill Bed is known as the East Berne Member (Ver Straeten and Brett, 2006, Figure 3) and where present in west-central New York is most often represented as dark-gray shale with little to no fauna. The stratigraphy of the East Berne Member in the Hudson Valley has been well studied by Ver Straeten (1994) where it preserves facies ranging from dark-gray shales near its base, grading upwards into siltstone interbedded with thicker and thicker sandstones (>1 m) near the top. The lower portions of the East Berne Member are relatively barren of fauna with infaunal bivalves being rarely found. Approximately 15-20 m down from the top of the unit is a distinct fossiliferous horizon known as the Dave Elliot Bed (Ver Straeten, 1994). This bed is traceable throughout the Hudson Valley where it is underlain by a layer of distinctive vertically-oriented concretions and represents a small-scale transgressive interval within the East Berne Member. Together, the East Berne Member and Dave Elliot Bed sit within a critical interval between two EE subunits and the description of the fauna of this interval represents the bulk of the data examined in this paper.

METHODS

The outcrop belt of Middle Devonian sediments exposed along the present-day Hudson Valley provides a transect nearly parallel to the hypothesized shoreline that existed during deposition. Nine outcrops of the upper East Berne Member were examined in the between Kingston in the south to East Berne in the north, with a single outcrop from western New York examined at the Seneca Stone Quarry located between Seneca and Cayuga lakes (Figure 5). At each section, the location of the Dave Elliot Bed was determined and volumetrically standardized samples were taken from multiple locations along the outcrop if possible. Samples were taken back to the lab where they were processed and all possible specimens identified to the species level and counted. Raw counts were standardized to the level of relative abundance using the categories: rare (0-4 specimens), moderately common (4-10 specimens), common (10-19 specimens), and abundant (>20 specimens). This method allows for the standardization of samples across various environments, downplaying the sometimes large numbers of certain taxa (i.e. the infaunal bivalve *Nuculoidea*), which in some samples are found almost to the exclusion of all other taxa. Additional presence-absence data from the northernmost locality for this study (location 9, Figure 5) was taken from Goldring (1935) as the Dave Elliot Bed is now covered at this section. Faunal and sedimento-

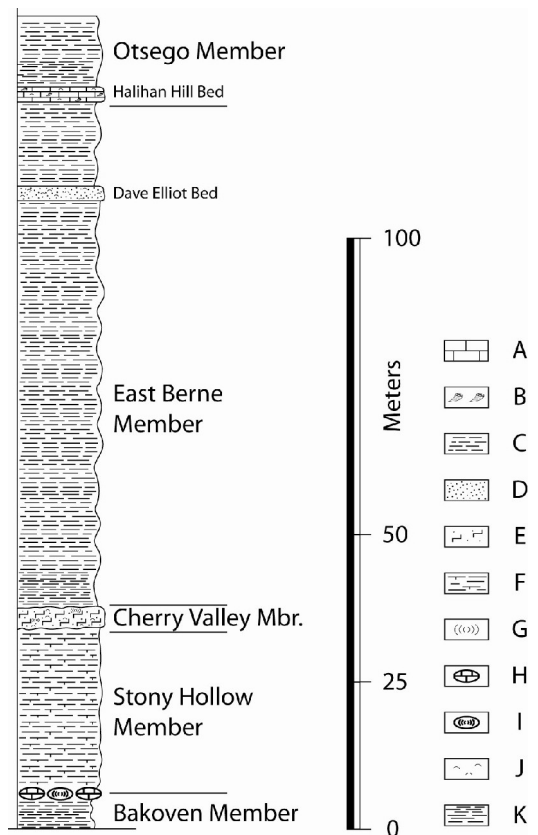


Figure 3. Simplified stratigraphic column for the study area. A) limestone, B) coral bed, C) gray shale, D) sandstone, E) calcareous sandstone, F) siltstone, G) goniatites, H) calcareous concretion, I) goniatite bearing concretion, J) shell bed, K) dark gray shale.

logic data were used to prepare paleoecological reconstructions of each of the 10 localities (see below).

Various statistical methods were used to examine the data from the Dave Elliot Bed. Comparison of faunal composition of individual samples from the Dave Elliot Bed was made using similarity coefficients and Detrended Correspondence Analysis using both relative abundance data and presence absence data as locality permitted. Similar methods were also used to compare samples from the Dave Elliot Bed to samples from the Centerfield Member cycle described in detail by Brett et al. (2007) to determine the similarity of biofacies through the Hamilton Group.

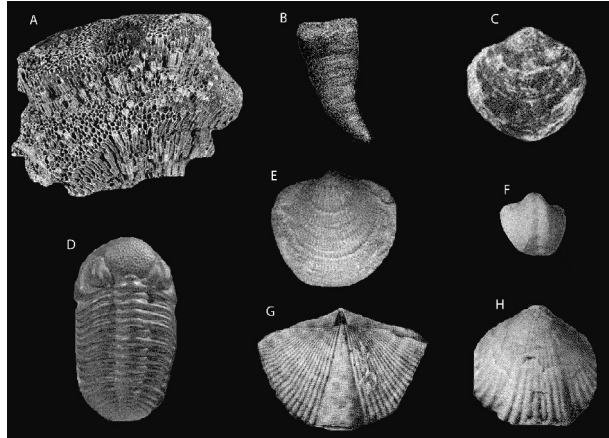


Figure 4. Hamilton Fauna: A) *Favosites*, B) *Stereolasma*, C) *Athyris*, D) *Eldredgeops*, E) *Pseudoatrypa*, F) *Ambocoelia*, G) *Mediospirifer*, H) *Eumetabolotoechia*.

DATA

Descriptions of the localities examined in this study will be discussed from south to north, beginning with

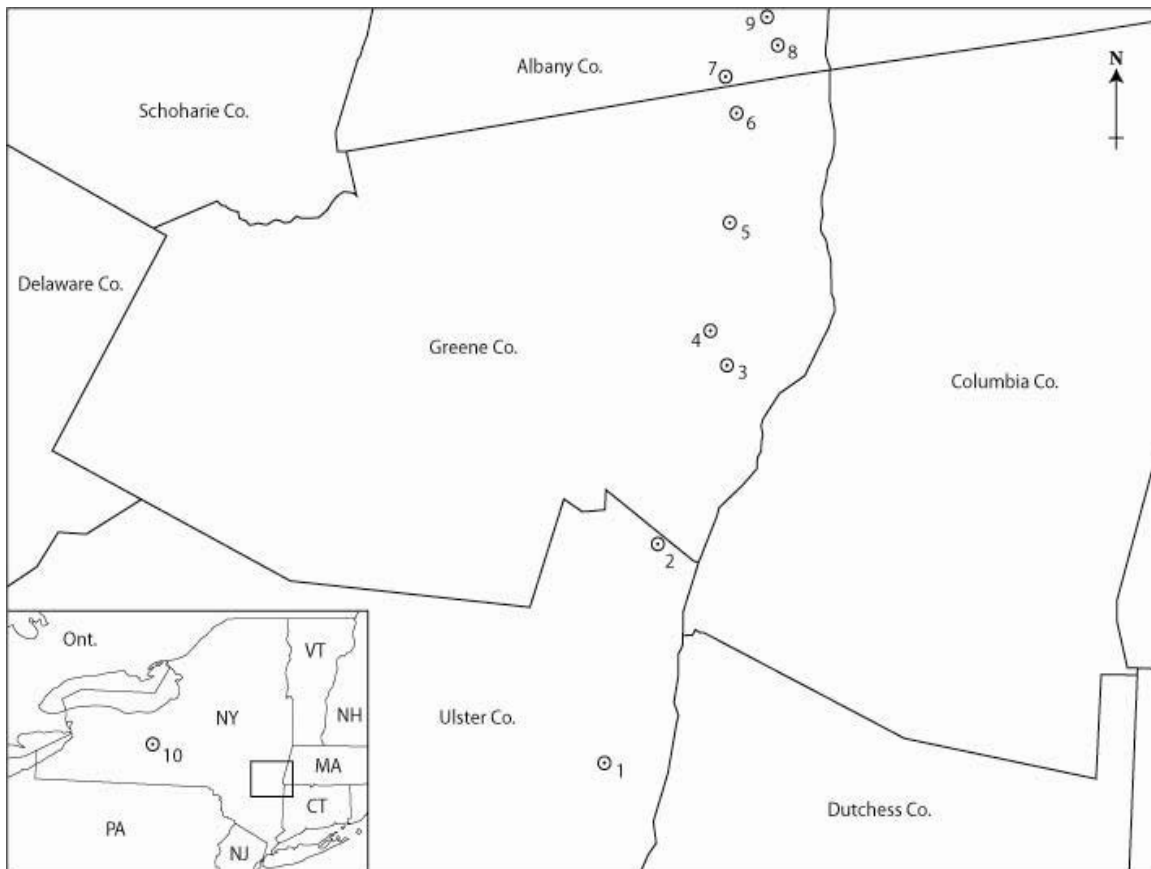


Figure 5. Map of sampling localities for the Dave Elliot Bed. Inset shows larger regional view, internal box represents enlarged area of mid-Hudson Valley. 1) Rte. 209; 2) Dave Elliot Rd.; 3) Sandy Plains Rd.; 4) Buttermilk Falls; 5) Earlton Quarry, Earlton; 6) Stanton Hill Rd.; 7) Gedney Hill Rd.; 8) Hannacroix Creek; 9) Cole Hill Rd., East Berne; 10) Seneca Stone Quarry.

Taxa \ Locality	Kingston 209	Dave Elliot Rd.	Sandy Plains Rd.	Buttermilk Falls	Earlton Quarry	Stanton Hill Rd.	Gedney Hill Rd.	Hammacraix Creek	Seneca Stone
Ambocoelia			*						*
Arcuaminites			*						
Athyris	*	*	*	*	*	*	*	*	*
Cyrtina	*	*	*	*	*	*	*	*	*
Cupulorostrum			*	*	*	*	*	*	*
Emmanuella	*								
Eoschuchertella	*			*	*	*	*	*	*
Eumetabolotoechia	*						*	*	*
Hallinetes	*	*	*	*	*	*	*	*	*
Lingula							*	*	*
Longispina		*	*	*	*	*	*	*	*
Mediospirifer		*	*	*	*	*	*	*	*
Mucrospirifer	*	*	*	*	*	*	*	*	*
Nucleospira		*	*	*	*	*	*	*	*
Orbiculoidea				*	*	*	*	*	*
Protoleptostrophia				*	*	*	*	*	*
Psuedoatrypa				*	*	*	*	*	*
Sinochonetes	*	*	*	*	*	*	*	*	*
Tropidoleptus		*	*	*	*	*	*	*	*
Trunculosia				*	*	*	*	*	*
Actinopteria				*	*	*	*	*	*
Goniophora				*	*	*	*	*	*
Grammysia	*	*	*	*	*	*	*	*	*
Leiopteria				*	*	*	*	*	*
Modiomorpha	*	*	*	*	*	*	*	*	*
Nuculites		*	*	*	*	*	*	*	*
Nuculoidea	*	*	*	*	*	*	*	*	*
Paleoneilo		*	*	*	*	*	*	*	*
Paracyclas				*	*	*	*	*	*
Pseudaviculopectin	*	*	*	*	*	*	*	*	*
Bellerophon				*	*	*	*	*	*
Bembexia		*	*	*	*	*	*	*	*
Platyceras	*	*	*	*	*	*	*	*	*
Tornoceras	*	*	*	*	*	*	*	*	*
Straight Cephalopod	*	*	*	*	*	*	*	*	*
Hyaliths	*	*	*	*	*	*	*	*	*
Tentaculitids				*	*	*	*	*	*
Aulocystis				*	*	*	*	*	*
Favosites				*	*	*	*	*	*
Heliophyllum				*	*	*	*	*	*
Heterophrentis		*	*	*	*	*	*	*	*
Stereolasma				*	*	*	*	*	*
Conularia				*	*	*	*	*	*
Bryozoan		*	*	*	*	*	*	*	*
Crinoid ossicles		*	*	*	*	*	*	*	*

● Abundant (>21)
● Common (11-20)
● Moderately Common (5-10)
● Rare (<4)

Table 1. Faunal list for all localities.

outcrops along the western side of Rte. 209 just west of Kingston where nearly the entire East Berne Member (all but the lowest ~10 m) is exposed in a large cliff (Figure 5). At this locality the East Berne Member consists of dark-gray shale grading upward into fine-grained silts with interbedded thin- to medium-bedded fine-grained sandstones. The bulk of the East Berne Member is relatively barren here with very rare, scattered infaunal bivalves and pyritic thread-burrow. There are two ~5 cm thick, compact, pyritic fossiliferous beds exposed near the middle-upper portion of the member separated by ~5 m of barren silty-shale. The upper fossiliferous bed has been identified as the Dave Elliot Bed, being underlain by the typical vertically-elongated concretions. Both beds are faunally similar, being dominated by large numbers of nuculid bivalves with rare leiorhynchid brachiopods, gastropods, and cephalopods (Table 1, Figure 6).

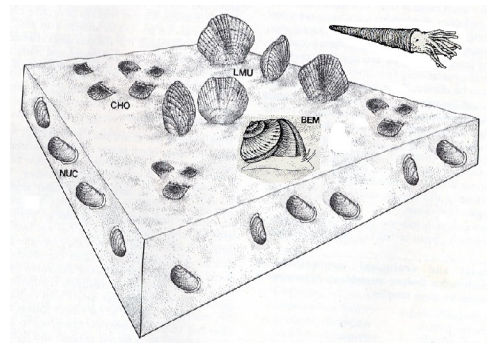


Figure 6. Paleoecological reconstruction for Rt. 209 locality. LMU: *Eumetabolotoechia*, NUC: *Nuculoidea*, BEM: *Bembexia*, CHO: *chonetids*.

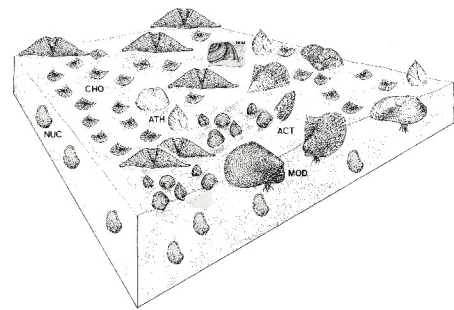


Figure 7. Paleoecological reconstruction for Dave Elliot Road. MOD: *Modiomorpha*, ATH: *Athyris*, ACT: *Actinopteria*

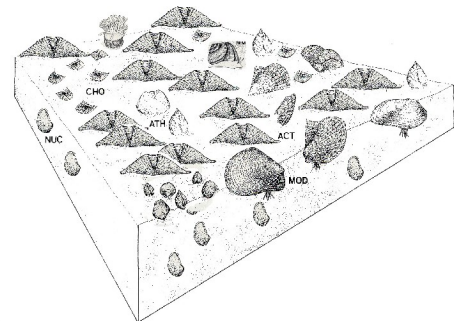


Figure 8. Paleoecological reconstruction for Sandy Plains Rd.

The next outcrop of the Dave Elliot Bed to the north is at the type locality at Dave Elliot Road west of Saugerties, a distance of 20 km north of Rte. 209 in Kingston (Figure 5). At this locality there is only one shell bed exposed within this portion of the East Berne. The Dave Elliot Bed here is a ~15 cm thick siltstone with fossil scattered throughout. The fauna of the Dave Elliot Bed is moderately diverse being dominated by small chonetid brachiopods along with abundant *Mucrospirifer*, *Athyris*, and *Emmanuella*. (Table 1, Figure 7).

~15 km. north of Dave Elliot Road, the Dave Elliot Bed examined is well-exposed along Sandy Plains Road just northwest of the village of Leeds (Figure 5) At this locality the Dave Elliot bed is preserved as a ~30 cm thick shelly sandstone at the top of an approximately 2 m coarsening upward succession. The fauna of the Dave Elliot Bed is dominated here by abundant spiriferid brachiopods along with abundant scattered crinoid columnals and moderately common bryozoans. This locality also contains the southernmost example of the typical Hamilton brachiopod *Tropidoleptus* (Table 1, Figure 8).

Just west and north (~2 km) of the exposure along Sandy Plains Road, the Dave Elliot Bed is exposed in a waterfall on Buttermilk Creek (Figure 5) The East Berne Member is well exposed in an abandoned quarry and in the creek here consisting of a coarsening upward succession of silty shale grading upward into thin- to medium bedded sandstones. Lithologically the Dave Elliot Bed closely resembles the exposure along Sandy Plains Road, however there are some distinct faunal differences between the two outcrops. The Dave Elliot Bed here contains abundant *Tropidoleptus* along with a greater abundance of chonetids and spiriferids (Table 1, Figure 9).

The next outcrop of the Dave Elliot Bed to the north is an exposure in an abandoned quarry at Earlton, ~10 km north of Buttermilk Creek (Figure 5). Most of the East Berne Member is covered at this outcrop, however it is possible to determine the stratigraphic location of the Dave Elliot Bed as an active quarry containing the Halian Hill Bed is located just uphill from exposures of the Dave Elliot Bed. The Dave Elliot Bed at this outcrop consists of scattered fossils within the upper ~10 cm of a ~1.5 m thick very hard sandstone. This outcrop preserves the most diverse assemblage of taxa found to date within the Dave Elliot Bed including abundant *Tropidoleptus*, *Mediospirifer*, and *Mucrospirifer* brachiopods, moderately common to common rugosan corals including *Stereolasma* and *Heterophrentis*, and rare favositid corals (Table 1, Figure 10).

Continuing ~10 km north, the next outcrop of the Dave Elliot Bed examined is a creek exposure in Stanton Hill Ravine (Figure 5). The middle through upper East Berne is well-exposed in this ravine and consists of medium- to thick-bedded sandstones interbedded with siltstones. The Dave Elliot Bed is lithologically very similar to the exposure at Earlton, however the fauna is noticeably less diverse and lacks the distinctive coral taxa seen at Earlton. The fauna is

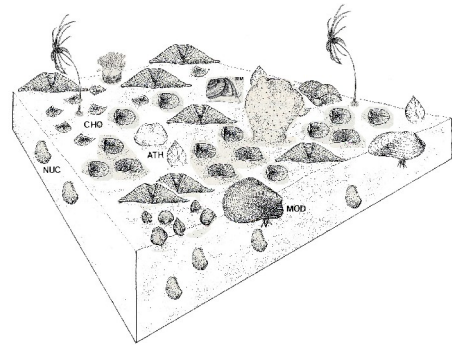


Figure 9. Paleoeological reconstruction for Buttermilk Falls.

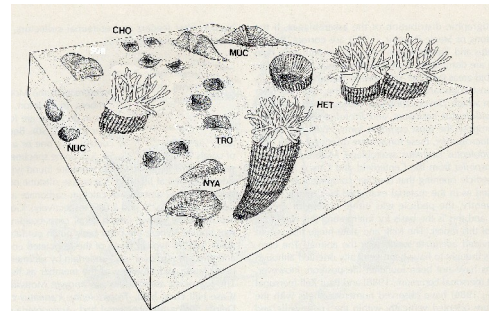


Figure 10. Paleoeological reconstruction for Earlton quarry. HET: *Heterophrentis*, MUC: *Mucrospirifer*, TRO: *Tropidoleptus*, NYA: *Nyassa*.

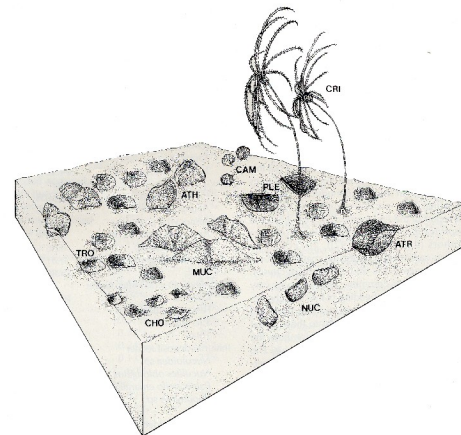


Figure 11. Paleoeological reconstruction for Stanton Hill Ravine. CRI: crinoids, ATR: *Pseudoatrypa*, CAM: *Camerotoechia*, PLE: *Protoleptostrophia*.

dominated by abundant *Tropidoleptus* and *Mediospirifer* brachiopods along with rare atrypids (Table 1, Figure 11).

The Dave Elliot Bed is next encountered in exposures along Gedney Hill Road, ~4 km north of Stanton Hill Ravine (Figure 5). The East Berne Member is moderately well-exposed along Gedney Hill Road from the Dave Elliot Bed down toward the bottom of the unit; glacial erosion has removed the upper East Berne Member above the Dave Elliot Bed sandstone with only till being found sitting directly on the bed. The Dave Elliot Bed here consists of ~1.5 m of resistant sandstone with fossils being scattered through the upper ~30 cm of the bed with a fauna dominated by *Tropidoleptus*, *Mucrospirifer*, and chonetid brachiopods along with rare *Heterophrentis* corals and conulariids (Table 1, Figure 12).

The next outcrop of the Dave Elliot Bed to the north (~10 km) from Gedney Hill Road is found in Hannacroix Ravine (Figure 5). Exposed within the ravine is a nearly complete section of the East Berne Member, however this section is markedly different from those to the south and multiple visits to the section were needed to determine the nature of deposition within the study interval. The East Berne Member has thickened quite a bit in this area and the lower and middle portions of the unit are distinctly finer-grained than sections to the south. The upper portion of the unit is approximately 25 m thick and contains a number of well-developed, coarsening-upward cycles, each approximately 3 m in thickness, that are capped by thicker and thicker sandstone beds with more and more diverse shell beds at the top of each cycle, continuing up to the exceedingly diverse Halihan Hill Bed. Additionally, the Dave Elliot Bed is not the lowest fossiliferous bed within the succession, mirroring the pattern seen at Rte. 209 in Kingston, except that there are three additional horizons of concentrated shells below the Dave Elliot Bed within the ravine. The Dave Elliot Bed is again here marked by a well-developed horizon of vertically-oriented concretions underlying the bed. The bed itself consists of a silty-sandstone with *Zoophycus* burrows and contains a moderately diverse fauna dominated by leiorhynchid, chonetid, and rarer spiriferid brachiopods along with conulariids (Table 1, Figure 13). The next shelly bed just below the Dave Elliot Bed in Hannacroix Ravine we have informally termed the Hannacroix Ravine Bed as it is well-developed and contains the zonally important goniatite *Tornoceras* aff. *mesopleuron* that pins this horizons to be just above the base of the Givetian stage within the Middle Devonian; previously this goniatite's lowest known occurrence was in the Halihan Hill Bed. The Hannacroix Ravine Bed contains a fauna dominated by nuculid bivalves with very rare leiorhynchid brachiopods and relatively common goniatites and nautiloids, much resembling the Dave Elliot Bed and lower bed at Rte. 209 near Kingston. The Hannacroix Creek Bed is marked by a zone of flattened concretions below it as well as two interesting 'gummy' clay horizons that are being currently investigated as to whether or not they may be bentonitic clay horizons.

Just ~5 km to the north and west of Hannacroix Creek is the type locality of the East Berne Member exposed along Cole Hill Road (Figure 5). Currently, the Dave Elliot Bed is covered at this locality; however Goldring (1935) de-

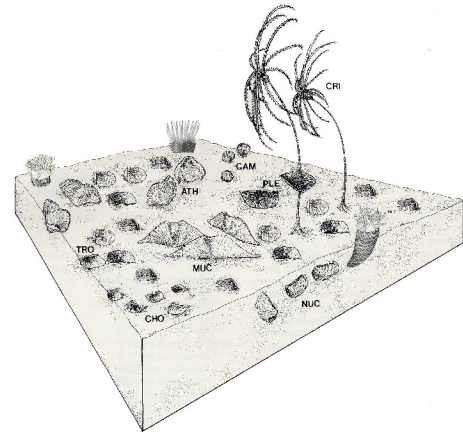


Figure 12. Paleocological reconstruction for Gedney Hill Rd.

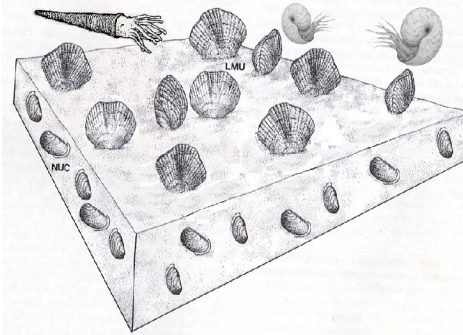


Figure 13. Paleocological reconstruction for the Dave Elliot Bed at Hannacroix Ravine.

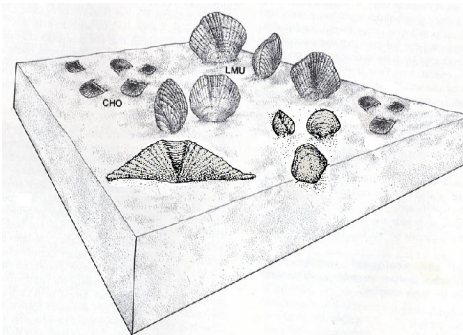


Figure 14. Paleocological reconstruction of the Dave Elliot Bed at Seneca Stone.

scribed the outcrop when the road was newly cut. It is possible from Goldring's detailed description of the lithology and fauna of the Halihan Hill Bed, which sits ~22 m above what was then the base of the outcrop, to estimate the level of the Dave Elliot Bed. The lowest ~6 m are described as relatively barren, blocky shales with concretions and rare nautiloids and goniatites. The next ~6 m were described as containing a fauna consisting of the brachiopods *Mucrospirifer*, *Mediospirifer*, *Arcuaminites*, *Nucleospira*, along with nuculid bivalves and *Modiomorpha* in blocky shales with thin sandstone beds becoming thicker near the top. Continuing up to the Halihan Hill Bed (~10 m) Goldring reports fossils becoming more abundant along with thicker sandstone beds, with the Halihan Hill Bed positioned within the upper 30 cm of a ~2 m sandstone. From her description of the section the Dave Elliot Bed most likely sits somewhere within the second 6 m of the section; this is consistent with patterns seen just to the south at Hannacroix Ravine. However, the fauna reported by Goldring for the hypothesized Dave Elliot interval at East Berne is markedly different from that observed at Hannacroix Ravine, containing a more diverse fauna composed of taxa interpreted to exist in somewhat shallower water conditions than those found at Hannacroix Ravine.

The Dave Elliot Bed has yet to be found north and west of Hannacroix Ravine throughout the outcrop belt of the East Berne Member except for a small exposure in the southwestern corner of the Seneca Stone Quarry south of Seneca Falls, a distance of ~200 km west of Hannacroix Ravine (Figure 5); a search is currently underway for exposures of the Dave Elliot Bed in the intervening area between these two widely separated areas. In between Hannacroix Ravine and Seneca Stone Quarry, the East Berne Member thins dramatically to little over approximately one meter in thickness and consists of dark-gray, fine-grained shale. The Dave Elliot Bed was discovered at Seneca Stone Quarry by Jeff Over of SUNY Geneseo and consists of a thin shell lag ~1-2 cm thick containing common chonetid, relatively common ambocoeliid and leiorhynchid and rare spirifereds brachiopods about 10 cm below the Halihan Hill Bed (Table 1, Figure 14).

DISCUSSION

Faunally, all of the taxa encountered within the Dave Elliot Bed to date have been elements of the Hamilton Fauna; no taxa of the underlying Stony Hollow Fauna have been encountered within the Dave Elliot Bed. With the existence of fossiliferous horizons below the Dave Elliot Bed within the East Berne Member (see discussion of Hannacroix Ravine section in road log), it is additionally possible to determine the content of the middle-lower East Berne in at least the Helderberg area and again, the fauna of this portion of the unit, though much less diverse than the overlying Dave Elliot and Hannacroix Ravine beds, contains no taxa distinctive of the underlying Stony Hollow Fauna. From these observations it is possible to postulate that the turnover between the Stony Hollow and Hamilton E.E. sub-Units is very rapid with little to no mixing of the two faunas.

With the identification of what is now the lowest known abundant and diverse example of the Hamilton Fauna it is possible to determine to what extent the extremely stable biofacies that typify the middle and upper Hamilton Group are developed (Figure 15), *i.e.*, how rapidly to the typical biofacies establish themselves within the basin after the rapid faunal turnover. Samples of the Dave Elliot Bed were compared with the typically developed sequence of biofacies seen through the Centerfield Cycle from the middle Hamilton Group (data from Gray, 1991) using Detrended Correspondence Analysis (DCA) to determine if the inferred sea floor

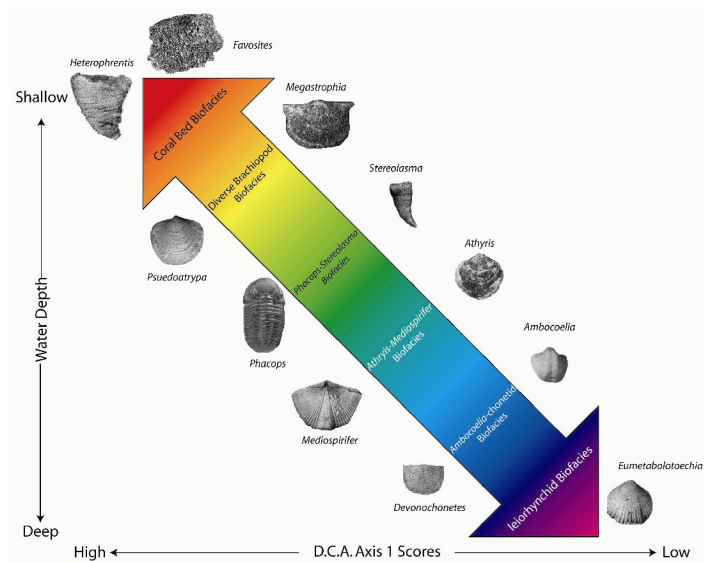


Figure 15. Diagram showing hypothesized gradient that existed along the sea floor during the Dave Elliot interval.

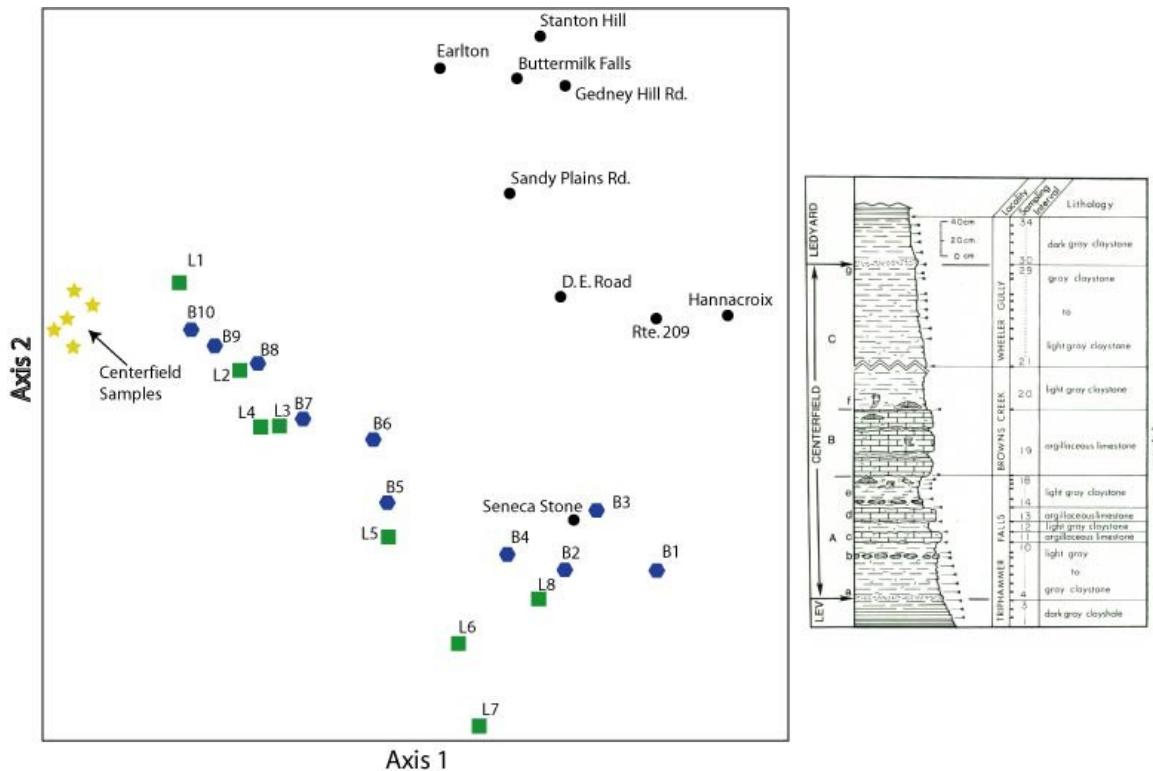


Figure 16. DCA plot for Dave Elliot Bed and Centerfield cycle samples. Stratigraphic column on right is for the Centerfield cycle (Gray, 1991). Black circles represent Dave Elliot Bed samples, blue hexagons represent the upper Levana Shale (LEV on column) samples of Gray, yellow stars represent the Centerfield Limestone samples of Gray, and green squares represent the Ledyard Shale samples of Gray.

gradient qualitatively hypothesized from field observations was statistically valid and how it compared to that seen through the Centerfield cycle.

Figure 16 shows the DCA plot for all the samples from the Dave Elliot Bed and the Centerfield cycle together. Previous analysis of the Centerfield cycle data has indicated that the samples shallow out of the Levana Shale into the Centerfield Limestone (shallowest water) and then deepen upwards through the overlying Ledyard Member (Brett et al., 2007). Within the DCA plot, this is shown as a distinct gradient along Axis 1, with the shallowest water samples plotting on one side of the axis and the deepest water samples The Dave Elliot Bed samples are separated off from the Centerfield cycle samples along axis 2, with the noted exception of the one sample of the Dave Elliot Bed from western New York (Seneca Stone). Along axis 1, the Dave Elliot Bed samples are arrayed from right to left in what was qualitatively determined to represent the deepest to shallowest water samples, thereby statistically supporting the field-based interpretation of a general shallowing from south to north, with a marked increase in water depth over a short distance between Gedney Hill and Hannacroix Ravine. The ‘compression’ of the Dave Elliot Bed samples along Axis 1 in relation to the Centerfield cycle samples and the separation of the two sets of samples along Axis 2 is most likely the result of the geographic location of the samples. The Hudson Valley area from which the Dave Elliot Bed samples were collected would have been much closer to the sources of sediment that were entering the basin during deposition and this probably had marked effects on which taxa could tolerate the higher amounts of sediment in the water column. It is interesting to note that the one Dave Elliot Bed sample that was taken in the vicinity of the Centerfield cycle samples plotted down amongst the Centerfield cycle samples along Axis 2, further supporting this interpretation. It is also possible that the shallowest Dave Elliot Bed sample (Earlton) was not as shallow as the core Centerfield cycle samples.

From these interpretations it is possible to now reconstruct the paleogeography of the shoreline within the basin during the time of deposition of the Dave Elliot Bed (Figure 17). From our analyses, we interpret an overall shallowing from Rte. 209 in Kingston to the samples from the Earlton Quarry. At Stanton Hill, just north of Earlton, no corals have been found and the sample is dominated by the brachiopod *Tropidoleptus*. Previous investigations have demonstrated that this association is consistent with shallow water-

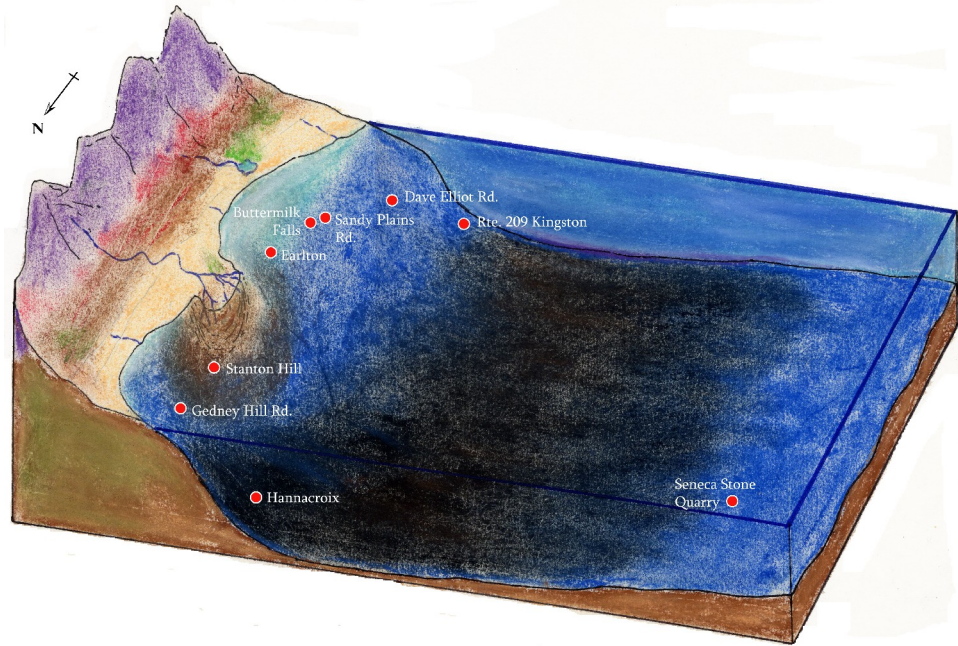


Figure 17. Paleogeographic reconstruction of the shoreline of the Appalachian Basin in the vicinity during deposition of the Dave Elliot Bed.

high sediment input, so Stanton Hill was most likely of a similar water depth as Earlton but was situated closer to a sediment source. Heading north from Stanton Hill we continue to see a pattern of high sedimentation; however the biofacies deepen somewhat towards Gedney Hill Road. Once north of Gedney Hill Road we see very rapid biofacies change with the next northerly sample from Hannacroix Ravine being exceedingly deep. This indicates the presence of some sort of paleovalley having existed on the sea floor in the vicinity of Hannacroix Ravine during sedimentation. This is further supported by the increased thickness of the upper East Berne Member in this area as there would have been additional accommodation space for sediment to have been deposited here.

CONCLUSIONS

The Dave Elliot Bed represents the lowest known occurrence of the Hamilton Fauna within the Appalachian Basin and is of earliest Givetian age. None of the warm-water incursion taxa of the underlying Stony Hollow Fauna were found within the Dave Elliot interval indicating a full and complete replacement of those taxa by cold-water adapted forms of the Hamilton Fauna. Additionally, this indicates a very rapid rate of faunal turnover and wholesale replacement of faunas within the basin with a very short interval of time. The typical biofacies of the overlying Hamilton Group were established throughout the water depth gradient and persisted with little if any change for the entirety of the Hamilton Fauna interval. The Dave Elliot Bed provides us with an important and interesting sample of time within a World-wide bioevent that helps to elucidate many factors that controlled faunal change through the interval.

ROAD LOG

Tot.	Int.	
0.0 mi.	0.0 mi.	Exit Plattekill Parking Lot at S.U.N.Y. New Paltz, turn RIGHT
0.05	0.05	Turn LEFT onto Rte. 32 heading North
0.25	0.2	Turn RIGHT onto Rte. 299 heading East
1.25	1.0	Turn RIGHT onto NYS Thruway entrance ramp, head North
5.75	4.5	Upper Ordovician Martinsburg Shale outcrops
9.75	4.0	Cross Rondout Creek
11.0	1.25	Lower Devonian Port Ewen Fm. outcrops
13.25	2.25	Lower Devonian Port Ewen, Alsen, and Becraft Fm.'s
14.75	1.5	Lower Devonian Schoharie Fm.
15.45	0.7	Lower-Middle Devonian Onondaga Fm.
15.85	0.4	Onondaga Fm
16.65	0.7	Cross Esopus Creek
16.75	0.1	Exit 19 for Kingston, get off Thruway
17.15	0.4	Turn RIGHT onto Rte. 28 heading West
18.0	0.85	Turn RIGHT onto Forest Hill Rd. then take immediate RIGHT onto City View Terrace

STOP 1 - City View Terrace. Our first stop this morning will be used to set up the stratigraphic and faunal story that will be the focus of today's trip. The outcrop along City View Terrace consists of shales and siltstones of the middle Union Springs Fm. and is the type locality for the base of the Stony Hollow Member with the upper-most portion of the underlying Bakover Member also being exposed (Figure 5). The contact between the two formations is marked by a layer of carbonate concretions as well as a discrete color change from dark-gray to black shales below to lighter-gray shales above; the concretions contain rare goniatites (*Cabrieroceras*) (Figure 18). This outcrop is a good example of the sediments that were deposited during the initial phases of the second tectophase of the Acadian Orogeny (Ettensohn, 1985). The downwarping of the edge of the craton due to back thrusting of the magmatic arc to the east created an over-deepened basin containing dysoxic to anoxic bottom waters allowing little to no benthic faunas to exist. Associated with this initial tectonic deepening is an eustatic sea level rise that brought warm, equatorial waters and associated faunas down to the south into the Appalachian Basin during this time; this is the local manifestation of the Kacak Bioevent. We see the first influx of warm-water adapted taxa in the upper Union Springs with the appearance of pelagic cephalopods previously unknown in much of the basin, such as *Cabrieroceras*. Additional taxa invade into the basin, displacing the preexisting cold-water adapted taxa, with a moderately diverse benthic fauna being present in the overlying Stony Hollow Member; this fauna has been termed the Stony Hollow Fauna (Figure 2). Today we will be primarily examining the interval that records a return to more 'normal' conditions for the Appalachian Basin near the end of the Kacak Bioevent showing a return of cold-water adapted taxa into the basin (Figure 4) associated with basin shallowing and associated sea level fall within the overlying East Berne Member of the Mt. Marion Fm.



Figure 18. Outcrop along City View Terrace. Contact between Bakover Member (below) and Stony Hollow Member (above) of Union Springs Fm. is at base of tree. Goniatite-bearing concretion at far right of picture marked by arrow.

18.0	0.0	Continue on City View Terrace to Potter Brothers' Ski Shop
18.1	0.1	Turn around in parking lot
18.3	0.2	Turn LEFT onto Forest Hill Rd. then take immediate LEFT onto Rte. 28 heading East
18.85	0.55	Turn RIGHT onto Rte. 209 entrance ramp heading North towards Rhinecliff
19.45	0.6	Middle Devonian Stony Hollow Member of Union Springs Fm.
20.75	1.3	Middle Devonian Otsego Member of Mt. Marion Fm.
20.8	0.05	Take exit for Sawkill Rd., turn LEFT, cross under Rte. 209 and immediately turn RIGHT heading back South on Rte. 209 towards Ellenville
22.25	1.45	Halihan Hill Bed in outcrop to right marking boundary between the East Berne Member (below) and Otsego Member (above) of Mt. Marion Fm.

STOP 2 - Route 209. At this stop we will be dropped off at the southern end of the outcrop and walk upwards through the section from the upper Stony Hollow Member, through boundary of the Union Springs and Mt. Marion Fm. at the base of the Cherry Valley Member, then into the overlying East Berne Member with a fine exposure of the Dave Elliot Bed, finishing up at the Halihan Hill Bed at the base of the Otsego Member where we will reboard the vehicles (Figure 5). Starting at the southern end of the outcrop we are in the upper siltstones and sandstones of the Stony Hollow Member (Figure 19). The unit is bioturbated and contains several shelly horizons including the lower and upper Proëtid beds containing a diverse example of the Stony Hollow Fauna (Figure 2). In the interval between the upper Proëtid Bed and the base of the Cherry Valley Member is the Hurley Member of Ver Straeten et al. (1994). This interval contains the distinctive goniatite *Agoniatites nodiferans*. Only the lower ~2 m of the Cherry Valley Member is exposed here consisting of thick-bedded sandstone with a depauperate fauna of dominantly leiorhynchid brachiopods (*Cherryvalleyrostrum*).

Sadly, the contact between the Cherry Valley Member and the overlying East Berne Member is covered here, but only for ~10 m or so. The lowest portions of the East Berne Member are medium-gray, fine-grained, very fissile shales that grade upwards with more and more silts. Near the middle of the member, we see the entrance of thicker-bedded siltstones and fine-grained sandstones. These beds are very homogenous inside and show little to no burrowing in nearly every instance at this outcrop. We interpret these beds to be storm-derived sediments that were most likely deposited nearly instantaneously into deep, dys-oxic waters.

Continuing up section we first come to a distinct rusty horizon that contains an incredible abundance of infaunal nuculid bivalves (*Nuculoidea*) (Figure 20).

Approximately 2 m above this bed is a distinctive horizon of vertically-oriented concretions formed around tubular pyritic burrows. This bed is ~1 m below the Dave Elliot Bed, another rusty horizon with abundant nuculid bivalves and rare leiorhynchid brachiopods and cephalopods. This bed represents a 'dirtier' water version of the leiorhynchid biofacies of Brett et al. (2007) (Figure 15). Con-



Figure 19. Southern portion of outcrop along Rte. 209 just west of Kingston exposing Stony Hollow Member (below), Hurley Member (middle) and Cherry Valley Member (above). Students are collecting from the upper Proëtid Bed within the Stony Hollow Member. White line denotes base of Cherry Valley Member.

tinuing upwards we see thicker and thicker siltstone and sandstone beds until near the northern end of the outcrop we encounter the diverse Halihan Hill Bed that contains abundant spiriferid brachiopods and common small rugosan corals.



Figure 20. Rte. 209 outcrop. Arrows point to distinctive fossiliferous horizons. Upper arrow points to the Dave Elliot Bed.

23.25	1.0	Stony Hollow Member
23.85	0.6	Take exit for Rte. 28 heading East
24.85	1.0	Go ¾ the way around the traffic circle and exit onto NYS Thruway and head North towards Saugerties
25.55	0.7	Stony Hollow Member
27.75	2.2	Cross Sawkill Creek
27.95	0.2	East Berne Member
30.55	2.6	Cross Plattekill Creek
31.0	0.45	Otsego Member
32.25	1.25	Mt. Marion to west of Thruway
34.05	1.8	Big Quarry in East Berne Member on north end of Mt. Marion
34.75	0.7	Take Exit 20 for Saugerties, outcrops of Onondaga Fm. just beyond exit ramp
35.15	0.4	Stay in middle lane after going through toll booth
35.25	0.1	Turn LEFT at traffic light and immediately enter Stewart's parking lot REST STOP at Stewart's/MacDonald's
35.25	0.0	Turn RIGHT out of Stewart's parking lot, stay in Right lane
35.45	0.2	Turn RIGHT onto Rte. 32 North
36.35	0.9	Onondaga Fm.
37.25	0.9	Turn RIGHT onto King's Highway, Co. Rte. 34
38.25	1.0	Kaatsban Stone Church (Kaatsban is Dutch for tennis court)
38.65	0.4	Turn LEFT onto Dave Elliot Rd.
40.15	1.9	Dave Elliot Road Quarry. Pull into quarry and park

STOP - 3 Dave Elliot Quarry. Stop 3 is the type locality of the Dave Elliot Bed along Dave Elliot Road (Figure 5, 21). This small quarry exposes ~10 m of the upper East Berne Member and dip markedly to the west. The Dave Elliot Bed consisting of a ~20 cm siltstone bed containing an abundance of chonetid brachiopods exposed near the eastern end of the quarry, holding up a small terrace; many of the brachiopods in the bed are distinctly white in color at this locality. Also present in the fauna of the bed here are rare to common spirifers and common emmanuellid brachiopods. The biofacies preserve here would be consistent with the chonetid biofacies of Brett et al. (2007) (Figure 15).

40.15 0.0 Continue West up hill on Dave Elliot Rd.

NYSGA 2009 Trip 10 - Bartholomew, Schramm, Ver Straeten and Over

- 40.35 0.2 Otsego Member
- 41.25 0.9 Turn RIGHT at 'T' onto High Falls Rd.
- 42.35 1.1 Otsego Member
- 42.75 0.4 Turn RIGHT, staying on High Falls Rd.
- 43.05 0.3 Stay STRAIGHT on Nelson Rd.
- 43.25 0.2 Cross Kaaterskill Creek just above High Falls, take immediate LEFT after bridge and bear LEFT onto Mossy Hill Rd.
- 43.35 0.1 Otsego Member
- 46.15 2.9 Turn RIGHT onto Rte. 23A
- 46.75 0.6 Stony Hollow and Cherry Valley members, base of Cherry Valley Member marks the base of the Mt. Marion Fm.
- 47.0 0.25 Stony Hollow Member
- 47.5 0.5 Upper Proetid Bed of Stony Hollow Member
- 48.1 0.6 Turn RIGHT at Stop Sign
- 48.7 0.6 Turn LEFT onto Vedder Mt. Rd.
- 49.7 1.0 Stony Hollow Member
- 51.2 2.5 Turn LEFT onto Co. Rte. 47
- 51.7 0.5 Cross Rte. 23
- 51.9 0.2 Turn RIGHT onto Co. Rte. 23B
- 52.2 0.3 Turn LEFT onto Green Lake Rd., Co. Rte. 49
- 52.7 0.5 Bear LEFT onto Sandy Plains Rd.
- 53.45 0.75 Pull over on side of road. Be careful as there is minimal shoulder here.



Figure 21. Dave Elliot Bed at Dave Elliot Rd.

STOP 4 - Sandy Plains Road. The Dave Elliot Bed at Sandy Plains Road (Figure 5) is exposed at the top of a ~1.5 m sandstone bed and is dominated by crinoidal debris, abundant spiriferid brachiopods along with with rare bryozoans. It is interesting to note that the upper portion of the East Berne Member has coarsened considerable in this area, with the sandstone containing the Dave Elliot Bed thickening by almost an order of magnitude (Figure 22). The biofacies preserved in the bed here most closely resemble a moderately diverse brachiopod assemblage of the Hamilton Group (Figure 15).



Figure 22. Dave Elliot Bed at Sandy Plains Rd.

53.45	0.0	Continue West on Sandy Plains Road
53.95	0.5	Otsego Member
54.0	0.05	Turn RIGHT onto Potic Mt. Rd.
55.95	1.95	Turn RIGHT onto Buttermilk Falls Rd.
56.7	0.75	Pull off for Buttermilk Falls

LUNCH STOP - Butter Milk Falls. Our lunch stop today is a picturesque spot at the Buttermilk Falls preserve just north of Leeds (Figure 5, 23). Exposed in the stream is the uppermost portion of the Stony Hollow Member and the overlying Cherry Valley Member. Continuing upstream the East Berne is exposed on private property and consists of silty shales with numerous thin-bedded sandstones. The main face of the falls where we will eat lunch is the upper Proëtid Bed and the distinctive brachiopods *Variatrypa arctica* and *Pentamerella* cf. *wintereri* have been found here representing the Stony Hollow Fauna. The smaller falls just upstream from the main falls is the Cherry Valley sandstone and is distinctly cleaved at this locality. The Dave Elliot Bed is present upstream and forms another ~4 m waterfalls and very closely resembles that seen at Sandy Plains Road just to the south (Figure 24).

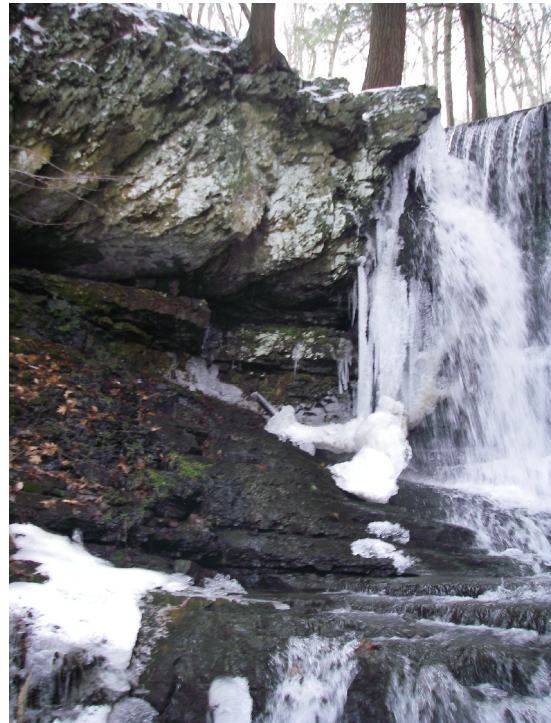


Figure 23. Upper Stony Hollow Member at Butter Milk Falls.

56.7	0.0	Continue down hill on Butter Milk Falls Rd.
57.0	0.3	Turn LEFT onto Co. Rte 49, Green Lake Rd.
57.4	0.4	Stony Hollow Member
58.0	0.6	Stay straight (left) on Co. Rte. 49
59.0	1.0	Onondaga Fm.
59.2	0.2	Turn LEFT, the immediately RIGHT, the bear LEFT
60.0	0.8	Turn RIGHT onto Lime Street
61.2	1.2	Onondaga Fm.
62.3	0.1	Turn LEFT at Stop Sign
62.4	0.1	Turn RIGHT onto Bronk Rd.
63.1	0.7	Continue STRAIGHT through Stop Sign
63.4	0.1	Cross Rte. 81, continue STRAIGHT onto Vanderberg Rd.
64.4	1.0	Turn LEFT onto Co. Rte. 26 then immediately turn RIGHT onto Lime Kiln Rd. Onondaga Fm. outcrops all along this road
66.6	2.2	Turn LEFT onto Reservoir Rd.
66.8	0.2	Turn RIGHT onto Valley View Rd.
67.1	0.3	Turn LEFT onto Co. Rte. 54
67.7	0.6	Turn RIGHT onto Jennings' Rd.
69.2	1.5	Stay STRAIGHT (right) onto Co. Rte. 51
70.7	1.5	Turn LEFT onto Shady Lane
71.45	0.75	Turn RIGHT onto Gedney Hill Rd.
72.2	0.75	Quarry on left, Albany Co. line
73.2	1.0	Pull off to LEFT into small parking area at base of hill



Figure 24. Dave Elliot Bed capping a waterfalls on Butter Milk Falls Creek

STOP 5 - Gedney Hill Road.

Stop 5 today is an exposure of the middle portion of the East Berne Member along Gedney Hill Road (Figure 5). We have skipped what are perhaps some of the most interesting exposures of the Dave Elliot Bed in the intervening area between Buttermilk Falls and here, however access to these exposures is limited at this time. We will walk up through the entirety of the exposure of the East Berne Member here as parking is limited. The Dave Elliot Bed is here represented by a ~1 m very hard quartz-rich sandstone bed (Figure 25) and contains an abundant fauna of brachiopods (*Athyris*,



Figure 25. Dave Elliot Bed at Gedney Hill Rd.

Mediospirifer, *Mucrospirifer*, *Tropidoleptus*) and rare corals and conulariids and would sit somewhere between the diverse brachiopod biofacies and the coral bed biofacies in the parlance of Brett et al. (2007) (Figure 15). In between here at Butter Milk Falls are the Earlton and Stanton Hill Ravine sections. The Dave Elliot Bed at Earlton contains the most diverse fauna yet discovered including rare tabulate corals, moderately-common rugosan corals, and abundant spiriferid and athyrid brachiopods. This biofacies is the closest we get within the interval to the coral bed biofacies of Brett et al. (2007) (Figure 15). The outcrop of the Dave Elliot Bed at Stanton Hill Ravine is similar in lithology to the exposure at Gedney Hill Road, however the fauna is less diverse (entirely lacking in corals) and almost entirely dominated by *Tropidoleptus*. This association would sit on the ‘dirtier’ side somewhere between the diverse brachiopod and *Athyris-Mediospirifer* biofacies of Brett et al. (2007) (Figure 15).

- 73.2 0.0 Continue down hill (north) on Gedney Hill Rd.
- 73.7 0.5 Take hard LEFT onto Rte. 143 heading North
- 78.9 5.2 Turn RIGHT onto Rte. 32 heading NORTH
- 82.6 3.7 Turn LEFT onto Co. Rte. 301
- 82.7 0.1 End of Gulch Cave to right
- 82.8 0.1 Onesquethaw Cave to left
- 83.2 0.4 Turn RIGHT continuing on Co. Rte. 301
- 83.5 0.3 Onondaga Fm.
- 84.7 1.2 Cross Onesquethaw Creek; nice fold in creek in upper Onondaga Fm.
- 84.9 0.2 Turn LEFT onto Rte. 443 heading West
- 85.0 0.1 Clarksville Cave
- 85.2 0.2 Cross Onesquethaw Creek
- 85.3 0.1 Onondaga Fm.
- 86.9 1.6 Turn LEFT into Stewart’s parking lot
REST STOP at Stewart’s
- 86.9 0.0 Turn RIGHT out of Stewart’s onto Rte. 443 heading East
- 88.6 1.7 Turn RIGHT onto Cass Hill Rd.
- 89.6 1.0 Otsego Member
- 90.0 0.4 Otsego Member
- 90.5 0.5 Turn LEFT into Nature Conservancy parking area for Hannacroix Ravine

STOP 6 - Hannacroix Ravine. The last stop today will be at Hannacroix Ravine (Figure 5). We consider this locality is somewhat of a ‘Rosetta Stone’ of sorts for the East Berne Member as it contains numerous individual fos-

siliferous horizons in addition to a number of well-developed small-scale cycles throughout the member. The abundance of thin, shelly horizons within this exposure confused us at first as to the proper identification of the true Dave Elliot interval, however this led to the discovery of some very important fossils that might have otherwise gone unnoticed. The actual Dave Elliot Bed consists of a ~0.75 m thick silty sandstone with *Zoophycus* burrows and a fauna dominated by spirifered brachiopods and conulariids and is underlain by the distinctive vertically-oriented concretions; the bed itself caps a small falls within the creek (Figure 25).



Figure 26. Author sitting on Dave Elliot Bed in Hannacroix Ravine.

We have, to date, found nine distinct fossiliferous horizons within the middle to upper East Berne Member at this locality (Figure 27, 28). Besides the Dave Elliot Bed, the horizon with the most well-developed fauna forms a ~30 cm interval with dispersed abundant nuculid bivalves, rare leiorhynchid brachiopods, and common nautiloid and goniatite cephalopods. This interval is underlain by two distinct ‘gummy’ clay horizons (possible bentonites, see discussion in main body of text). We have provisionally named this interval the Hannacroix Ravine Bed. Of great importance was the discovery at this site of the zonally important goniatite *Tornoceras* aff. *mesopleuron* (Figure 29, graciously identified for us by Dr. R. Thomas Becker of Muenster, Germany). At the standard section for the base of the Givetian in Morocco, this taxa first appears just above the base of the *P. hemiansatus* conodont zone that marks the base of the Givetian Stage and the occurrence of this taxa in the Dave Elliot Bed here marks its lowest record in eastern North America to date and helps to establish the base of the Givetian Stage as occurring somewhere below this bed and above the top of the Cherry Valley Member which contains conodonts indicative of the upper Eifelian (*T. kockelianus* Zone) and puts the lower East Berne in the *P. ensensis* Zone (Figure 30).

Between the Dave Elliot Bed and the Halihan Hill Bed at the top of the East Berne, there are at least four distinct fossiliferous horizons, each becoming progressively more diverse as you proceed up-section. Between the Hannacroix Ravine Bed and the Dave Elliot Bed are two, thin fossiliferous horizons dominated by scattered nuculid bivalves. There are also two nuclulid-dominated horizons below the Hannacroix Ravine Bed that can be traced through the ravine and out onto the road where the lowest forms a ~2 m falls in a side gully entering from a ditch to the south of the main ravine; the upper of the two beds is visible further up the ditch to the south. A more detailed investigation of the entire East Berne Member is planned for this section. To date no fossiliferous horizons have been found downstream of the road culvert, although a picturesque waterfall is present about 100 yards downstream providing an accessible lower section that needs more work.

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END OF TRIP

90.5	0.0	End of trip, return to New Paltz – Turn RIGHT out of parking lot onto Cass Hill Road
92.4	1.9	Turn RIGHT onto Rte. 443, cross Onesquethaw Creek
92.7	0.3	Turn RIGHT onto Co. Rte. 301
94.2	1.5	Turn LEFT continuing on Co. Rte. 301
98.7	4.5	Stay STRAIGHT continuing onto Co. Rte. 396
103.3	4.6	Cross Rt. 9W, stay on Co. Rte. 396
105.3	2.0	Turn LEFT onto Rte. 144 at 'T' intersection
105.7	0.4	Turn LEFT onto NYS Thru-way entrance ramp, head South
165.2	59.5	Take Exit 18 for New Paltz
166.1	0.9	Turn LEFT at light onto Rte. 299 heading West
167.1	1.0	Turn LEFT at light onto Rte. 32 heading South
167.3	0.2	Turn RIGHT onto Plattekill Ave. then immediately LEFT into parking lot

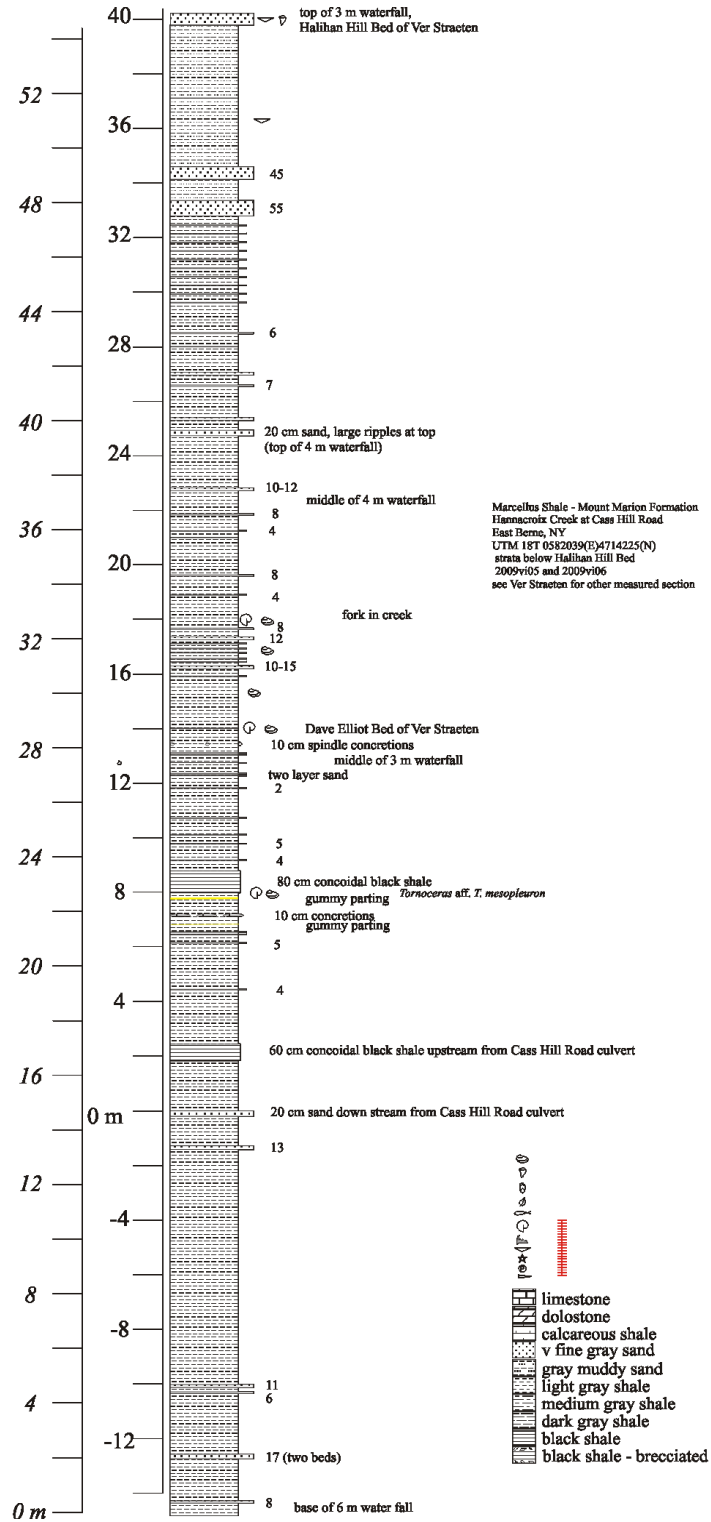


Figure 27. Stratigraphic column for Hannacroix Ravine (by Jeff Over and students).

NYSGA 2009 Trip 10 - Bartholomew, Schramm, Ver Straeten and Over

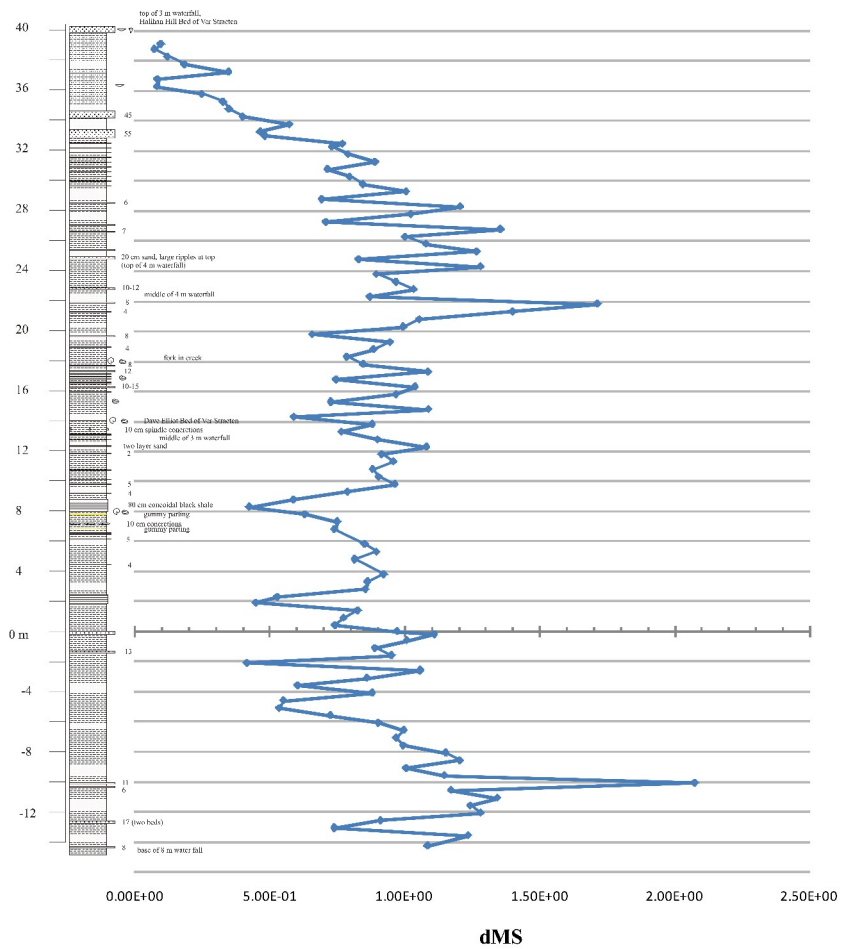


Figure 28. Magnetic susceptibility curve for Hannacroix Ravine section by Jeff Over and students.



Figure 29. Goniatite *Tornoceras* aff. *mesopleuron* from the Hannacroix Ravine Bed. Full specimen is ~5 cm across.

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Stage	Conodont zones	Goniatite faunas	Unit
Givetian	<i>semialternans</i>	<i>P. amplexum</i>	up. Tully
	<i>ansatus</i>		mid. Tully
		<i>T. uniangulare</i>	low. Tully
			Windom
			Portland Pt.
	<i>rhenanus/ varcus</i>	<i>M. n. sp./S. unilobatus</i>	Jaycox
		<i>T. amuletum/T. ualdenense</i>	Wanakah
	<i>timorensis</i>		Ledyard
			Centerfield
	<i>hemiansatus</i>	<i>T. arkonense</i>	Butternut
		Pompey	
<i>Parodiceras/T. mesopleuron</i>		Delphi Stat.	
? - ?		Stafford	
	<i>ensensis</i>	Chittenango/ Cardiff	
	<i>kockelianus</i>	---D---E---B---	
	<i>australis</i>	Berne	
Eifelian		<i>A. vanuxemi</i>	Cherry Valley
		<i>C. plebeifome</i>	Hurley
			Bakoven

Figure 30. Current understanding of goniatite and conodont biostratigraphy for the Middle Devonian of New York.

The Classic Barrovian Metamorphic Sequence of Dutchess County and Its Structural and Stratigraphic Context in the Taconic Orogeny

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Introduction

Metamorphism in Dutchess County, NY has been considered a classic example of Barrovian-type metamorphism since early mapping of the region (Knopf, 1927; Balk, 1936; Barth, 1936). Interestingly enough, few workers have extended the work that Balk and Barth began. People have been willing to accept it as classic without further testing (notable exceptions are the work of Bence, Vidale, Whitney and their coworkers). In his 1998 NYSGA field guide to Dutchess County, Donnelly remarks that "The area remains one of the clearest Barrovian sequences in the world." And yet, there are many unanswered questions about the metamorphism. Some questions arise because of the complexities of the stratigraphy and the timing of episodes of deformation, metamorphism, and fluid flow. Others arise because of new analytical techniques that can be brought to bear to help elucidate the complex metamorphic history of the rocks.

The metamorphic sequence in Dutchess County begins with chlorite grade rocks at the Hudson River (Figure 1; Stop 1 in this guide; all pelitic rocks east of the Shawangunk Ridge in Ulster County contain chlorite, although whether detrital or authigenic/metamorphic is not always clear) and progresses beyond the second sillimanite (sillimanite-orthoclase) isograd near the Connecticut border (Stop 8). Rocks now exposed over a distance of about 30 km were tectonically buried as much as 25 km (Bence and McLelland, 1976; Whitney et al., 1996b) during the Taconic Orogeny. The sequence was subsequently tilted to its present orientation during uplift to the east associated with the Acadian Orogeny. In addition to isograd mineral growth, the prograde sequence also demonstrates fabric development from phacoidal cleavage at the Hudson River, through phyllites and schists in the middle of the county, to gneissic fabric near the Connecticut border.

Balk (1936) and Barth (1936) believed that metamorphism in Dutchess County was due to "steeping" of the pelitic rocks in igneous fluids emanating from magmas deeper within the crust. They cited as evidence the fact that granitoid intrusions are found to the east and south (in Connecticut and Westchester County, NY, respectively), on strike with the increasing metamorphic gradient mapped in the county. Some of these granites, however, contain muscovite suggesting that they are peraluminous S-type granitoids formed by partial melting of a source dominated by sedimentary material (Winter, 2001, p. 350). Some outcrops in the sillimanite zone of Dutchess County contain Qtz-fsp-mica veins which Barth (1936) interpreted as apophyses from the granitoid intrusions. These veins have also been interpreted as leucosomes from anatexis of the sediments (Bence and McLelland, 1976; Whitney, personal communication, 2009). Most workers now believe that the heat for metamorphism was generated by tectonic burial of the rocks during the Taconic Orogeny although fluid flow was probably an important heat-transfer mechanism (Whitney et al., 1996b).

The classic nature of the metamorphic sequence in Dutchess County, and its excellence for teaching, mean that a number of very good field guides are available for the area (Bence and McLelland, 1976; Donnelly, 1998; Whitney and Peck, 2004). Bence and McLelland's 1976 NYSGA field guide provided a thorough discussion of the sequence, and identified many of the outcrops that have since become almost mandatory on local field trips (Stops 2, 4, 6, 7 and 9 in this guide). They also included whole rock chemical analyses, and microprobe analyses of individual minerals that are very helpful for understanding metamorphic conditions. Additional sources of whole rock analyses are Vidale (1974b), Whitney et al. (1996b) and Donnelly (1998). Whitney et al. (1996b) also published microprobe analyses of individual mineral grains.

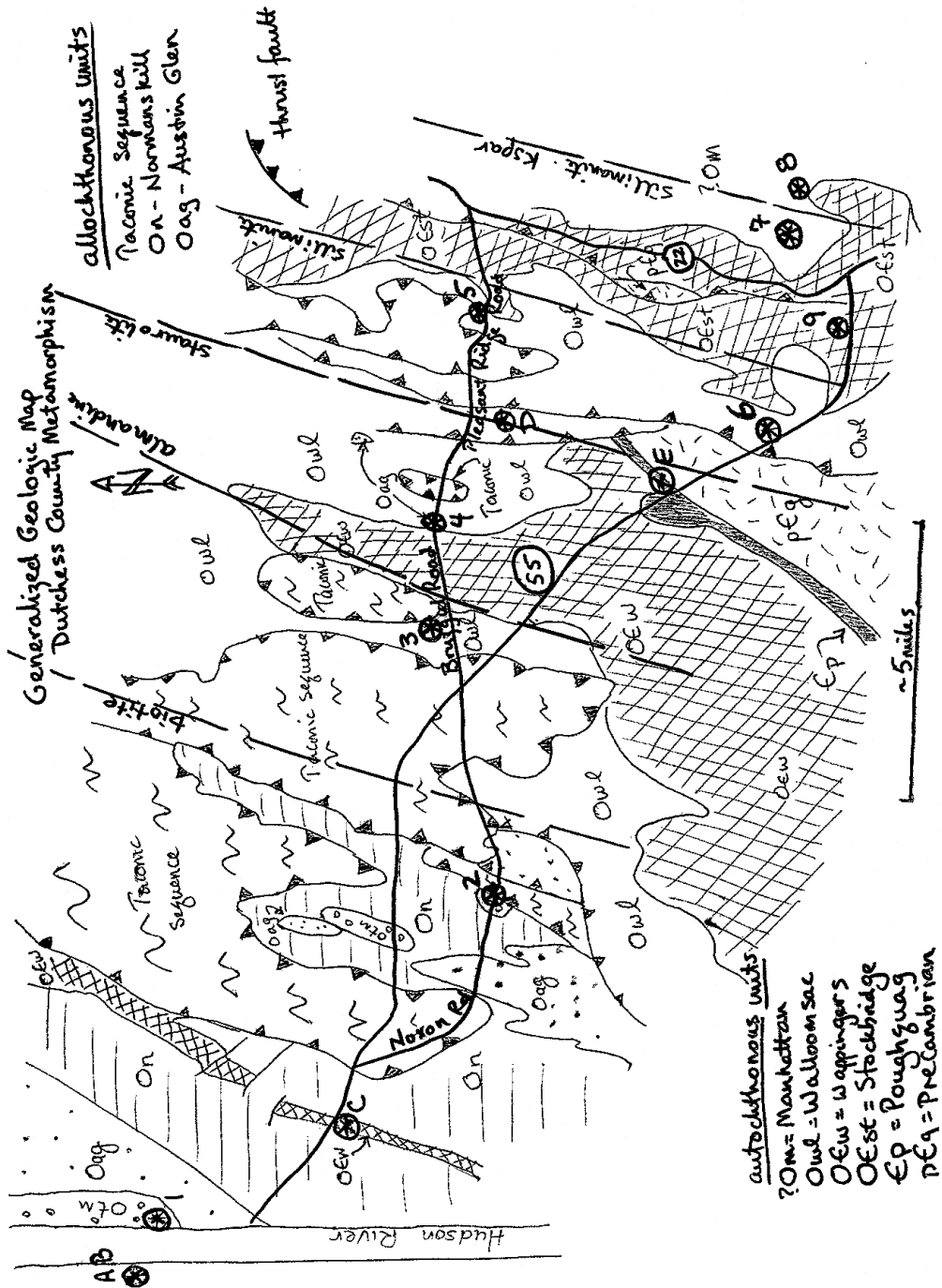


Figure 1. Generalized geologic map of Dutchess County showing main stratigraphic and structural features, metamorphic isograds, and stop location. (after Fisher et al., 1970).

This field trip will follow Bence and McLelland (1976) and Whitney and Peck (2004) for the higher grade (chlorite grade and higher) outcrops. To this, we add several additional topics:

- We will make stops in the lower part of the chlorite zone at which we discuss the evidence for low grade metamorphism and the difficulties in distinguishing metamorphic from detrital chlorite
- We stop at exposures of the Taconic melange and discuss its significance.
- We comment on some recent proposals relevant to the sometimes confusing stratigraphic nomenclature.
- We comment on the development of metamorphic fabrics across the county.
- We provide photomicrographs for discussion of relationships that are difficult to examine in outcrop.
- We include additional equations and AFM projections to illustrate possible reactions.

Stratigraphic Nomenclature

We do not propose revisions to the stratigraphy of Dutchess County, but note that some terms are the subject of confusion which has been discussed in the literature. In particular the terms “Snake Hill Shale”, and “Normanskill” Formation or Group cause confusion, in part because of the stratal disruption of these units along the Taconic front, but also because formational names have been applied to both biostratigraphic and lithostratigraphic units. Similarly, usage of the term “Taconic melange” requires clarification.

Dark shales and slates associated with the Taconic front and further west are sometimes referred to as "Snake Hill Shale" (e.g., Fisher and Warthin, 1976; Fisher, 1977). In Dutchess County shales in this stratigraphic position are also mapped as the Normanskill Formation (On), although Rickard and Fisher (1973) argued, and most investigators now agree, that that Normanskill Formation is entirely allochthonous. The term "Snake Hill Shale" is also sometimes used to refer to the matrix of the Taconic melange (Otm) and has been used to refer to exposures of Taconic melange matrix in the Poughkeepsie area (e.g., outcrops in the Arlington and Spackenkill neighborhoods of the Town of Poughkeepsie near Vassar College). Kidd et al. (1995) and English et al. (2006) discourage this broad usage of the term "Snake Hill Shale" (also pers. comm. G. Baird, 2009). We accept this restricted usage, which regards the Snake Hill as coherent stratigraphic blocks of lithologies similar to the "type" example bordering Saratoga Lake. These rocks have been caught up in Taconic thrusts, and are likely blocks with melange.

Similarly, the type locality of the “Normanskill” Formation (or Group) at Normanskill Gorge south of Albany is a block within the Taconic melange (Vollmer, 1981), where it comprises Austin Glen aspect graywacke and shale. However the term Normanskill Formation was extended from there to include Austin Glen (greywacke and shale), Mount Merino (black and green chert and siliceous argillites), and later the Indian River (red slate) as members (Ruedemann, 1942; Fisher, 1977), and later as formations when Normanskill was elevated to Group status (Fisher, 1977; see discussions in Vollmer, 1981; Cunningham, 1991; and Kidd et al., 1995). -We favor the proposal of Kidd et al. (1995) to abandon using the term “Normanskill” as a group or formation name, as it has strong biostratigraphic connotations and has been widely applied to very diverse lithologies.

The Austin Glen “type” locality is near Catskill, New York and it is widely exposed along the west side of the Hudson River (Vollmer, 1981; Marshak, 1990; Cunningham, 1991) including large exposures at the Mid-Hudson Bridge (Cunningham, 1991; Optional Stops A and B) directly west of Poughkeepsie. These rocks are highly deformed and are in most cases clearly tectonically transported (Vollmer, 1981; Cunningham, 1991). They are often associated with Mount Merino, but depositional contacts are not preserved. Generally, exposures of the Austin Glen and Mount Merino (many are on the west side of the Hudson River) are not considered part of the Taconic Allochthon as traditionally defined, because they lie west of Logan's (or Emmon's) line (e.g., Fisher, 1977), however structurally they are essentially lower thrust sheets (“slices”) (Bosworth and Vollmer, 1981).

To the west, deformation in Austin Glen and related flysch formations (e.g., Schenectady Formation) decreases, and they overlie the carbonate shelf sequence (Fisher, 1977). The term Pawlet has been used for Austin Glen equivalent rocks in the northern Taconics (Zen, 1961) where it has been shown to be conformable with the underlying Mount Merino, Indian River and Poultny Formations of the Taconic sequence (Rowley et al, 1979). The westerly prograd-

ation of the Austin Glen/Pawlet flysch over the slope/shelf sequence has been explained as due to the migration of a foreland basin as the Taconic Allochthon progressed westward over the continental margin and onto shelf sequence (Bird and Dewey, 1970; Rowley and Kidd, 1981). Note that “Austin Glen” has been used to include both flysch deposited on the continental shelf in front of, and subsequently overridden by, the Taconic Allochthon, and flysch deposited on the allochthonous Taconic continental rise sequence (i.e., Mount Merino and Indian River) and transported with it.

Within Dutchess County, metapelites mapped as conformable with the underlying Wappinger Group limestones are referred to as the Walloomsac Formation (Fisher, 1977), and are thus considered to be exposed structurally within windows beneath the Taconic Allochthon, or brought up as thrust slices along with the underlying shelf carbonate and Precambrian basement. The Walloomsac is thought to be correlative with flysch sequences deposited west of the allochthon, as well as the Manhattan schist (Fisher, 1977), mapped in eastern Dutchess County.

The Taconic melange is not a stratigraphic unit, but rather a complex tectonic unit consisting of centimeter to hundred meter scale blocks of various lithologies, dominated by graywacke, within a phacoidally cleaved scaly matrix. The blocks include stratally disrupted sequences (“broken formations”) as well as olistostromic components. The melange occurs in distinct zones, which may be meters to kilometers across, and which represent high strain zones associated with the emplacement of the Taconic Allochthon (Vollmer, 1981; Vollmer and Bosworth, 1984; Caine, 1991; Plesch, 1994; Kidd et al., 1995). Kidd et al. (1995) discuss other terms used for this unit, including “Poughkeepsie Melange” proposed by Fisher (Fisher & Warthin 1976, Fisher 1977), and propose the new name “Cohoes Melange” be used with the well-exposed sections along the Mohawk River as “type” sections (Plesch 1994). The Taconic melange is overlain by the Silurian-Devonian Helberburg Group along the Taconic unconformity south of Albany (Vollmer 1981; Kidd et al., 1995). Outcrops of melange matrix are widely exposed in the Poughkeepsie area. The typical irregular phacoidal cleavage (Stop 1) becomes more slaty in appearance to the east, and it becomes more difficult to discern small blocks and fragments within the melange.

Geologic History

The geologic history of the Dutchess County area, and of the whole Taconic belt that occupies much of eastern New York, is one of sedimentation, tectonic collision, and metamorphism. Ages of the various sedimentary rocks are constrained by graptolite zones, environments of deposition have been interpreted from textures, and provenances deduced by detrital clasts and geochemistry. The paleogeography, however, is less certain for it is very complex, and no one interpretation is unanimously accepted. In this field guide we follow the paleogeographic reconstructions of Rowley and Kidd (1981) shown in Figure 2.

In brief, the geologic history of Dutchess County is as follows (with formation names and abbreviations as shown on the Lower Hudson Sheet of the state geologic map (Fisher et al., 1970). Precambrian rifting of the Grenville-age supercontinent resulted in formation of a rifted margin sedimentary environment on the edge of the Iapetus Ocean (Figure 2A and B). Initial Paleozoic deposition in rift basins of the Grenville crust was the Poughquag Quartzite (Cpq). Carbonate platform deposition then ensued, forming the thick limestones and dolostones of the Wappinger Group (OCw, probably correlative with the Stockbridge Marble (OCst) in the eastern part of the county). During this same time interval, fine-grained clastic sediments of the Everett (Eev), Nassau (En), Germantown (Eg), Stuyvesant Falls (Osf), Mount Merino and Indian River (Omi) formations were deposited on the continental rise.

Eastward dipping subduction of the oceanic crust resulted in closing of the Iapetus Ocean basin. As the subduction zone approached ancestral North America (Figure 2C), the sediments on the continental slope and rise were incorporated into the accretionary wedge causing imbrication of the rise sediments by thrust faulting while flysch continued to be deposited in the migrating basin. Austin Glen/Pawlet (Oag) was deposited conformably on the Taconic continental rise sequence while Walloomsac (Owl) was deposited on shelf carbonates farther west.

Continued collision caused the obduction of the rise sediments onto the continental shelf carbonate and shale sediments, along with the already deposited Austin Glen/Pawlet flysch (Figure 2D). Austin Glen flysch continued to be deposited although the locus of deposition moved continuously west in advance of the collision zone. At the same time, previously deposited flysch sediments were incorporated into the accretionary wedge, further complicating the structural relationships. Obduction of the continental rise depositional sequence completely onto the shelf sequence

formed the present-day Taconic Allochthon (Figure 2E). At this time metamorphism of the rise sediments occurred, accompanied by the development of a slaty cleavage.

Austen Glen and related flysch (e.g., Utica, Schenectady) continued to be deposited to the west of the accretionary wedge. The climax of the Taconic Orogeny came when Grenville basement contacted the arc basement (Figure 2F). At this time basement faults, including some formed during the initial rifting of the Grenville supercontinent, were reactivated as reverse faults uplifting basement complexes such as the Berkshire, Housatonic, and Hudson Highlands massifs and transporting them to the west. The entire mass was then metamorphosed during the Taconic Orogeny, forming the Barrovian sequence we will see today. However, textural evidence of two episodes of deformation (folded cleavages, folded veins, and crenulation cleavages), are common in rocks of the Taconic Allochthon, the first fabric may have formed during obduction, the second during regional metamorphism.

Geochronology

Based on Rb-Sr and K-Ar studies, Long (1962) calculated a minimum age of 430 Ma for metamorphism during the Taconic Orogeny, followed by regional reheating at about 360 Ma. Taylor et al. (1999) used Lu-Hf systematics on garnet-whole rock pairs to date the metamorphism at around 440 Ma. The range of ages they calculated, along with data from both the core and rim of a garnet, suggested that prograde garnet growth occurred in only about 10 million years (446 Ma core, 436 Ma rim), implying a relatively short metamorphic event.

Metamorphism and P-T estimates

Metamorphic grade increases to the east-southeast across the county from practically unmetamorphosed shale at the Hudson River at Poughkeepsie, NY to high-grade gneiss at the Connecticut border east of Pawling, NY. The first appearances of the isograd minerals biotite, garnet, staurolite, kyanite, sillimanite, and orthoclase have been located (Barth, 1936; Vidale, 1974a; Whitney et al., 1996b). Sutter et al. (1985) have identified two different metamorphic events in Dutchess County. The first (their T-2 domain) was characterized by Barrovian metamorphism from chlorite to garnet grade along a gentle metamorphic gradient. The second (T-3 domain) was along a steeper metamorphic gradient resulting in closely spaced isograds from garnet to the first appearance of sillimanite-orthoclase. The data suggest that this higher grade event was a late structural event, and that the zone of maximum intensity of metamorphism corresponds to the area of maximum crustal thickening from imbricate thrusting late in the Taconian orogeny. Several inconsistencies in the simple story of prograde metamorphism, including the fact that fibrolitic

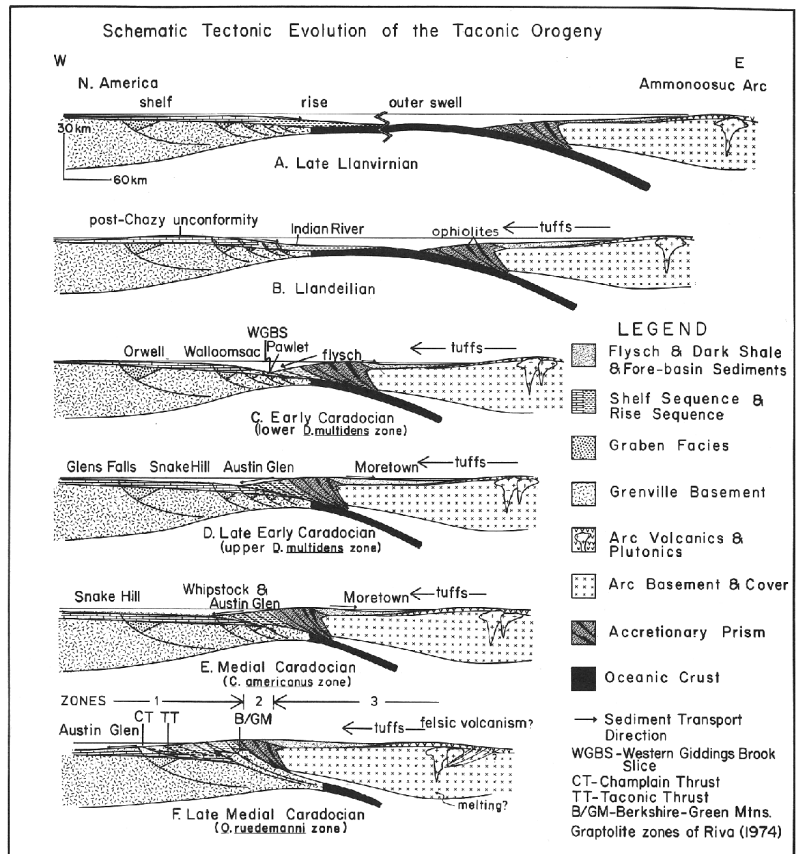


Figure 2. Sequential tectonic cross-sections for the Taconic Orogeny in New England. See text for discussion (from Rowley and Kidd, 1980).

sillimanite occurs at the same localities as the first appearance of kyanite (Figures 16 and 17), as well as the observation that garnets in the greenschist and amphibolite facies rocks (biotite, garnet and staurolite zones) appear to have been affected by circulating fluids (Whitney, 1996; Whitney et al., 1996a and b), may be related to this second, higher grade event. In addition, a range of protolith compositions, often in the same outcrop, demonstrate well the effects of bulk composition on mineral paragenesis.

Metamorphic grade increases to the east-southeast across the county from practically unmetamorphosed shale at the Hudson River at Poughkeepsie, NY to high-grade gneiss at the Connecticut border east of Pawling, NY. The first appearances of the isograd minerals biotite, garnet, staurolite, kyanite, sillimanite, and orthoclase have been located (Barth, 1936; Vidale, 1974a; Whitney et al., 1996b). Exceptions to this simple story of prograde metamorphism include the fact that fibrolitic sillimanite occurs at the same localities as the first appearance of kyanite (Stop 6), and the observation that garnets in the greenschist and amphibolite facies rocks (biotite, garnet and staurolite zones) appear to have been affected by circulating fluids (Whitney, 1996; Whitney et al., 1996a and b). In addition, a range of protolith compositions, often in the same outcrop, demonstrate well the effects of bulk composition on mineral paragenesis.

Where appropriate, each field trip stop description includes the possible mineral reactions as proposed by either Bence and McLelland (1976) or Whitney et al. (1996b). It is an interesting challenge to look for hand specimen evidence of the reactions; in some cases we have included thin section photomicrographs of reactions for which evidence is difficult to see in hand specimen.

Temperature estimates for the metamorphism of Dutchess County rocks have been made by a variety of techniques, and are relatively consistent. Oxygen isotope data (Garlick and Epstein, 1967) suggests a temperature range of 480-625°C for garnet through sillimanite zone metamorphism. Whitney et al. (1996b) used garnet-biotite geothermometry (Ferry and Spear, 1978) to calculate a temperature range of 435-590°C for the same zones, and added calculations from the sillimanite-orthoclase zone of around 730°C. Using the GPBM geobarometer (Ghent and Stout 1981) they also calculated a pressure range of 2-6 kbar for garnet through sillimanite-orthoclase rocks. Their results are plotted in Figure 3 (Whitney et al., 1996b) in which core and rim analyses from garnets at the same grade are plotted with the same symbols, illustrating possible P-T-t paths.

Veins and their relationship to matrix minerals

Quartz and calcite veins are common across Dutchess County. The calcite is presumably due to the dis-

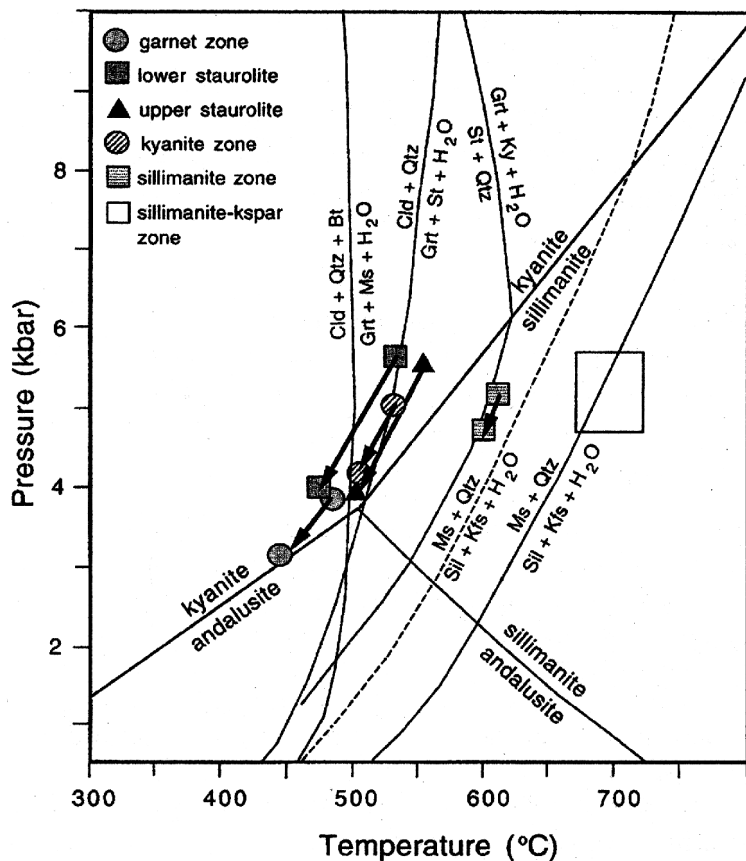


Figure 3. Petrogenetic grid for Dutchess County rocks (from Whitney et al., 1996b).

solution of the many carbonate rocks in the stratigraphic section. The quartz may be derived either from pressure solution during structural deformation, or from silica released to the system by metamorphic reactions especially at the lower grades. Thin sections clearly show evidence for pressure solution during slaty and crenulation cleavage formation. Veins exposed in outcrop are commonly folded, with a crenulation cleavage parallel to their axial surfaces (e.g, Stop 3).

Vidale (1974a) mapped veins throughout the county and found that their mineralogy correlated with increasing grade, with quartz and quartz-calcite veins below the staurolite isograd, quartz-plagioclase veins in the staurolite, kyanite and sillimanite zones, and quartz-plagioclase-orthoclase above the sillimanite-orthoclase isograd. Vidale (1974a) concluded that the different vein-types were related to increasing metamorphic grade, and that the material for the vein was derived from the surrounding matrix. It has since been proposed that some of the veins in the higher-grade rocks could be anatectic melts resulting from high temperatures and abundant aqueous fluids (Bence and McLelland, 1976).

Whitney et al. (1996a) proposed that the fluids which modified the metamorphic sequence migrated through the rocks both along fractures (to create veins) and along channels parallel to the foliation. These migrating fluids modified garnets at or near the peak metamorphic conditions (~525-550°C, ~4-5 kbar), just before the garnets stopped growing. Sr-isotopic studies of fluid inclusions in co-existing quartz and garnet (Whitney et al., 1999) indicate that the fluids in the minerals may have been different, suggesting that fluids trapped in garnets were not in equilibrium with matrix fluids trapped in quartz. In contrast, Whitney and Morrison (1997) determined that garnet and quartz oxygen isotopes were in equilibrium at least at higher grades. These studies indicate that fluid flow has significantly affected the rocks in this sequence and many complexities remain to be worked out.

Fabric development and metamorphic differentiation

Metamorphic fabrics at lower grades typically develop from the growth and rotation of platy minerals (phyllosilicates, graphite, hematite, ilmenite) perpendicular to the maximum shortening direction. The rocks here provide evidence that mass diffusive transfer (pressure solution) is an important mechanism for the development of cleavage domains particularly at lower grades. The classic progression of metamorphic rock fabric, from slate, to phyllite, shist and gneiss, closely correlates with increasing metamorphic grade across the Dutchess County. Crenulation cleavage is common, illustrating multiple deformation phases likely associated with allochthon emplacement (thin skin), later Taconic collisional-related shortening (basement involved), and probable Acadian overprint.

Index minerals, such as biotite, chloritoid, and staurolite have been observed to grow across fabrics, and kyanite appears to be randomly oriented within the foliation plans. This suggests that the metamorphic peak was probably reached after the main deformation phase ended. However, some porphyroblasts, especially garnet and chloritoid do show clear evidence of rotation relative to an earlier foliation (see Whitney et al., 1996; and examples at individual stops below).

Above the sillimanite isograd, schistosity begins to be replaced by gneissic layering, a fabric that then predominates in a zone sub-parallel to the NY-CT border. Vidale (1974b) concluded that this layering was a product of metamorphic differentiation possibly caused by the diffusion of components in response to local pressure gradients related to tectonic stress.

Structure and geologic cross section

The complex tectonic history of the Taconic Orogeny has resulted in a complex structural picture for Dutchess County. A schematic cross section approximately along the line of the field trip is shown in Figure 4. Formation names and symbols are taken from the Lower Hudson sheet of the NYS Geologic Map (Fisher et al., 1970) except as noted below. Important features of the cross section include: remnants of the Taconic Klippe (labeled "Taconic sequence") which underlie topographic highs in the central part of the county; pC basement rocks and their Paleozoic cover raised along eastward-verging reverse or thrust faults (Ratcliffe and Burton, 1990); an overturned syncline involving the Wappinger Group carbonates and the Walloomsac formation in the eastern part of the section (McLelland and Fisher, 1976); and an interesting topographic inversion whereby carbonate rocks of the Wappinger Group are ridge-forming units in the western part of the county, and valley-forming units to the east.

Generalized Geologic Cross Section along NYS 55

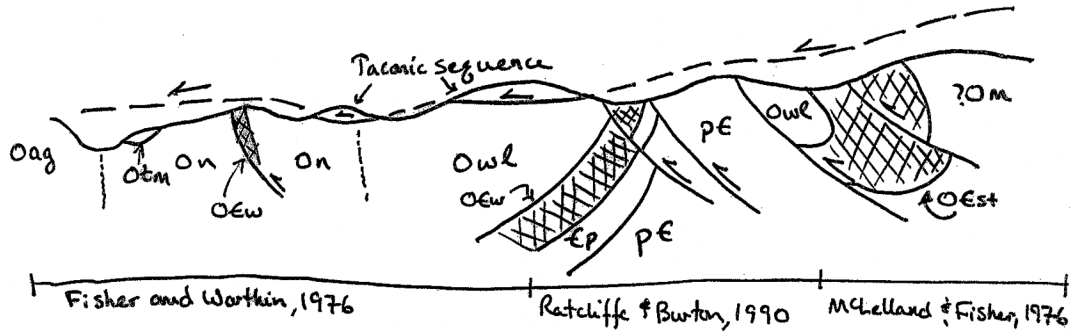


Figure 4. Generalized geologic cross section along NY Route 55 (adopted from Fisher and Warthin, 1976; Ratcliffe and Burton, 1990; and McLelland and Fisher, 1976).

ACKNOWLEDGEMENTS

Donna Whitney; Gordon Baird; Ed Landing; Donald Hodder; and Frederick Vollmer's SUNY New Paltz Petrology classes of 1995, 2003 and 2009.

ROAD LOG

Miles	Route Description
0.0	0.0 Wooster Science Building (loading dock), SUNY New Paltz. Exit lot and turn left (north) onto NY 32 (S. Manheim Blvd.).
0.3	0.3 Turn right (east) at traffic light onto NY 299 (Main Street). Cross over NYS Thruway.
1.2	0.9 Traffic light at intersection with NYS Thruway entrance. Stay on NY 299.
6.8	5.6 Turn right (south) at traffic light onto US 9W.
8.7	1.9 Traffic light at Haviland Road. Continue straight, <i>or turn left for two optional stops.</i>
	(0.7) Small outcrop on left in Austin Glen Formation.

Optional Stop A - Austin Glen Formation (UTM 18T 0587118 4617922). Graywacke beds here show sedimentary structures indicative of deep marine turbidity deposits, including ripple marks, graded beds, and sole marks. Although these sedimentary structures show that the beds are upright, cleavage dips more gently than the upright bedding indicating downward facing structures (Figure 5). This illustrates the complex structural relationships found in this belt due to refolding and probable post-cleavage block rotation in front of the Taconic Allochthon.

(0.3) Continue to parking lot in Johnson Iorio Memorial Park.



Figure 5. Bedding-cleavage relationships in Austin Glen graywacke at Optional Stop A. Ripple marks, sole marks and graded beds (bottom right) demonstrate that these beds are upright. Cleavage (parallel to hammer handle) dips more gently than the bedding, indicating that the cleavage and associated folds are downward facing. Such complex relationships are common in this belt.



Figure 6. (right) Graywacke and shale of the Austin Glen Formation at Optional Stop B, Johnson Iorio Memorial Park. Thickest beds in photograph are approximately 20 cm.

Optional Stop B - Austin Glen Formation at Johnson Iorio Memorial Park (UTM 18T 0587209 4617483). Outcrops of Middle Ordovician Austin Glen Formation thick bedded graywacke and shale show sedimentary structures typical of these deep marine turbidity current deposits that were shed to the west off the rising Taconic Mountains (Figure 6). These deposits were subsequently deformed as the Taconic Allochthon overrode them from the east. Across the Hudson River can be seen outcrops of the Taconic melange, the main Taconic thrust zone (Stop 1), and the Taconic Range to the east. The rock in these cuts is particularly well exposed because of the high percentage of sandstone, more common shale rich exposures may be more representative of a protolith for metapelite outcrops visited at subsequent stops.

From optional stops, return to traffic light at Haviland Road and turn left (south) on US 9W.

- | | | |
|-----|-----|---|
| 8.8 | 0.1 | Take ramp to right following signs for Mid-Hudson Bridge and US 44 - NY 55. |
| 9.5 | 0.7 | Toll booths for Mid-Hudson Bridge. |

Road cuts from here to the Mid-Hudson Bridge show excellent exposures of the Middle Ordovician Austin Glen Formation thick bedded graywacke and shale. The outcrop shows the variation in bed thickness and shale content, as well as numerous faults. See the descriptions above for similar exposures at Optional Stops A and B.

- | | | |
|------|-----|--|
| | | Cross the Mid-Hudson Bridge and keep to the left for the next intersection. |
| 11.2 | 1.7 | Turn left (north) at traffic light onto Jefferson Extension (aka S. Perry Street). |
| 11.2 | 1.7 | Turn left (west) at stop sign onto Union Street, which curves sharply to right (north) and becomes S. Clover Street. |
| 11.5 | 0.3 | Turn left (west) at traffic light on Main Street. Proceed under US 9, and over railroad bridge. |
| 11.7 | 0.2 | Turn left at traffic light onto Rinaldi Blvd. |
| 11.8 | 0.1 | Turn right onto Gerald Drive, following sign for Truck Route, proceeding through tunnel under Mid-Hudson Bridge. |
| 12.0 | 0.2 | Turn right onto Hendryk Drive into Kaal Park. |
| 12.1 | 0.1 | Park in lot at Kaal Park (City of Poughkeepsie Park). |

Stop 1 - Taconic Melange at Kaal Park (UTM 18T 0588112 4617257). Taconic melange at Kaal Park is the type locality of the Poughkeepsie Melange of Fisher. The Taconic melange is well exposed here, although large sections

are covered with poison ivy. The phacoidal, or scaly, cleavage is typical of melange fabric world-wide. The shale breaks into centimeter-scale lens-like phacoids whose surfaces are polished and striated. Blocks within the melange are mainly graywacke here, and probably represent a disrupted bedded sequence similar to exposures of Austin Glen on the west side of the river at Optional Stop A. Locally, however, limestone blocks occur within the melange suggesting a sedimentary olistostromal component. The Taconic melange is a thick (kilometer-scale) tectonostratigraphic unit that occurs along the western margin of the Taconic Allochthons in the northern Appalachians, and represents a complex syn-sedimentary fault zone associated with allochthon emplacement. To the north, the Taconic melange is unconformably overlain by the Silurian Helderberg Group (Vollmer, 1981), establishing its Taconian age.



Figure 7. Taconic melange at Stop 1, Kaal Park showing graywacke blocks in phacoidally cleaved shale matrix.

Many exposures of "Normanskill" or "Snake Hill" shale in the Poughkeepsie area show clasts suggestive of melange, and the Taconic melange is suggested to be a potential protolith for some of the metapelites (such as Walloomsac) examined at subsequent stops.

- | | | |
|------|-----|---|
| | | Retrace route to US 44 - NY 55, returning on Hendryk and Gerald Drives and Rinaldi Blvd. |
| 12.6 | 0.5 | Turn right (east) at traffic light from Rinaldi Blvd. onto Main Street. Turn right (south) at light onto N. Perry, and left (east) at traffic light on US 44 - NY 55. |
| 14.7 | 2.1 | Traffic light at Raymond Avenue (turn right for Vassar College). Continue straight on US 44 - NY 55. Pass outcrops of Taconic Melange before Raymond Avenue. |
| 14.9 | 0.2 | Bear right onto NY 55, leaving US 44. |
| 16.2 | 1.3 | Outcrop of Wappinger Group carbonates on right. <i>For optional stop C, park on Henmond Road to the right just past the outcrop and return on foot.</i> |

Optional Stop C - Wappinger Group Carbonates (UTM 18T 0593529 461 5609). Outcrop of Wappinger Group carbonates, possible protolith for metacarbonates to be examined at Stop 9.

- | | | |
|------|-----|---|
| 16.7 | 0.5 | Cross Wappinger Creek. |
| 17.1 | 0.4 | Bear right at traffic light onto DCR 21 (Noxon Road). |
| 19.8 | 2.7 | Cross DCR 49 (Titusville Road) at traffic light. |
| 22.2 | 2.4 | Cross Sprout Creek. |
| 22.4 | 0.2 | Pull off on left near large outcrops on both sides of road. |

Stop 2 - Chlorite Zone at Noxon Road (UTM 18T 0594925 4615132). These outcrops are within the Taconic Allochthon, and show considerable structural complexity. The rock here is chlorite grade, and exhibits slaty cleavage, abundant chevron folds, and crenulation cleavage that folds an earlier foliation subparallel to bedding. Thin sections show that cleavage formation is dominated by pressure solution, a likely source of quartz veins (Figure 8 and 9). Witney et al. (1996) suggest that the chlorite forms from clay minerals. These reactions release silica that may also contribute SiO₂ to the formation of quartz veins.

Four distinct lithologic units are present within these outcrops: graywacke and dark shale, green slates, red slates, and slate with interbedded micrite. The contacts do not trace directly across the road, and are either part of steeply plunging structures or are discontinuous. Fisher and Warthin (1976) and Bence and McLelland (1976) assign the graywacke unit to Mount Merino (at the far west end), and the other three to the Indian River Formation. Structur-

ally this outcrop lies within the Giddings Brook Slice, the the structurally lowest slice of the Taconic Allochthon (Stanley and Ratcliffe, 1985).

On the south side of the road, west of the main outcrops, are dark crenulated slates and folded graywacke beds. Bottom structures exposed in a meter-scale synform indicate the folds are upward facing. This unit is in contact with green slates further east on the north side of the outcrop, where quartz veining indicate faulting along the contact. It seems likely that this unit is equivalent to the Austin Glen/Pawlet, and represents flysh deposited on the continental rise and subsequently carried with the allochthon (see main text).

The bulk of the main outcrop consists of highly folded and crenulated green slate with some interbedded chert and siltstone, however on the south side of the road is a large section of red slates of typical Indian River aspect. Bence and McLelland (1976) report hematite is present in the red slates, and absent in the green slates. The presence of hematite in Indian River has been attributed to both primary and secondary origins (e.g., Bird and Dewey, 1970; Bence and McLelland, 1976), and may represent erosion of a weathered horizon exposed as a foreland bulge during the Taconic Orogeny. If the graywacke unit is Austin Glen/Pawlet, then the Mount Merino may be represented by the green slates with cherty interlayers.

At the eastern end of the outcrop on the north side, a unit of banded gray micrites and interlayered slate is in contact with the green slate. The contact between the two units is complex, and disruption of the micrite beds is suggestive of soft sediment deformation with a strong tectonic overprint. This unit is remarkably similar to thinly bedded brecciated micrites at Schodack Landing, also within the Giddings Brook slice, which have been interpreted as slump deposits on the continental slope (Bird and Dewey, 1975; Friedman, 1979). This unit likely has a similar origin.

As we drive east from this stop we will leave the Giddings Brook slice, and shortly enter the structurally higher Everett slice. The remaining metapelite stops, except possibly Stop 5, are below the Everett slice (Stanley and Ratcliffe, 1985).

Pull out WATCHING CAREFULLY FOR TRAFFIC and continue on DCR 21 (Noxon Road).

- 23.1 0.7 Turn left at intersection, following DCR 21 (Noxon Road).
- 23.3 0.2 Cross Jackson Creek
- 23.6 0.3 Cross underpass below Taconic Parkway.

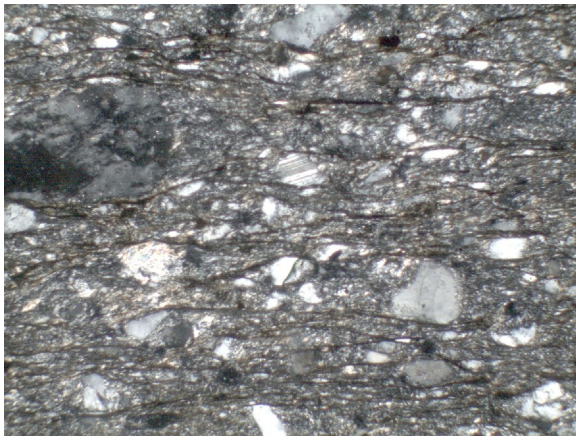


Figure 8. Thin section of graywacke (Austin Glen/Pawlet?) from Stop 2. Grains include abundant quartz, chert, lithic fragments, and feldspar (note plagioclase grain at center). A pressure solution cleavage (horizontal) truncates quartz grains. Field width 1.46 mm, CPL.

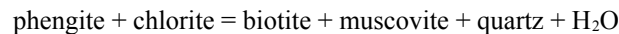


Figure 9. Thin section of kink fold and crenulation cleavage from Stop 2. The early foliation is bedding parallel and is folded by the later crenulation cleavage (vertical). The high phyllosilicate density in the crenulation cleavage domains suggests preferential dissolution of quartz. Field width 3.67 mm, PPL.

- 25.2 1.6 Cross DCR 82 at traffic light. Continue on DCR 21 (Noxon Road).
 26.8 1.6 Cross NY 55 at traffic light. Continue on DCR 21 (now Bruzgul Road).
 28.5 1.7 After following outcrops down hill on left, turn left on Perkins Lane (dirt) and pull off on right.

Stop 3 - Biotite Zone at Perkins Lane (18T 0608117 4612444). Black slates in biotite zone, although biotite is not visible in hand sample. These rocks are assigned to the Walloomsac Formation and are believed to be deposited on Wappinger Group shelf carbonates, which underlie Clove Valley just to the east. Thus, tectonically, these are exposed in a window and represent flysch overridden by the allochthon, or are brought up with basement by late reverse faults.

The main cleavage in the outcrop is relatively low angle, suggestive of high shear strains associated with allochthon emplacement. Numerous folded veins are related to shortening across the cleavage (Figure 10). Whitney, et al. (1996b) propose that biotite could form by the reaction:



An AFM projection for the upper biotite zone shows that at this grade biotite is stable for a wide range of pelitic compositions (Figure 11A).



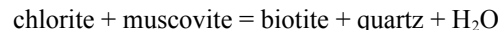
Figure 10. Folded veins with gently dipping axial planes parallel to crenulation cleavage at Stop 3.

- 28.8 0.3 *Lunch stop.* Turn right into Tymor Park (Town of Unionvale).

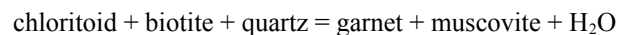
Lunch Stop - Tymor Park. South of the picnic area a dirt road runs through a spectacular gorge in the Wappingers Group dolostones to Furnace Pond. The pond was named either for a lime mill that stood on its shore (on Lime Mill Road), or for the Clove Springs Ironworks whose furnaces processed up to 700 tons of iron ore from Union Vale mines during the 19th century. Remains of the ironworks can be seen south of the park on Furnace Road.

- 29.7 0.9 Small traffic circle at intersection with DCR 9. Continue straight on DCR 21 (now Wingdale Mountain Road).
 30.4 0.7 Pull off on right, opposite gated dirt road.

Stop 4 - Garnet Zone at Wingdale Mountain Road (UTM 18T 0610910 4612446). Garnet zone with chloritoid. At this point we have crossed the garnet isograd, and garnets are now present in hand sample, although not always easy to find. Downhill from the parking area are biotite-bearing dark schists and metagraywackes (Figure 12). Garnet is found in the more fine-grained beds, which are presumably more aluminous than the graywackes. Whitney et al. (1996b) propose the reaction:



Across from the pull out, near a gated entrance, are chloritoid-bearing schists (Figures 13 and 14), suggesting the reaction:



Bence and McLelland (1976) attribute the mineralogical difference between these two outcrops to a more aluminous protolith, and the separation of the two bulk compositions by the garnet-chlorite tie line in an AKFM projection (Figure 11B). Note that the tie line switch in the biotite (Stop 3) to garnet (Stop 4) zone transition (Figures 11A and 11B) allows garnet to occur in typical pelitic compositions.

In thin sections from the lower outcrop, biotite is found growing across a crenulation cleavage, suggesting post-tectonic growth (Figure 12). Thin sections from the chloritoid-bearing outcrop show clear rotation of chloritoid and garnet porphyroblasts associated with shear strain (Figures 13 and 14). These porphyroblasts have overgrown an early foliation which has then been rotated during continued or later deformation.

Continue on DCR 21 (Wingdale Mountain Road).

32.9 2.5 *Alternate stop location with better parking than Stop 5 without requiring permission.* Pull out is on left at hair pin turn. It is suggested that cars turn around below and return to park on right, facing up hill. Outcrops are across from pull out and up hill on right.

Alternate Stop 5 - Stauroilite Zone (UTM 18T 0614569 4611725). See Whitney and Peck (2004) for a description of this outcrop.

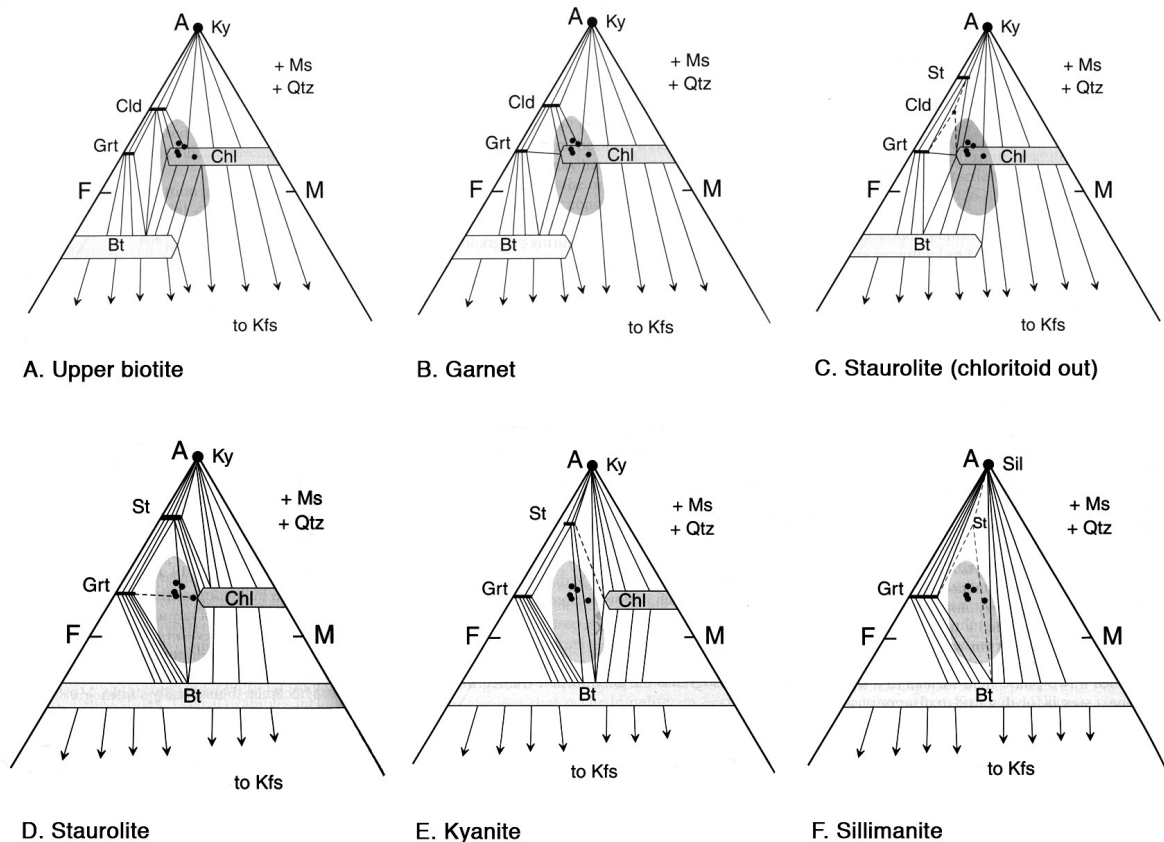
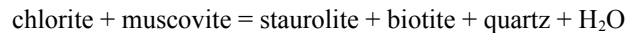


Figure 11. AKFM projections at several grades of metamorphism. The shaded area represents the composition of typical pelites (from Winter, 2001).

33.8 0.9 Pull off on right past Beebe Hill Road being VERY CAREFUL OF NARROW SHOULDER. Please do not park on Beebe Hill Road which is private and special permission must be obtained. You may wish to visit the alternate stop above for better parking.

Stop 5 - Staurolite Zone at Beebe Hill Road (UTM 18T 0615310 4611610). Coarse biotite-muscovite schists with garnet and staurolite porphyroblasts. Rocks here are now well within the staurolite zone, and contain centimeter scale garnet and staurolite porphyroblasts in the muscovite schist. This is the upper staurolite zone of Whitney, et al. (1996b), and because there is no chlorite in the rocks, the likely reaction is:



A tie line switch from garnet-chlorite to staurolite-biotite occurs at this grade, making staurolite stable over a wide range of pelitic compositions (Figure 11D).

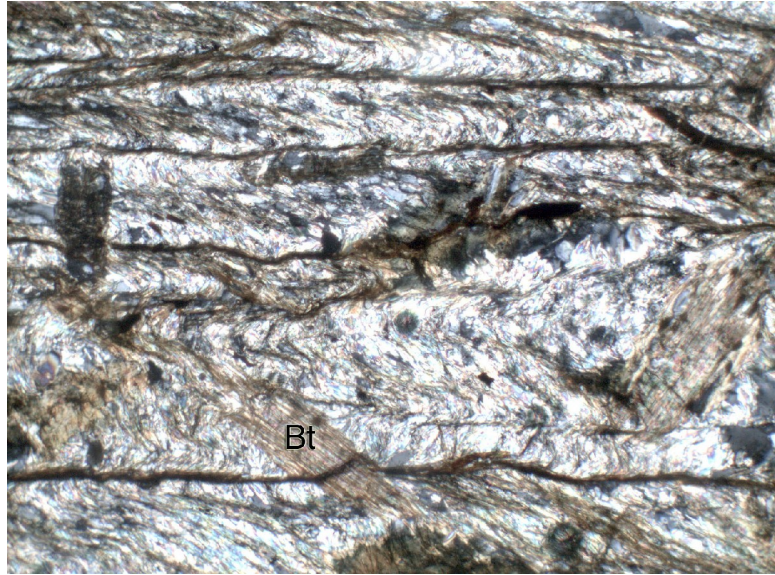


Figure 12. Thin section of non-chloritoid bearing schist from Stop 4 showing biotite overgrowing a crenulation cleavage. The earlier foliation (vertical) is folded by a crenulation cleavage (horizontal), and cross cut by the biotite, suggestion post-tectonic growth of the index mineral. Field width 1.46 mm, PPL.



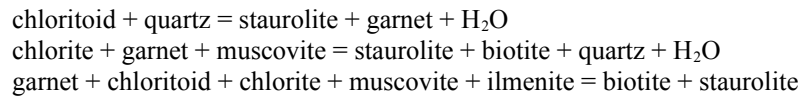
Figure 13. Thin section of chloritoid (center left) garnet schist from Stop 4. Top to right shear is shown by higher phyllosilicate density at top left and bottom right of garnet, and by bent ilmenite grains (opaque). Field width 3.67 mm, PPL.



Figure 14. Thin section showing rotated chloritoid in schist from Stop 4. Top to right shear is shown by: high phyllosilicate density at top left and bottom right, corresponding low density at top right and bottom left, rotation of an early overgrown foliation, and by bent ilmenite grain (top left). Field width 3.67 mm, CPL.

- Turn around CAREFULLY and return on DCR 21 (Wingdale Mountain Road).
- 35.4 1.6 Turn left on Pleasant Ridge Extension.
- 35.9 0.5 Turn left at stop sign onto Pleasant Ridge Road.
- 36.3 0.4 Intersection with Still Road. *Turn right onto Still Road for optional stop.*
- (0.7) Follow Still Road and turn around at Mennella Road.
- (0.1) Park on right across from small low outcrop of schists on left.

Optional Stop D - Lower Staurolite Zone at Still Road (UTM 18T 0611053 4611579). Please DO NOT HAMMER on outcrop. Chloritoid is partially replaced by staurolite, and propyroblasts are difficult to see in hand sample. As noted by Bence and McLelland (1976), five coexisting ferromagnesian phases, garnet-chloritoid-staurolite-biotite-chlorite, can be identified in thin section, a disequilibrium assemblage violating the mineralogical phase rule (Figures 11C and 15). Three different reactions are possible here (Whitney et al., 1996b):



Careful study is needed to determine which reaction(s) are most likely because garnet may be either a product and a reactant. Figure 11C shows the chloritoid-out reaction, and the phase stability triangles for the five phases.

- (0.7) Turn right at stop sign back onto Pleasant Ridge Road.
- 39.5 3.2 Intersection of Pleasant Ridge Road and NY 55.

Optional Stop E - Poughquag Quartzite (UTM 18T 0610711 4606921). The Poughquag Quartzite is the basal unit overlying Precambrian gneisses here, and is correlative with the Potsdam sandstone and similar Cambrian basal units. On NY 55 to the left (east) the unconformity of the Poughquag Quartzite with underlying Precambrian gneisses is crossed proceeding uphill. These basement rocks were carried up on reverse faults, presumably later in the orogeny, and along steeper faults, when the continental basement eventually became involved in the collisional event.

- Turn left on NY 55.
- 40.9 1.4 Cross Appalachian Trail.
- 41.2 0.3 Pass intersection with Old Route 55 on left.
- 42.4 1.2 Pull over on right shoulder near large outcrops on both sides of the road.

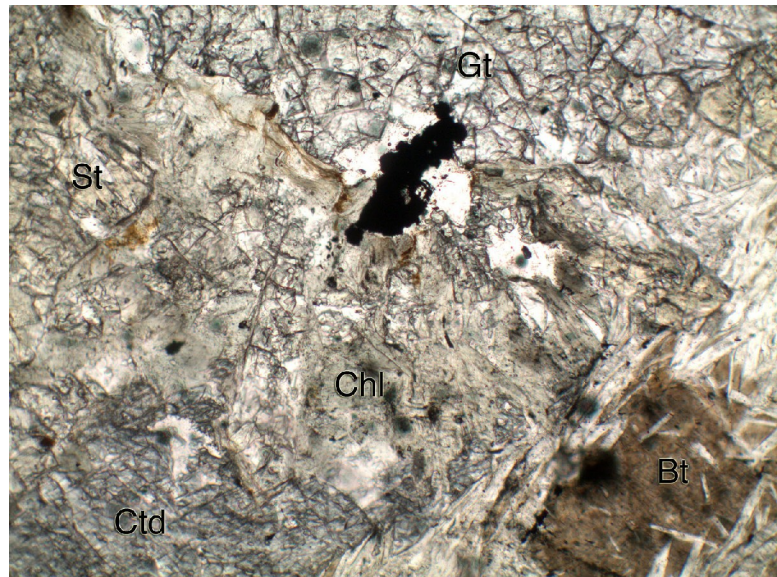
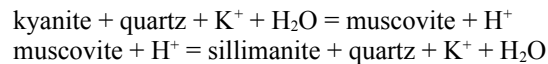


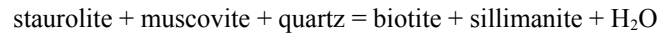
Figure 15. Thin section showing the five coexisting phases garnet-chloritoid-staurolite-biotite-chlorite at Optional Stop D. Coexistence of the five phases is an apparent violation of the phase rule (Figure 11C). Field width 1.46 mm, PPL.

Stop 6 - Kyanite Zone (UTM 18T 0613449 4603217). Kyanite is present as small 1-2 centimeter blue blades, but is difficult to find. It appears on weathered surfaces as white sprays or rosettes on foliation planes (Figure 16), apparently with no preferred linear alignment. **DO NOT HAMMER WEATHERED SURFACES**, samples may easily be obtained from fresh surfaces on existing road outcrops.

In thin section the assemblage kyanite-sillimanite-staurolite-garnet has been observed (Figures 17 and 18). The kyanite appears resorbed (Figure 17), and the presence of sillimanite as fibrolite (Figure 18) in the rocks suggests a captured divariant reaction, or disequilibrium (see Figures 3, 11E and 11F). Fibrolite has been observed in close proximity (millimeters) to the kyanite. One possibility is that linked reactions are occurring in adjacent local domains separated by a chemical potential gradient as proposed by Carmichael (see Blatt et al., 2006):



This allows the reaction to proceed without direct contact of the polymorphs. However, in the sample described here, this requires chemical potential gradients at a millimeter scale. Another possible reaction to explain the presence of fibrolite (Whitney et al., 1996b) is:



Again, this seems to require a strong chemical potential gradient and extremely limited fluid mobility.



Figure 16. Kyanite sprays in outcrop at Stop 6.

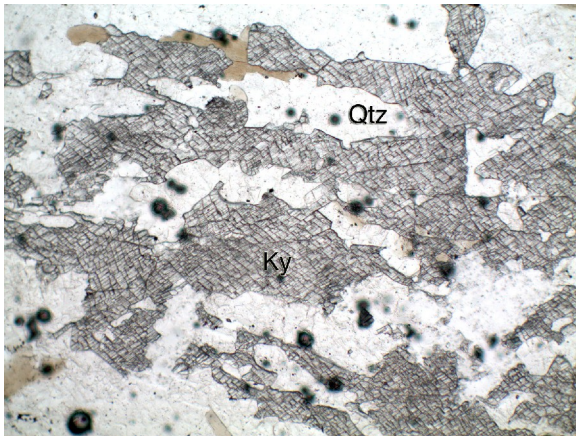


Figure 17. Thin section showing kyanite from Stop 6. In hand sample kyanite occurs as elongate needles, but shows irregular outlines in thin section suggesting it may be resorbed. Field width 1.46 mm, PPL.

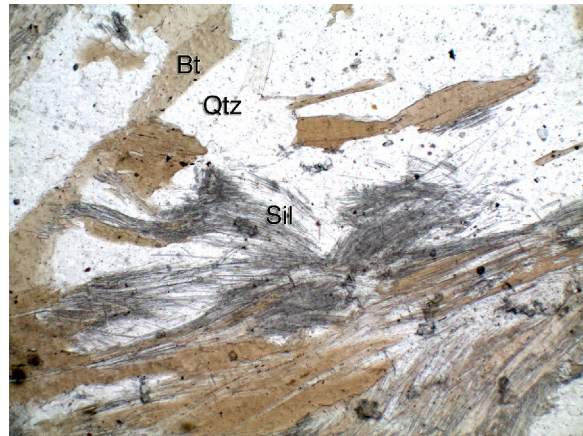


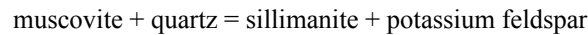
Figure 18. In the same thin section as Figure 17, within millimeters, is abundant sillimanite occurring as fibrolite. The occurrence of both implies a captured divariant reaction or disequilibrium. Field width 1.46 mm, PPL.

Whitney et al. (1996) noted that fibrolite first occurs in the kyanite zone. They discuss this in detail, and present several hypotheses to explain this. Variation in Mg content between layers and a reaction involving staurolite is ruled out by them, as they did not find evidence of staurolite in fibrolitic sillimanite-bearing layers. They suggest instead that the fibrolite may have grown at a later time (later Taconic or possibly Acadian), and that the reaction may be related to fluid infiltration, with fibrolite growth controlled by differences in local permeability and possible metasomatic Al-enrichment.

As discussed in the main text, two metamorphic events have been identified in Dutchess County (Sutter et al., 1985), and may be related to these perplexing apparent disequilibrium assemblages.

- | | | |
|------|-----|---|
| 45.1 | 2.7 | Pass outcrops of tremolite-diopside marble, which we will return to at Stop 9. |
| 46.1 | 1.0 | Cross over NY 22, taking ramp to follow NY 22 N. |
| 46.7 | 0.6 | Turn right at light onto Quaker Hill Road. |
| 49.3 | 2.6 | Park on right using CAUTION DUE TO NARROW SHOULDER. Hazard blinkers and flag person recommended. Alternate parking is just below hairpin curve at bottom of hill (UTM 18T 0619799 4602388). |

Stop 7 - Sillimanite Zone at Quaker Hill Road (UTM 18T 0620040 4602299). Prismatic sillimanite crystals are evident in thin section (Figure 19), but not visible in outcrop. Small pegmatite dikes containing tourmaline-muscovite-potassium feldspar-quartz intrude the sillimanite-garnet-muscovite schists, and provide evidence of anatectic melting (as opposed to quartz veins formed by lower temperature pressure solution dissolution and precipitation). In thin section (Figure 19) there is evidence for the reaction:



Note that the sillimanite between the muscovite and potassium feldspar has a coarser prismatic form than that between the quartz and potassium feldspar. A possible explanation is that ion mobility was greater along the muscovite-potassium feldspar interface than the quartz-potassium feldspar interface due to easier fluid migration, so growth was favored over nucleation.

- | | | |
|------|-----|---|
| | | Continue uphill on Quaker Hill Road. |
| 50.0 | 0.7 | Turn right at stop sign on DCR 66 (Old Quaker Hill Road). |
| 50.7 | 0.7 | Turn left on Burgess Road (dirt), park on right. |

Stop 8 - Sillimanite-Potassium Feldspar Zone at Burgess Road (UTM 18T 0621121 4600990). Manhattan schist here displays more pronounced gneissic banding, and tourmaline-muscovite-potassium feldspar-quartz pegmatite dikes. Whitney and Peck (2004) assign this outcrop to the Sillimanite-Potassium Feldspar zone.

- | | | |
|------|-----|---|
| | | Turn around and return on DCR 66 (Old Quaker Hill Road) |
| 51.6 | 0.9 | Turn left on Quaker Hill Road. |

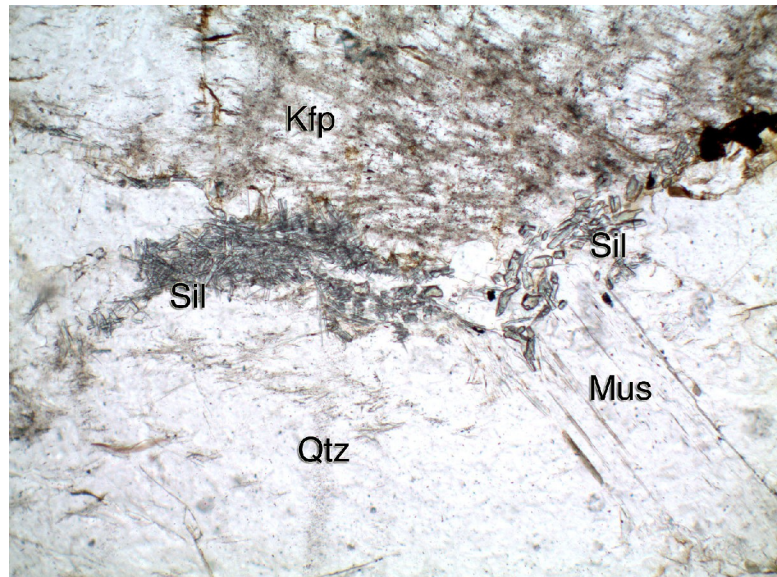


Figure 19. Thin section from Stop 7 showing prismatic sillimanite (as opposed to fibrolite) and reaction textures. Note that the sillimanite between the muscovite and potassium feldspar is a coarser prismatic form than that between the quartz and potassium feldspar. Field width 1.46 mm, PPL.



Figure 20. Diopside tablets at Stop 9.



Figure 21. Diopside sprays at Stop 9.

- | | | |
|------|-----|--|
| 54.8 | 3.2 | Turn left at traffic light on NY 22. |
| 54.9 | 0.1 | Bear right onto NY 55 toward Poughkeepsie. |
| 55.9 | 1.0 | Pull over on right shoulder near large outcrops. |

Stop 9 - Metacarbonates of the Briarcliff Dolostone (UTM 18T 0615950 4601071). Outcrops of diopside-trochilite-phlogopite marble that represent the metamorphism of carbonates to metamorphic grades approaching the sillimanite zone. One layer in particular contains a high concentration of diopside tablets and sprays (Figures 9 and 10). The diopsidic composition of these unusual forms was confirmed by thin section and X-ray diffraction analysis. Remnants of the less ductile silica-bearing beds within the marble record extreme ductile folding and boudinage, testifying to the high temperature deformation they enjoyed.

END OF FIELD TRIP. To return to New Paltz take NY 55 west, through Poughkeepsie to the Mid-Hudson Bridge. Once over the bridge follow US 9W north to NY 299. Take NY 299 west to New Paltz and NY 32 (or the NYS Thruway). Follow NY 32 south to the SUNY New Paltz Campus.

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The Shawangunk and Martinsburg Formations Revisited: Sedimentology, Stratigraphy, Mineralogy, Geochemistry, Structure and Paleontology

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INTRODUCTION

In southeastern New York Middle Silurian Shawangunk Formation (Figure 1), containing gray conglomerate, sandstone and shale, lies unconformably above the Ordovician Martinsburg Formation, consisting of shales and graywackes. In southwestern New York, near the Port Jervis area, The Shawangunk Formation is overlain by the Bloomsburg Red Beds, the same stratigraphic sequence that occurs in Pennsylvania and New Jersey to the southwest. The Shawangunk Formation thins gradually from Port Jervis to its pinchout near Hidden Valley and Binnewater, New York. Two tongues of the upper part of the Shawangunk are: the Ellenville Tongue that extends from the Ellenville-Accord area to its feather edge just southwest of the New York-New Jersey border and, the High View Tongue that is restricted to the Wurtsboro area (Epstein and Lyttle, 1987; Epstein, 1993).

Early in the Paleozoic carbonate banks lay along the east coast of the ancient North American continent. During the Ordovician plate convergence commenced in the closing of the Iapetus Sea and a deep basin developed into which thick muds and dirty sands were deposited. These were later lithified into the Martinsburg Formation. Eventually, with continued compression, these sediments were folded and faulted during the complex deformation of the Taconic Orogeny. The trend of these folds in southeastern New York is approximately N20E. As one proceeds westward across the Wallkill Valley these structures become less intense. Subsequent to the Taconic Orogeny mountains rose to the east and coarse sediments were transported westward and deposited as the conglomerates and sandstones of the Shawangunk Formation across the beveled folds of the Martinsburg. Deposition occurred on a plain of alluviation and in a marine basin to the northwest. Erosion of the source area was intense, and the climate, based on the mineralogy of the rocks, was warm and at least semiarid. The source was composed predominately of sedimentary and low-grade metamorphic rocks with exceptionally abundant quartz veins and small local areas of gneiss and granite. As the source highlands were eroded, the steep braided streams of the Shawangunk gave way to more gentle-gradient streams of the Bloomsburg Red Beds.

The Taconic Orogeny

During the Ordovician Period there existed an ocean, known as the Iapetus or proto-Atlantic, to the east of the Shawangunk Mountains. Bisecting this body of water was a narrow, mountainous landmass called the Taconic Island Arc, similar in shape to today's Caribbean or Aleutian islands. Landmasses that bordered the Iapetus Ocean began to move toward one another. On the east the landmass was present-day Western Europe composed largely of granitic rocks, whereas on the west it was present-day North America. The movement of these landmasses, or tectonic plates, resulted in the shrinking of the Iapetus Ocean and eventual collision of the Taconic Island Arc with eastern proto-North America, resulting in the formation of large mountains as high as the present-day Himalayas. The collision resulted in the (new) Taconic Mountains riding up over the edge of the proto-North American continent and pushing it down to form a basin. The sediments shed by these mountains were deposited in this basin that was more than 500m (1,500 ft) deep. This process, known as the Taconic Orogeny, or mountain-building episode, occurred during the Middle Ordovician Period about 450 million years ago. The Taconian event was the result of shelf, slope, and island arc accretion. It was during this time period that the Martinsburg Formation was deposited.

The Acadian Orogeny

During the Devonian Period (about 418-362 Million years ago) the Iapetus Ocean became much narrower and shallower, finally becoming a shallow basin. Landmasses representing North America and Western Europe began to move toward one another. This collision, known as the Acadian Orogeny, took some tens of millions of years to occur and resulted in the formation of large mountains, the roots of which are actually what we now call the Berkshires in Massachusetts. The Avalonian Terrane was accreted during the Acadian Orogeny. Eventually, these mountains were weathered and eroded. The sediment that they shed accumulated to the west forming the Catskill Delta during the middle of the Devonian Period. Some workers estimate that the volume of sediments that was dumped into this basin approached 70,000 cubic miles. The collision of two large landmasses known as Laurasia and Gondwana began after the Devonian and lasted through the Permian Period. This collision, locally called the Alleghany Orogeny, overprinted much of the deformation produced during the Acadian and Taconic orogenies and produced the major structures of the central and southern Appalachians. With Laurasia and Gondwana now sutured together, the supercontinent Pangaea had been formed. This was the last major orogenic event to affect the present-day east coast of North America. The relationship between the overlying Shawangunk and the underlying Martinsburg formations is an angular unconformity. That is, there is a significant gap in the rock record that resulted from a change that caused deposition of the Martinsburg to cease for a considerable amount of time during which there was uplift and erosion with loss of the previously formed record. In other words, although the Shawangunk overlies the Martinsburg, it is not in stratigraphic succession; there is a hiatus of between 10-30 million years between the formation of the Martinsburg and the deposition of the Shawangunk.

STRATIGRAPHY

The following is a list of formations that occur in the field trip area from the mid-Hudson Valley to Port Jervis, listed from youngest to oldest. The last carbonate unit is the Onondaga Limestone, above which occur the clastics of the Middle Devonian Hamilton Group.

Plattekill Formation of Fletcher (1962)
(Middle Devonian): red and gray shale, siltstone and sandstone; 500+ feet thick.

Ashokan Formation (Middle Devonian): thin- to thick-bedded olive gray sandstone with minor siltstone and shale; 500-700 feet thick.

Mount Marion Formation (Middle Devonian): olive-gray to dark-gray, platy, very fine to medium grained sandstone, siltstone and shale; >1000 feet thick.

Bakoven Shale (Middle Devonian): dark gray shale; 200-300 feet thick.

Onondaga Limestone (Middle Devonian): fossiliferous limestone with the following members • Edgecliff at base (light-weathering chert), Nedrow above (dark-weathering chert; shaly) and Moorehouse at top (dark-weathering chert). The Seneca Member, missing in the Hudson Valley, can be observed at Cherry Valley; 100 feet thick.



Figure 1. Mohonk Lake lies in a faulted, glacially scoured basin bounded by the light, quartz pebble Shawangunk Conglomerate. Beyond the Shawangunk lies the Port Jervis trough underlain by glacial sediments and occasional outcrops of Onondaga Limestone (see Feldman, 1985). Note the Catskill Mountains in the distance. The valleys in the Catskills are aligned along linears that are presumably controlled by structural weakness.

- Schoharie Formation** (Lower Devonian): thin- to medium-bedded, calcareous mudstone and limestone, more calcareous toward top. From top to bottom - Saugerties, Aquetuck, and Carlisle Center members; 180-215 feet thick.
- Esopus Formation** (Lower Devonian): dark, laminated and massive non-calcareous, siliceous argillaceous siltstone and silty shale; 200 feet thick (thickens to southwest).
- Glenerie Formation** (Lower Devonian): thin- to medium-bedded siliceous limestone, chert and shale; 50-80 feet thick.
- Connelly Conglomerate** (Lower Devonian): dark, thin- to thick-bedded pebble conglomerate, quartz arenite, shale and chert; 0-20 feet thick.
- Port Ewen Formation** (Lower Devonian): dark, fine- to medium grained, sparsely fossiliferous, calcareous, partly cherty, irregularly bedded mudstone and limestone; 0-175(?) feet thick (180 feet thick near Port Jervis).
- Alsen Formation** (Lower Devonian): fine- to coarse-grained, irregularly bedded, thin- to medium-bedded, argillaceous to partly cherty limestone; 20 feet thick.
- Becraft Limestone** (Lower Devonian): massive, very light to dark gray and pink, coarse-grained crinoidal limestone with thin-bedded limestone and shaly partings near the bottom in places; 30-50 feet thick (thins toward High Falls; 3 feet thick near Port Jervis).
- New Scotland Formation** (Lower Devonian): calcareous mudstone and silty, fine- to medium grained, thin- to medium-bedded limestone; may contain some chert; 100 feet thick.
- Kalkberg Limestone** (Lower Devonian): thin- to medium-bedded, moderately irregularly bedded limestone, finer grained than the Coeymans Formation below, with abundant beds and nodules of chert and interbedded calcareous and argillaceous shales; 70 feet thick.
- Ravena Limestone Member of the Coeymans Formation** (Lower Devonian): wavy bedded, fine- to medium-grained and occasionally coarse grained limestone with abundant thin shaly partings; 15-20 feet thick.
- Thacher Member of the Manlius Limestone** (Lower Devonian): laminated to thin-bedded, fine-grained, cross-laminated, graded, microchanneled, mudcracked, locally biostromal limestone with shale partings; 40-55 feet thick.
- Rondout Formation** (Lower Devonian and Upper Silurian): Fossiliferous, fine- to coarse-grained, thin- to thick-bedded limestone and barren laminated, argillaceous dolomite. Limestone lentils come and go but the more persistent ones have been named (from top to bottom) - Whiteport Dolomite, Glasco Limestone and Rosendale members; 10-50 feet thick.
- Binnewater Sandstone** (Upper Silurian): fine-grained, thin- to thick-bedded, cross-bedded and planar cross-bedded, rippled quartz arenite, with gray shale and shaly carbonate. Probably grades southwestward into the Poxono Island Formation; 0-35 feet thick.
- Poxono Island Formation (Upper Silurian):** poorly exposed gray and greenish dolomite and shale, possibly with red shales in the lower part; 0-500 feet thick.
- High Falls Shale** (Upper Silurian): red and green, laminated to massive, calcareous shale and siltstone, occasional thin argillaceous limestone and dolostone; ripple marks, dessication cracks; 0-80 feet thick.
- Bloomsburg Red Beds** (Upper Silurian): grayish-red and gray shale, siltstone and sandstone; 0-700 feet thick.
- Tongue of the Bloomsburg Red Beds:** grayish-red siltstone and shale and slightly conglomeratic, partly cross-bedded sandstone with pebbles of milky quartz, jasper, and rock fragments, and gray sandstone; 0-300 feet thick.
- Shawangunk Formation** (Middle Silurian): cross-bedded and planar-bedded, channeled, quartz-pebble conglomerate (rose quartz conspicuous in upper part), quartzite, minor gray shale and siltstone and lesser red to green shale. Lower contact unconformable; 0-1,400 feet thick.
- Tongue of the Shawangunk Formation:** cross-bedded, cross-laminated (distinctive very light and medium-dark-gray laminae), planar-bedded, thin- to thick-bedded, medium-grained quartzite and conglomerate with quartz pebbles as much as 2 inches long and greenish-gray silty shale and siltstone; 0-350 feet thick.
- Diamictite** (Lower Silurian or Upper Ordovician): diamictite (colluviums and shale-chip gravel with exotic pebbles and fault gauge of sheared clay and quartz veins; lower contact unconformable; less than 1 foot thick.
- Martinsburg Formation** (Upper and Middle Ordovician): greater than 10,000 feet thick.
- Shale and Graywacke at Mamakating:** dominantly thick sequences of thin- to medium-bedded, medium dark gray shale interbedded with very thin to thick-bedded graywacke (as much as 6 feet thick) alternating with thinner

sequences of medium-bedded gray-wacke interbedded with less thin- to medium-bedded shale. Grades downward and laterally into the sandstone at Pine Bush.

Sandstone at Pine Bush: medium-grained, medium- to thick-bedded, medium gray, speckled light-olive-gray- and light-olive-brown-weathering quartzitic sandstone interbedded with, and containing rip-up clasts of thin- to medium-bedded, medium-dark-gray, greenish gray-weathering shale and fine-grained siltstone. Lower contact with Bushkill Member is interpreted to be conformable, but in many places it is marked by a thrust fault; grades upward and laterally in the shale and graywacke Mamakating.

Bushkill Member: laminated to thin-bedded shale and slate containing fine-grained graywacke siltstone; bed thickness of shales does not exceed 2 inches and bed thickness of graywackes rarely exceeds 12 inches; lower contact conformable with underlying Balmville Limestone of Holzwasser (1926), but often disrupted by thrust faulting.



Figure 2. The Shawangunk Conglomerate, a quartz pebble conglomerate that is quartzitic in places, cross-bedded and planar-bedded, with minor gray shale and siltstone.

ROAD LOG AND STOP DESCRIPTIONS

We will be traveling through tick-infested areas; take the proper precautions and be sure to use DEET, especially on da FEET!

Total Miles	Miles Between Stops	
0.0	0.0	Leave parking lot at SUNY New Paltz. This locality is on blacktop of Holocene age, conformably overlying fill of unknown age brought in by contractors. Make a left turn on Plattekill Avenue and continue to stop sign; bear right.
0.4	0.4	Turn left at stop sign onto Main Street (Starbucks on left) and proceed through town; cross bridge over Wallkill River.
0.8	0.4	Make first right turn onto Springtown Road.
1.3	0.5	Make first left onto Mountain Rest Road. Continue until stop sign at four corners.
2.5	1.2	Continue straight up steep hill. Several sharp curves will signal approach to the top of hill. Note outcrops of Martinsburg Formation on right, consisting of thin-bedded shale, minor siltstone, and very rusty weathering with fine-grained graywacke. Cleavage is poorly developed.
4.7	2.2	Turn left at Gatehouse onto Mohonk Mountain House grounds. Stop at gatehouse and check in. Proceed to guest parking about 2 miles ahead.
6.9	2.2	Park cars in last guest parking lot.

STOP 1. WE WILL BE SPENDING ABOUT 2 HOURS HERE EXAMINING VARIOUS OUTCROPS ON THE MOHONK MOUNTAIN HOUSE GROUNDS. Please, no hammers, only cameras!

The Shawangunk (pronounced *Shon-gum*) Formation (Figure 1) was deposited during the Silurian Period, approximately 438-418 million years ago. At that time a shallow sea covered the southeastern part of New York State into which drained rivers and streams. The bottoms of these streams were layered with pieces of abraded quartz. Constant motion of the water eroded the chunks of quartz into sub-rounded ‘pebbles’ that were eventually “glued” together by silica-rich cement carried by percolating ground water. The name of the sedimentary rock formed by this process is conglomerate (Figure 2). The exact source of the quartz that makes up the conglomerate is not clear. It may have formed by braided streams or alluvial fan deposition. A braided stream divides into an interlacing network of several small branching and reuniting shallow channels. An alluvial fan is a low outspread flat to gently sloping mass of rock material shaped like an open fan, deposited by a stream at the place where it issues from a narrow mountain valley upon a plain (Jackson, 1997). The Shawangunk Mountains (Figure 3) extend to the southwest into New Jersey where they are called the Kittatinny Mountains. The Shawangunk Formation ranges in thickness from 733m (2,200 ft) near Ellenville, to the southwest, to <1m (about 1 ft) near Hidden Valley, to the northeast, finally ‘pinching out’ and disappearing. Stratigraphically below the Shawangunk lies the Middle to Late Ordovician age (489-439 million years old) Martinsburg Formation, consisting mostly of shale, up to a thickness of almost 3333m (10,000 ft).

Glacial Geology

Note the beautiful glacially paved surface with striations and chattermarks. Chattermarks are small, closely spaced curved scars or cracks made by chipping a brittle bedrock surface, in this case the conglomerate, by rock fragments carried in the base of a glacier. Each chattermark is roughly transverse to the direction of ice movement and its ‘horns’ point in the direction the glacier moved. Compare the direction of the striations here (about



Figure 3. The Shawangunk Ridge that continues southwestward into New Jersey and Pennsylvania where it is overlain by the Bloomsburg Red Beds. In the foreground note the famous Trapps (Gunks) that, according to Van Diver (1985), as all mountain climbers know, constitute the best rock-climbing region of the eastern United States.

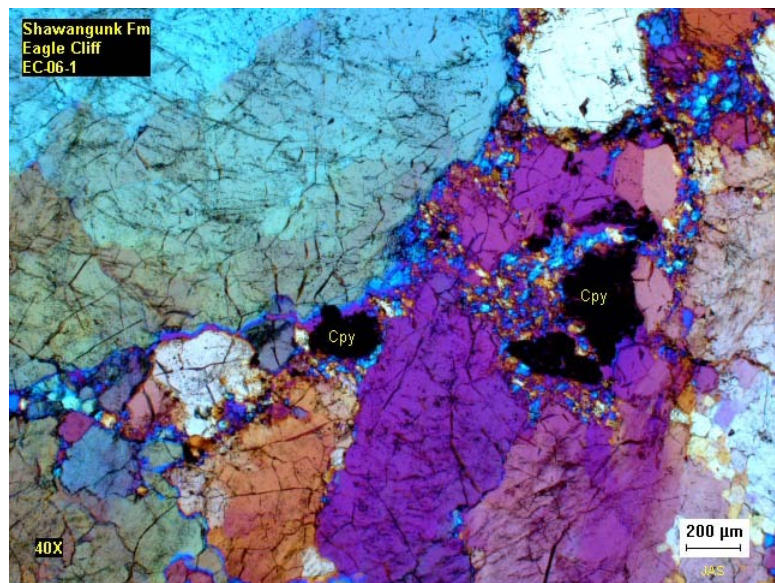


Figure 4. Chalcopyrite (copper-iron sulfide, Cpy) in the Shawangunk Conglomerate along Eagle Cliff Road, lake side.

10 degrees to the northeast) with the direction of the striations at the top of the crevice (about 20 degrees to the northwest). As the glacier moved down through the valley a small lobe moved around the ridge changing direction slightly resulting in a change in strike.

Mineralization within the Shawangunk Formation

The sulfide mineralization observed within the Shawangunk Formation along Eagle cliff consists of predominately marcasite and pyrite with small amounts of chalcopyrite (Figure 4). The sulfides are disseminated within the conglomerate. Emplacement appears to be associated with fault structures. SEM/EDS analysis of the sulfides also shows trace amounts of lead and zinc. Gossan in the form of goethite, hematite and jarosite has formed in places as a result of sulfide weathering. In some areas the water soluble sulfate minerals melanterite $\text{Fe}(\text{SO}_4) \cdot 7\text{H}_2\text{O}$ and rozenite $\text{Fe}(\text{SO}_4) \cdot 4\text{H}_2\text{O}$ are observed as a white to yellowish encrustation on the surface of sulfide containing exposures, when weather conditions permit (Figure 5). One from Eagle Cliff (valley side) sample contained two hydrated Fe-sulfate phases: melanterite $\text{Fe}(\text{SO}_4) \cdot 7\text{H}_2\text{O}$ and rozenite $\text{Fe}(\text{SO}_4) \cdot 4\text{H}_2\text{O}$. Rozenite is a lower hydrate containing 4 moles of water, as compared to melanterite which contains 7 moles of water. Rozenite can be formed in one of two ways: 1. Dehydration of melanterite or, 2. during lower relative humidity conditions; at room temperature rozenite is the stable phase below 70-80% RH. Examination by polarized light microscopy of the sample shows euhedral (well formed) crystals of rozenite encased in melanterite. This relationship indicates that the rozenite crystallized first, followed by the melanterite. Additionally, no alteration/dehydration textures were observed in the melanterite. From this we conclude that the rozenite crystallized during lower RH conditions, followed by melanterite which crystallized during increased RH conditions (i.e., above 70-80% RH depending on ambient temperature).

The Sky Lakes

Looking toward the southwest observe the ridge forming the “backbone” of the Shawangunks. Continuing along the ridge, located at its highest point about 30km (10 mi) away, is Sam’s Point and the Ice Caves, so called because of the cold air trapped at the bottom of open vertical joints. Within the deep recesses of the caves one can find ice that lasts all



Figure 5. Melanterite, a white precipitate, can be found in the Shawangunk Conglomerate along Eagle Cliff Road, valley side.



Figure 6. Mohonk Lake with a pH of 7 near the surface. This is one of five sky lakes on the ridge; the others are: Minnewaska, Awosting, Mud Pond and Maratanza.

summer long. Along the top of the Shawangunk Ridge are located four “Sky Lakes” in addition to Mohonk Lake (Figure 6), that are so-called because they receive their water supply solely from rainwater. These lakes are Minnewaska (closest to Mohonk), Awosting, Mud Pond and Maratanza. The lakes most likely overlie faults that weakened the bedrock, which was then scooped up by the passing glaciers. Interestingly, Mohonk Lake (Figure 6) is the only one of the five sky lakes that has greenish water; the others have clear blue water. The reason for this is that the acidic pH of the water of the four sky lakes (4.0) precludes algae and fish from living in them. However, Mohonk Lake’s water is buffered by the alkaline rocks of the underlying Martinsburg Formation that floors the southeastern portion of the lake and has a pH of 7-7.5. This raises the pH level such that the conditions are more conducive to life. Between the Shawangunks and the Catskill Front is a broad valley known as the Port Jervis trough. The trough was formed by erosion and glacial scouring of the relatively weak sediments beneath the Catskill Formation. The floor of the valley is fairly flat due to the deposition of glacial debris (till) and lake sediments. Looking toward the southeast one can see the Hudson Valley (the Hudson River is about 18km [6 mi] to the east).

Continue to the Huntington Lookout summer cottage. The view here is spectacular. Toward the southwest we see the Shawangunk Ridge and the Trapps (famous among technical rock climbers). To the northwest one can see the Catskill Mountains, also known as the Catskill Front, rising in the distance to a height of 700m (2,100 ft) above the Hudson River. The highest peak in the Catskill Mountains is Slide Mountain which rises 1,282m (3,846 ft) above sea level. These mountains began to rise in New England and the Canadian Maritime Provinces during the Devonian Period (due to a collision of tectonic plates known as the Acadian Orogeny). Throughout the Middle and Late Devonian Period the mountains were eroded by streams and rivers that flowed toward the west, carrying vast quantities of mud and sand that were deposited on the floor of a great inland sea as the rivers lost velocity. Closer to the ancient shoreline the heavier particles settled out first, leaving mostly sands and some shale, whereas farther from the shoreline the deposits consisted of mainly shales with some sand and, farthest from the shoreline we find only shale. Thus, as the rivers and streams lost velocity, the heaviest particles (larger sand-size grains) settled out first and the smallest, finer particles settled out last forming the fine-grained shales.

Make a right turn out of parking lot and proceed toward Gatehouse.

- 8.2 1.3 On right observe the (covered) contact between the Shawangunk and Martinsburg at Woodland Bridge. Note the relatively shallow dip of the shales typical of the proximity to the contact with the Shawangunk.

- 8.3 0.1 Make right turn into shale pit.

STOP 2. SHALE PIT. In this shale pit (Figure 7) the Martinsburg contains fault-related structures interpreted to be of two different ages. A *mélange*, or broken formation, is part of a *mélange*



Figure 7. Shale pit in the Martinsburg Formation dominated by medium-dark gray shale with fine-grained graywacke. The graywackes are fossiliferous and contain a fair amount of pyrite. The Martinsburg here contains fault-related structures interpreted to be of two different ages. A *mélange* zone that is Taconic in age has been carried westward over the Silurian Shawangunk in the hanging wall of the younger Kleine Kill thrust (See Epstein and Lyttle, 1987) that is Taconic in age with a strike of N5E-N10E. Most of the small faults (see eastern wall of the pit) and slickensided surfaces that are common at this location are probably related to this younger thrust fault. The trace of the Kleine Kill thrust has a strike of N15E and is interpreted to be Alleghanian in age.

zone that is Taconic in age that has been carried westward over the Silurian Shawangunk in the hanging wall of the younger Kleine Kill thrust. Most of the small faults and slickensided surfaces that are common in this locality are probably related to this younger thrust fault. The trace of the Kleine Kill thrust nearby has a strike of N15E and is interpreted to be Alleghanian in age. The fault zones in the Martinsburg which clearly do not cut Silurian and younger rocks generally have: (1) a trend more northerly than the northeast-trending folds and faults in the Silurian rock, (2) a diagnostic scaly cleavage, (3) tightly folded "floating" knockers of graywacke whose axes plunge predominantly to the northeast with variable azimuth, and (4) very little, if any, vein quartz. The faults that definitely cut both the Ordovician Martinsburg and the Silurian Shawangunk, and which we interpret to be Alleghanian in age, generally have: (1) a slightly more easterly strike, (2) well-developed "pencils" in the shale formed by the intersection of bedding and cleavage, and (3) vein quartz parallel to bedding and/or cleavage that commonly contains shale fragments.



Figure 8. Oscillation ripples in Martinsburg Formation, shale pit.

The exposure here consists of predominantly dark gray shales and siltstones interbedded with fine grained graywacke beds, occasional prominent pyrite layers and disseminated sphalerite, chalcopyrite and galena. Oscillation ripples (Figure 8) occur on some bedding surfaces. Carbonaceous material occurs mostly as fine-grained patches throughout the matrix. The studied section is tectonically stressed with shiny quartz slickensided surfaces, parallel cross-laminated strata and ripple marks. Crinoid stems, some disarticulated, and free columnals occur on different bedding surfaces, indicating a possible change in current regime. Scattered linear to sinusoidal horizontal burrow structures ranging in diameter from .5-3 cm are found on the silty beds. Some of the burrows are infilled with coarse quartz grains. The faunal constituents include brachiopods (93%), crinoids (*Ectenocrinus*; 3%), bivalves (3%), ostracodes (<1%), corals (<1%), trilobites (*Cryptolithus*, <1%) (Figure), conulariids (<1%) and unidentified burrowers (<1%). The brachiopods are represented by a low diversity assemblage of dalmanellids and what may be a new species of *Sowerbyella*. The fauna can be classified into distinct trophic groups: (1) high-level suspensions feeders (crinoids, corals); (2) low-level suspension feeders (brachiopods, bivalves); (3) animals that collect food from the sediment surface (ostracodes, trilobites); and (4) animals

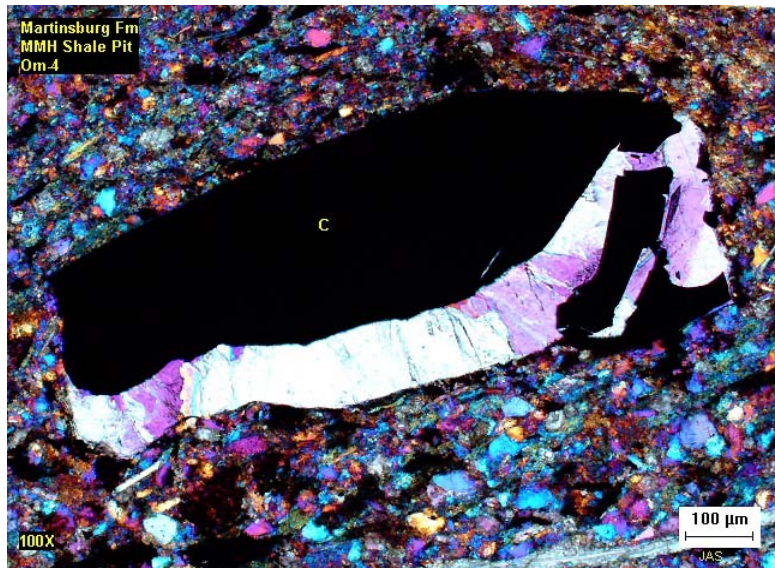


Figure 9. Carbonaceous material (thin section) in Martinsburg Formation, shale pit.

that feed within the sediment (burrowers). This partition of feeding niches leads to a reduced competitive trophic structure and therefore increased community stability.

Mineralization within the Martinsburg Formation

Within the Martinsburg Formation at the Mohonk shale pit sulfide mineralization has also been observed. The sulfides include pyrite, chalcopyrite, galena and sphalerite. The sulfides are found associated with calcite veins and as disseminations throughout the shale. Carbonaceous material has also been observed in thin section (Figure 9).

		Make right turn out of shale pit and proceed to gatehouse. At stop sign make right turn onto Mountain Rest Road. RESET ODOMETER TO ZERO!
0.0	0.0	Turn right at stop sign just past Gatehouse (Mountain Rest Road).
2.1	2.1	Make right turn at the bottom of the hill onto Butterville Road. As we progress in a southerly direction on Butterville Road observe cliffs of the extremely resistant Shawangunk Formation on the right. The lowland and hills in the foreground are underlain by the shales and graywackes of the Martinsburg Formation, striking toward the northeast. Note Sky Top (tower) from which one can see six states (New York, New Jersey, Pennsylvania, Connecticut, Vermont, and Massachusetts). Sky Top sits atop a rock scramble and steep climb through the “Crevice” and ending with an egress called the “Lemon Squeeze.” This is the signature hike of Mohonk Mountain House that is adjacent to the 6,400 acre Mohonk Preserve.
3.7	1.6	At the stop sign make a right turn onto NY 299 heading west. On the right note the overturned shales and thin graywackes of the Martinsburg Formation. Here there are several narrow fault zones with vein quartz.
4.8	1.1	On the right note the steeply dipping and overturned Martinsburg shales and thin graywackes. Ahead view the “Trapps” a world famous site for rock technical climbers located on the Mohonk Preserve. The Mohonk Preserve, founded in 1963, was established to protect the northern Shawangunk Ridge. Its mission is to protect the ecology of the area and to provide for public environmental education and recreation.
5.0	0.2	Jenkins-Lueken orchards.
5.5	0.5	Steeply dipping and overturned Martinsburg shales and thin graywackes on right. View of the Shawangunk cliffs at the Trapps straight ahead.
6.0	0.5	Note Martinsburg with graywackes on right. The bedding appears to be right-side-up however, the poorly developed cleavage in the shales dips steeply to the west suggesting that both bedding and cleavage have been rotated by faulting.
7.3	1.3	At stop sign make right turn onto US 44/55. We will be climbing to the top of the Shawangunk Mountains for the next mile or so.
7.8	0.5	Visitor’s Center, Mohonk Preserve on right. Shortly we will be making a sharp left turn with the Martinsburg (on the right) and shale interbedded with thin-bedded (up to 5 in), cross-bedded to planar laminated siltstone, and minor fine-grained sandstone. Soft-sediment slump folds are common in the siltstones. Cleavage is absent or very poorly developed, except near tight folds and narrow fault zones. Most of these faults and folds do not affect the overlying Shawangunk Formation and must be Taconic in age. They trend from N5E to N20E, whereas structures in the overlying Shawangunk trend in a more easterly direction.
8.6	0.8	On the left is a scenic overview of the Wallkill Valley.
8.7	0.1	The contact between the Shawangunk and Martinsburg formations is exposed at two spots approximately 400 feet southwest of the road at the base of the cliffs. Fairly regular and linear mullions can be observed in the basal Shawangunk at the contact.
8.8	0.1	As we pass under the Trapps Bridge note the conglomerates and quartzites of the Shawangunk dipping to the northwest (31NW).

10.2 1.4

Cross Coxing Kill. As we cross Coxing Kill note that the Coxing Kill Valley coincides with the trough of a broad, open syncline that plunges gently to the northeast. This is also a favorite site for skinny dipping. For the next 0.5 mile we will begin to cross a broad open anticline that exposes a window of Martinsburg that is about 2 miles long. To the right note an outcrop of basal Shawangunk with unconformably underlying shales of the Martinsburg exposed about 40 feet away. Dips in both



Figure 10. “Pods” in Shawangunk Conglomerate are often conical with flat bottoms and associated with quartz pebbles. Note that the quartzite is recrystallized. The matrix filling the pods is dark gray and seems to follow the bedding and/or crossbedding and cleavage.

units are very gentle and the angle of unconformity is as little as 2°. The divergence in strike, however, is as much as 38°, with the strike in the Martinsburg being more northerly. The Martinsburg here is in the broad open-fold Taconic tectonic zone (zone 3 of Epstein and Lytle, 1987).

10.7 0.5

Peters Kill parking area on right.

11.8 1.1

Entrance to Minnewaska State Park on the left.

12.0 0.2

Trailhead leading to Lake Awosting on left.

12.5 0.5

Cross Sanders Kill.

13.8 1.3

Turn into small parking area with scenic overview of the Rondout Valley underlain by Upper Silurian through Middle Devonian rocks that are buried by a variety of glacial sediments. In the distance note the Middle and Upper Devonian clastics forming the Catskill Mountains. There is a thrust fault with nice slickensides across the road from the parking area. BE CAREFUL OF ONCOMING CARS!



Figure 11. Large “pod” in Shawangunk Conglomerate. Note quartz pebbles at bottom right.

STOP 3. MYSTERIOUS ‘PODS’. Walk back along NY 44/55 (north side of road) and observe the mysterious ‘pods’ of Feldman et al., (2009) that follow the beds (N20E; 38NW) in the upper part of the Shawangunk Formation. The pods (Figures 10, 11) are often conical with flat bottoms and associated with quartz pebbles. Note that the quartzite is recrystallized. The matrix filling the pods is often dark gray and seems to follow the bedding and/or crossbedding. WHAT ARE THEY? The pods range from 0.2 to 7 cm across (at the base) and 0.2 to 6 cm in height. Internally, the ‘pods’ are mostly non-recrystallized quartz grains with interstitial mica and possibly clays (Figure 12). In contrast, the areas outside the ‘pods’ consist of pressure solution welded quartz grains consistent with the quartzitic nature of the formation. The pods are restricted to a relatively narrow stratigraphic interval and show no evidence of internal bedding. The pods may be: (1) microbial mounds/algal mats, (2) sponges, (3) mud balls or SOMETHING ELSE! They are similar in general outline to some thrombolites in that the internal texture is non-laminated but there is no indication of microbial activity such as clotting. The possibility that they are sponges or algal mats is somewhat doubtful because the braided stream depositional environment was one of relatively fast-moving fresh water draining mountains to the southeast; in addition, most sponges live in a marine environment. The conical shape does not favor a mud ball origin. Hint: note that the gray matrix follows the cleavage. PLEASE DO NOT COLLECT SAMPLES; ACTIVE RESEARCH SITE!

END OF TRIP

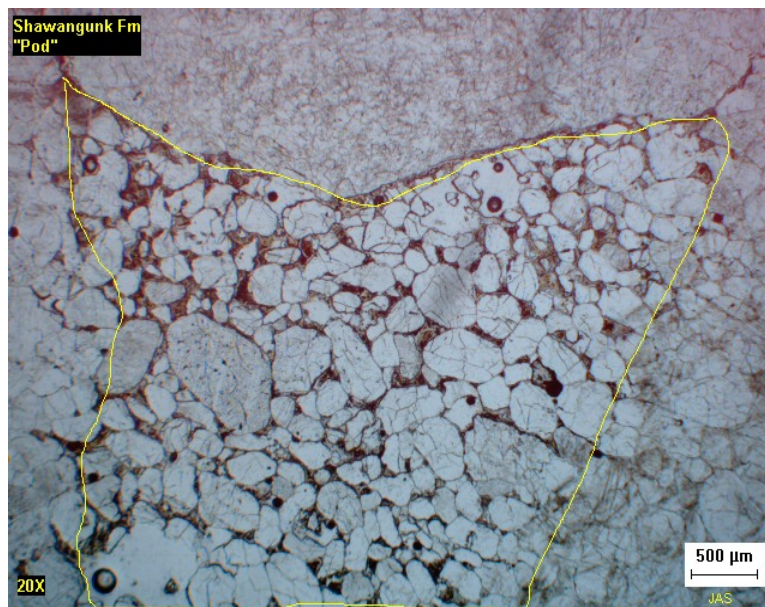


Figure 12. Thin section of “pod” containing non-recrystallized quartz grains with interstitial mica and possibly clays. In contrast, the areas outside the pod consist of pressure solution welded quartz grains.

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