New York State Geological Association

82nd Annual Meeting

Field Trip Guidebook

Hosted by
College of Staten Island/CUNY
Staten Island, NY, 10314
September 24-26, 2010

Edited by
Alan I. Benimoff

Published by the
New York State Geological Association
www.nysga.net
www.nysgaonline.org

ISBN 1061-8724

Front Cover Image Credit

Geologic map of Staten Island, NY from New York City folio, Paterson, Harlem, Staten Island, and Brooklyn quadrangles, New York-New Jersey *Folio 83*, published 1902 by the United States Geologic Survey

Table of Contents

Delaware Water Gap, A Geology Classroom

By Jack B. Epstein
U.S. Geological Survey

INTRODUCTION

The Delaware Water Gap National Recreation Area (DEWA) contains a rich geologic and cultural history within its 68,714 acre boundary. Following the border between New Jersey and Pennsylvania, the Delaware River has cut a magnificent gorge through Kittatinny Mountain, the Delaware Water Gap, to which all other gaps in the Appalachian Mountains have been compared. Proximity to many institutions of learning in this densely populated area of the northeastern United States (Fig. 1) makes DEWA an ideal locality to study the geology of this part of the Appalachian Mountains. This one-day field trip comprises an overview discussion of structure, stratigraphy, geomorphology, and glacial geology within the gap. It will be highlighted by hiking a choice of several trails with geologic guides, ranging from gentle to difficult. It is hoped that the "professional" discussions at the stops, loaded with typical geologic jargon, can be translated into simple language that can be understood and assimilated by earth science students along the trails. This trip is mainly targeted for earth science educators and for Pennsylvania geologists needing to meet state-mandated education requirements for licensing professional geologists. The National Park Service, the U.S. Geological Survey, the New Jersey Geological Survey, and local schoolteachers had prepared "The Many Faces of Delaware Water Gap: A Curriculum Guide for Grades 3–6" (Ferrence et al., 2003). Portions of this guide, "The Many faces of Delaware Water Gap" appear as two appendices in this field guide and is also available by contacting the Park (http://www.nps.gov/dewa/forteachers/curriculummaterials.htm). The trip will also be useful for instruction at the graduate level. Much of the information presented in this guidebook is modified from Epstein (2006, 2010).

Figure 1 The Delaware Water Gap National Recreation Area lies within the center of the northeast megalopolis and within a few hours' drive of one-third of the Nation's urban population.

Please note: It is illegal to collect rocks, plants, or parts of animal life in National Park lands.

PREFACE (FROM "THE MANY FACES OF DELAWARE WATER GAP")

Geology is a topic that many students, both children and adults, find confusing. And because they do not see how geology relates to their own personal situation, they also find it dull. Yet beneath their feet are stories hidden in the rocks. If these rocks could speak, they would tell of dramas that have unfolded. Of colliding continents, of oceans come and gone, of inhabitants that are no more, of mountains of ice that scraped away the evidence of time gone by. These rocks have seen much. They have controlled the fate of the Delaware Valley and have passively influenced everything from the local environment to human history and economics.

To engage students in the study of geology, we must teach them the basics, but we must also help them to understand how the rocks that they see relate to more intangible ideas. We must help them to discover clues to the ancient past. We must encourage them to explore how local geology controls the distribution of plants and animals, through its influence on soil types, precipitation, and temperature. We must engage them in inquiry about how the landscape has controlled human settlement, keeping folks out of hard to reach places or luring them in for valuable natural resources. We must facilitate an understanding of how geology is tied to our freedom, having changed the pathways and outcomes of wars.

Delaware Water Gap National Recreation Area is an outdoor classroom waiting to be discovered. Hidden in the ridges and valleys of the park, the patient explorer can find a seemingly endless array of animals and plants, each living in a microhabitat controlled by the rocks. The plant and animal communities range from aquatic river communities to cactus barrens, providing myriad opportunities for students to study biological and geological concepts and the inter-relationships between the two.

Today, the Poconos is a major destination (as it has been since the early 1800s), attracting visitors with hundreds of waterfalls, heart-shaped lakes, and the famous Delaware Water Gap. Though transportation has changed from carriage, to train, to automobile, the scenic river valley continues to inspire its guests. It is a landscape shaped by glaciers, a mountain shaped by time, an economy shaped by geology.

Teachers, geologists, and National Park Service staff worked in partnership to develop these materials. It is our hope that this curriculum will make it easier for students to understand geology and recognize the significance of the geologic story preserved in our National Parks. It is our belief that a visit to the park, for a personal encounter with the local geology, will help students to internalize the materials studied in class and will inspire curiosity about the role of geology in their lives and those of their ancestors. We hope these materials make it easier for you to use park resources to teach geology.

DELAWARE WATER GAP NATIONAL RECREATION AREA

Following a disastrous flood on the Delaware River in 1955, the U.S. Army Corps of Engineers had proposed building a dam across the river at Tocks Island, about eight miles upstream from the Gap, and forming a reservoir nearly 40 miles long ending near Port Jervis, New York. The dam was authorized by Congress in 1962, but following much controversy, the dam was considered environmentally and structurally unsound, and it was officially de-authorized in 2002. The land in the Delaware Valley that was acquired by Congress for the proposed dam was officially designated a Recreation Area in 1965 and in 1992 the river became part of the National Wild and Scenic Rivers System.

The DEWA is the most heavily visited National Park Service (NPS) facility in the northeastern United States, attracting more than five million visitors yearly. It offers an oasis of solitude in the midst of this urbanized megalopolis. Recreational opportunities include fishing, hunting, boating, swimming, and hiking. The area also offers insight into the biologic diversity, cultural history, and geologic evolution of this part of the Appalachian Mountains.

DEWA straddles two physiographic provinces—the Pocono Plateau on the northwest and the Valley and Ridge to the southeast. Three Paleozoic orogenies affected patterns of sedimentation, resulting in a wide range of depositional environments, ranging from deep-sea, shallow marine, to terrestrial. Additionally, the area also experienced a varied and complex history of deformation. The Pocono Plateau is underlain by flat-lying to gently inclined sandstone and shale of Middle and Late Devonian age. More complexly folded rocks form the Valley and Ridge Province. The rocks include a variety of shale, sandstone, limestone, and dolomite of Ordovician, Silurian, and Devonian age, aggregating ~9000 feet in thickness within the park boundaries (Table 1). The entire stratigraphic sequence exposed in the Recreation Area spans ~65 million years. The Delaware Valley is carved into the softer shale and limestone, whereas the surrounding mountains are held up by the more resistant sandstone. The northeast-trend of the folds in the area controls the orientation of these features.

During the past 2 million years, the area experienced several periods of glaciation. Retreat of the latest (Wisconsinan) glacier, which began ~20,000 years ago (Witte, 2001), left behind varied scenery. The glacier carved and sculpted the landscape, deposited a variety of glacial drift, carved out lakes, and rearranged some of the stream drainage. The present Delaware River has cut through a silt and sand terrace that was occupied by Native Americans ~11,000 years ago. The area offers a wide variety of geologic subjects for future study and re-evaluation, including the regional framework and plate tectonic history of the Appalachian Mountains. Many controversial concepts in geomorphology are emphasized in this classic area, including accordant summits and peneplains, the formation of wind and water gaps, and the origin of waterfalls as they relate to the complex glacial history of the area.

Age	Formation	Approximate Thickness (feet)	Description
Devonian	Catskill Formation	500+	Sandstone and lesser shale; forms uplan Pocono Plateau
	Mahantango Formation	2,000	Siltstone and silty shale; forms steep slo northwest of the Delaware River
	Marcellus Shale	800	Shale; underlies the Delaware River northeast of Flatbrook Bend
	Buttermilk Falls Limestone	275	Limestone, calcareous shale, and chert
	Schoharie Formation	100	Calcareous, argillaceous siltstone
	Esopus Formation	180	Shaley siltstone and shale
	Oriskany Group	100	Sandstone, calcareous shale and siltston chert
	Helderberg Group	300	Limestone, calcareous shale, calcareous sandstone
	Rondout Formation	30	Dolomite, calcareous shale, limestone
Silurian	Decker Formation	80	Arenaceous limestone, calcareous sands dolomite
	Bossardville Limestone	100	Limestone
	Poxono Island Formation	500	Dolomite, shale, limestone; underlies Delaware River southwest of Flatbrook
	Bloomsburg Red Beds	1,500	Red sandstone, siltstone, shale; forms northwest slope of Kittatinny Mountain
	Shawangunk Formation	1,400	Sandstone and conglomerate; holds up Kittatinny Mountain
Ordovician	Martinsburg Formation	1,000+	Slate and graywacke; forms southeast sl of Kittatinny Mountain

Table 1. Generalized description of rock units in the Delaware Water Gap National Recreation Area, New Jersey and Pennsylvania. The Martinsburg, Shawangunk, and Bloomsburg are the only formations that will be visited on this field trip.

STRATIGRAPHY

Delaware Water Gap owes its notoriety to the depth to which the river has cut through Kittatinny Mountain. Exposures of 3000 feet of Silurian clastic rocks are nearly continuous; the entire Shawangunk Formation, with its three members, and most of the Bloomsburg Red Beds are visible (Fig. 2). The underlying Martinsburg crops out in widely scattered exposures and small abandoned quarries just south of the gap.

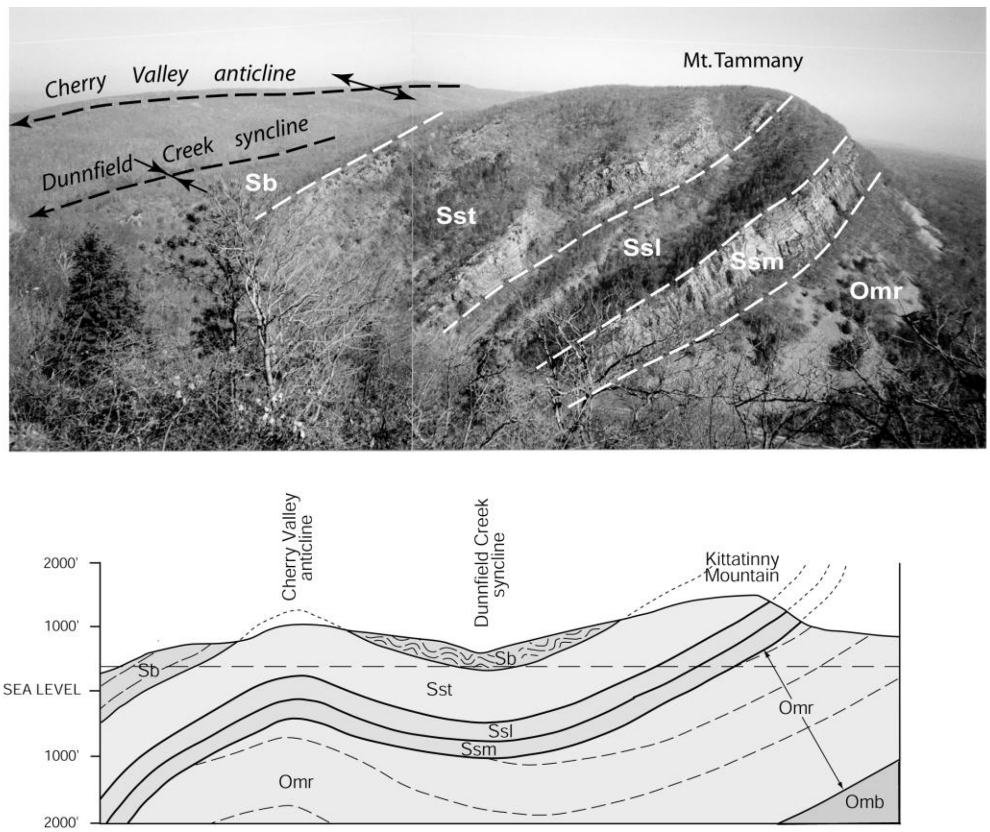


Figure 2. Delaware Water Gap in New Jersey as viewed from atop Kittatinny Mountain (Mt. Minsi) on the Pennsylvania side. Omb, Bushkill Member of the Martinsburg Formation; Omr, Ramseyburg Member of the Martinsburg Formation; Ssm, Minsi Member of the Shawangunk Formation; Ssl, Lizard Creek Member of the Shawangunk Formation; Sst, Tammany Member of the Shawangunk Formation; Sb, Bloomsburg Red Beds. Small-scale folds in the Bloomsburg are located only in the Dunnfield Creek syncline. The angular discordance at the Ss-Om Taconic contact is less than 5 degrees (Beerbower, 1956). Horizontal dashed line shows rocks possibly visible along I-80 in New Jersey.

The Martinsburg is more than 15,000 feet thick in eastern Pennsylvania, consisting of three members: a lower Bushkill Member of thin-bedded slates, middle Ramseyburg Member with abundant greywacke packets, and an upper Pen Argyl Member with medium- to thick-bedded slate and some greywacke (Drake and Epstein, 1967). These sediments were deposited in a rapidly subsiding flysch-turbidite basin (Van Houten,

1954) formed during Middle Ordovician continental plate collision. The highland source for the Martinsburg was "Appalachia" to the southeast, and the sediments covered a foundered Cambrian and Ordovician east-facing carbonate bank. The greywackes were probably deposited in submarine channels and were triggered by earthquakes during the Ordovician. The contact between the Pen Argyl and Ramseyburg Members disappears under the Shawangunk just within the confines of Delaware Water Gap National Recreation Area one mile west of Delaware Water Gap (Epstein, 1973). The Pen Argyl does not reappear in New Jersey to the northwest. Several small slate quarries and prospects in the Ramseyburg Member, all long since abandoned, are found within the DEWA boundaries (Epstein, 1974a). The deepening of the Ordovician basin in which the Martinsburg detritus was deposited was followed by tectonic uplift reflecting intense Taconic mountain building, which peaked with emergence of the area during the Late Ordovician. This period of orogenic activity and regional uplift was followed by deposition of a thick clastic wedge, the lowest unit of which consists of coarse terrestrial deposits of the Shawangunk Formation. The contact between the Shawangunk and Martinsburg is a regional angular unconformity. The discordance in dip is not more than 15 degrees in northeastern Pennsylvania, New Jersey, and southeastern New York (Epstein and Lyttle, 2001).

The Shawangunk was divided into three members at the gap (Epstein and Epstein, 1972). The upper and lower conglomeratic-sandstone members, the Minsi and Tammany are believed to be fluvial in origin and are interposed by a transitional marine-continental facies (the Lizard Creek Member). The fluvial sediments are characterized by alternations of polymictic conglomerate with quartz pebbles more than 2 inches long, conglomeratic sandstone, and sandstone (cemented with silica to form quartzite), and subordinate siltstone and shale. The bedforms (planar beds and cross-bedding) indicate rapid flow conditions. Cross-bed trends are generally unidirectional to the northwest. The minor shale and siltstone beds are thin, and at least one is mudcracked, indicating subaerial exposure. These mudcracks may be seen at mileage 104.9 at the south entrance to Delaware Water Gap on the New Jersey side by looking up ~50 feet at an overhanging ledge (see Fig. 13). These features indicate that deposition was by steep braided streams flowing toward the northwest with high competency and erratic fluctuations in current flow and channel depth. Rapid runoff was undoubtedly aided by lack of vegetation cover during the Silurian. The finer sediments present are believed to be relicts of overbank and backwater deposits. Most of these were flushed away downstream to be deposited in the marine and transitional environment represented by the Lizard Creek Member of the Shawangunk Formation.

The Lizard Creek Member contains a variety of rock types, and a quantity of sedimentary structures that suggest that the streams represented by the other members of the Shawangunk flowed into a complex transitional (continental-marine) environment, including tidal flats, tidal channels, barrier bars and beaches, estuarine, and shallow neritic. These are generally energetic environments, and many structures, including flaser bedding (ripple lensing), uneven bedding, rapid alternations of grain size, and deformed and reworked rock fragments and fossils support this interpretation (Epstein and others., 1974). The occurrence of collophane, siderite, and chlorite nodules and *Lingula* (phosphatic brachiopod) fragments at Lehigh Gap, 25 miles to the southwest, indicate near-shore marine deposition. Many of the sandstones in the Lizard Creek are supermature, laminated, rippled, and contain heavy minerals concentrated in laminae. These are believed to be beach or bar deposits associated with the tidal flats.

The outcrop pattern of the Shawangunk Formation and the coarseness of some of the sediments, suggest that they were deposited on a coastal plain of alluviation with a source to the southeast and a marine basin to the northwest. Erosion of the source area was intense and the climate, based on study of the mineralogy of the rocks, was warm and at least semi-arid. The source was composed predominantly 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 rocks in the Bloomsburg are in well-defined to poorly defined upward fining cycles that are characteristic of meandering streams. The cycles are as much as 13 feet thick and ideally consist of basal cross-bedded to planar-bedded sandstones that truncates finer rocks below. These sandstones were deposited in stream channels and point bars through lateral accretion as the stream meandered. Red shale clasts as long as three inches were derived from caving of surrounding mud banks. These grade up into laminated finer sandstone and siltstone with small-scale ripples indicating decreasing flow conditions. These are interpreted as levee and crevasse-splay deposits. Next are finer overbank and floodplain deposits containing irregular carbonate concretions. Burrowing suggests a low-energy tranquil environment; mudcracks indicate periods of desiccation. The concretions are probably caliche precipitated by evaporation at the surface. Fish scales in a few beds (seen near the tollbooth along I-80 in Pennsylvania) suggest marine transgressions onto the low-lying fluvial plains, perhaps in a tidal-flat environment.

The source for the Bloomsburg differed from that of the Shawangunk Formation because the red beds required the presence of iron-rich minerals suggesting an igneous or metamorphic source. Evidently, the source area was eroded down into deeper Precambrian rocks.

Upper Silurian and Lower Devonian rocks younger than the Bloomsburg Red Beds hold up Godfrey Ridge just north of the town of Delaware Water Gap. These rocks span the complete range of sedimentary types and reflect an equally complex series of depositional environments, including shallow marine shelf, supratidal and intertidal flats, barrier bars, and many neritic zones. Fossils are plentiful in many of the units.

STRUCTURAL GEOLOGY

Four rock sequences of differing tectonic style have been recognized in northeastern Pennsylvania. These *lithotectonic units* are presumed to be bound by décollements (detachments along a basal shearing plane or zone). Type and amplitude of folds are controlled by lithic variations within each lithotectonic unit.

Lithotectonic unit 1 comprises the Martinsburg Formation. Slaty cleavage is generally well developed in its pelitic rocks, and may be seen in several exposures along U.S. 611 south of the Gap. The slight angular discordance with the overlying Shawangunk is buried beneath talus at the Gap. At Stop 1 we will discuss this angular unconformity.

Lithotectonic unit 2 is made up of resistant, competent quartzite and conglomerate of the Shawangunk Formation and the overlying finer clastic rocks of the Bloomsburg Red Beds. Concentric folding by slippage along bedding planes is common. Cleavage is found within the shale and siltstones of this unit, but it is not so well developed as in the Martinsburg where slates have been commercially extracted. The reason for this is not because of different time of formation (e.g., Taconic or Alleghanian), but because of slight lithologic differences; the shale of the Martinsburg Formation were more uniform and of finer grain than those in the Silurian clastic rocks. Folds are generally open and upright (Fig. 2), but some limbs are overturned (discussed at Stop 1). In the Water Gap, the Bloomsburg is disharmonicaly folded, appearing as many small folds in the core of the Dunnfield Creek syncline. These can be seen south of the town of Delaware Water Gap in Pennsylvania. Cleavage in the Bloomsburg Formation dips southeast and appears to have been rotated during later folding. Numerous bedding-plane faults, many with small ramps in the Bloomsburg contain slickensides with steps that indicate northwest translation of overlying beds, regardless of position within a given fold (see Figure 18). Dragging of cleavage along some of these faults indicate that faulting postdated cleavage development, which in turn, predated folding.

The home for rocks of lithotectonic unit 3 is in a narrow ridge (Godfrey and Wallpack ridges) northwest of Kittatinny and Blue Mountains. Folds in this sequence in the southwestern part of the area are of smaller scale than surrounding units. Axes of these folds are doubly plunging and die out within short distances, making for complex outcrop patterns (Epstein, 1973, 1989). Folding becomes less intense in the northeast part of the Recreation Area where units 2 and 3 dip uniformly to the northwest.

There is a sharp contrast between the structure of lithotectonic units 4 and 3. Unit 4 makes up rocks of the Pocono Plateau north of the Delaware River. These rocks dip gently to the northwest and are interrupted throughout the area by only sparse and gentle upright folds. Cleavage is present, but not as well developed as in underlying rocks. Southwest of the field trip area, however, cleavage in Middle Devonian shale and siltstones is so well developed that these rocks were quarried for slate in the past near Lehigh Gap.

Three décollements, or zones of décollement in relatively incompetent rocks, are believed to separate the four lithotectonic units. The Martinsburg-Shawangunk contact is interpreted to be a zone of detachment between lithotectonic units 1 and 2. Thin fault gouge and breccia, ~2 inches thick, is present at the contact, such as at Yards Creek, ~5 miles northeast of Delaware Water Gap (Inners and Fleeger, 2001, p. 171). Elsewhere, such as at Lehigh Gap (Epstein and Epstein, 1969), and at exposures in southeastern New York (Epstein and Lyttle, 1987), thicker fault gouge, bedding-plane slickensides containing microscarps or steps, and drag folds indicate northwest movement of the overlying Shawangunk Formation.

I have concluded that the dominant northwest-verging folds and related regional slaty cleavage in all rocks in the Delaware Water Gap area were produced during the Alleghenian orogeny and are superimposed upon Taconic structures in pre-Silurian rocks (Epstein and Epstein, 1967; Epstein, 1974b). The regional slaty cleavage formed after the rocks were indurated at, or just below, conditions of low-grade metamorphism. Estrangement of the effects of the two orogenies is still the subject of considerable debate. An Alleghanian age for the cleavage in the Martinsburg at Lehigh Gap, 25 miles to the southwest, was established by Wintsch and others (1996) with whole rock 40Ar/39Ar dating of the micas in the slate.

GEOMORPHOLOGY OF DELAWARE WATER GAP

Following uplift during the late Paleozoic orogeny, the original divide of the Appalachian Mountains probably lay somewhere to the east within the area of the present Piedmont or Valley and Ridge physiographic province. During rifting and opening of the Atlantic Ocean, that divide shifted westward toward its present position in the Appalachian Plateau. The divide migration was due to the erosional advantage of the steeper streams that flowed eastward toward the Atlantic Ocean as compared to the gentler gradient of streams that flowed westward toward the continental interior. The manner of migration of that divide and how the streams cut through the resistant ridges are critical elements in any discussion of Appalachian geomorphic development and have been a source of considerable controversy for more than a century.

Several wind and water gaps are present in Kittatinny and Blue Mountains, the southernmost ridge of the Ridge and Valley Province. Viewed from a distance, these gaps or low sags interrupt the fairly flat ridge top that was termed the “Schooley peneplain” by Davis (1889) and popularized by Johnson (1931). Ideas on the origin of these gaps are critical factors in the discussions about the geomorphic development of the Appalachians. Those hypotheses that favor down cutting (superposition) from an initial coastal plain cover (Johnson, 1931; Strahler, 1945) require that the location of the gaps be a matter of chance. Those hypotheses favoring the present drainage divide having been inherited from the pattern already established following the Alleghanian orogeny and controlled by the topography and structure prevalent at the time (Meyerhoff and Olmstead, 1936) or by headward erosion into zones of structural weakness (headward piracy, Thompson, 1949) require that there be evidence for structural weakness at the gap sites.

Geologic mapping during the past few decades indicates that the gaps along the ridge held up by the Shawangunk Formation in northeastern Pennsylvania and northern New Jersey are located at sites where there are structures that are not present between these sites. The general conclusion can be made that the gaps are located at sites of structural weakness. If this opinion is accepted, then those hypotheses which suggest that streams sought out weaknesses in the rock during headward erosion are favored.

The following are features that appear to be related to most gap sites: (1) dying out of folds along plunge within short distances; (2) narrow outcrop widths of resistant beds because of steep dips; (3) more intense folding locally than nearby; (4) abrupt change in strike owing to kinking along strike; (5) intense overturning of beds and resultant increase in shearing; (6) cross faulting; and (7) progressive erosion exposing narrower widths of resistant rocks compared to areas nearby (Epstein, 1966). Details at Delaware Water Gap will be discussed at Stop 1.

GLACIAL GEOLOGY

The latest (Wisconsinan) glacial advance into eastern Pennsylvania and northern New Jersey resulted in the deposition of a conspicuous terminal moraine which crosses the Delaware River ~11 miles south of the gap near Belvidere, New Jersey (Fig. 3). The moraine reached heights of more than 100 feet in places. The glacier was at least 2,000 feet thick in the Gap at one time. As the glacier retreated from its terminal position north of Blue Mountain, the meltwater was dammed between the terminal moraine, the surrounding hills, and the retreating ice front. A series of stratified sand and gravel deposits sequentially were laid down in the lake that formed and as the glacier retreated. The lake has been named Lake Sciota, after the classic delta and varved lake-bottom sediments that are found there (Epstein and Epstein, 1967). The lake reached a depth of ~200 feet in places. Initially, the outlet for the lake was over the terminal moraine at Saylorsburg and the water flowed west toward the Lehigh River. As the glacier retreated northeastward past the Delaware River, the waters drained through the gap and the lake ceased to exist.

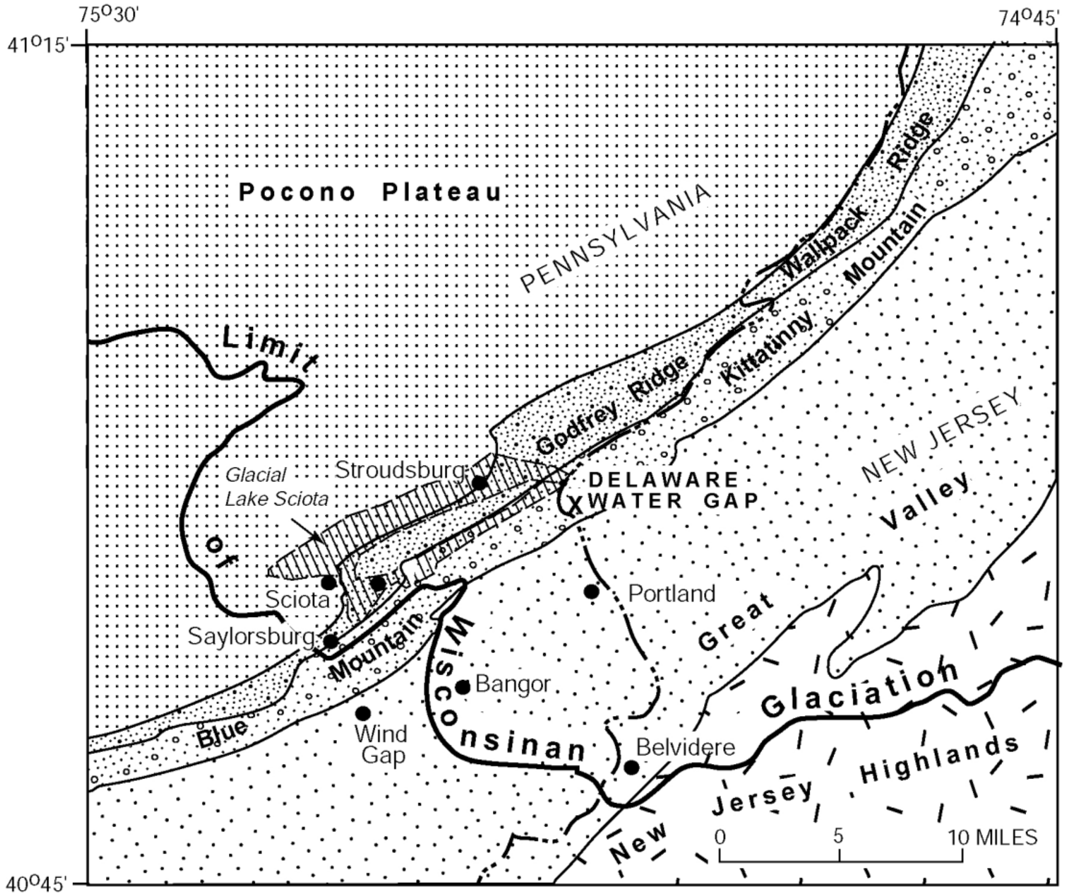


Figure 3 Physiographic map of part of easternmost Pennsylvania and northwestern New Jersey showing the position of the maximum advance of the Wisconsinan glacier. Modified from Epstein (1969).

A variety of glacial deposits formed in the Delaware Water Gap area, comprising varying proportions of gravel, sand, silt, and clay. On the basis of texture, internal structure, bedding and sorting characteristics, and generally well-preserved landforms, the deposits have been subdivided into till (ground, end, and terminal moraine) and stratified drift (delta, glacial-lake-bottom, kame, kame-terrace, and outwash deposits). Below the gap is an outwash terrace, more than 150 feet high on both sides of the river, comprising very coarse gravel with boulders exceeding eight feet in length. This deposit may be seen at mile 91.9 and 99.8 of the road log. Numerous striae, grooves, and roches moutonnee formed by Wisconsin glacial erosion are found on bedrock surfaces in most parts of the area. Striae trends show that the ice was strongly deflected by underlying bedrock topography. Whereas the average direction of flow of the ice sheet around Delaware Water Gap was ~ S. 20°W., the base of the ice traveled more southwestward parallel to the valley bottoms and about due south over the ridge top. Several of these glacial features can be seen along the trail of Stop 5.

Bedrock topography has been subdued in many places by the drift cover. Examples of drainage modifications are numerous. Talus deposits, congelifractates, rock streams, and rock cities are believed to be partly of periglacial origin. Numerous lakes, mostly in kettle holes, have made the Pocono area the tourist attraction that it is.

ROAD LOG AND STOP DESCRIPTIONS

Figure 4 shows the stop locations

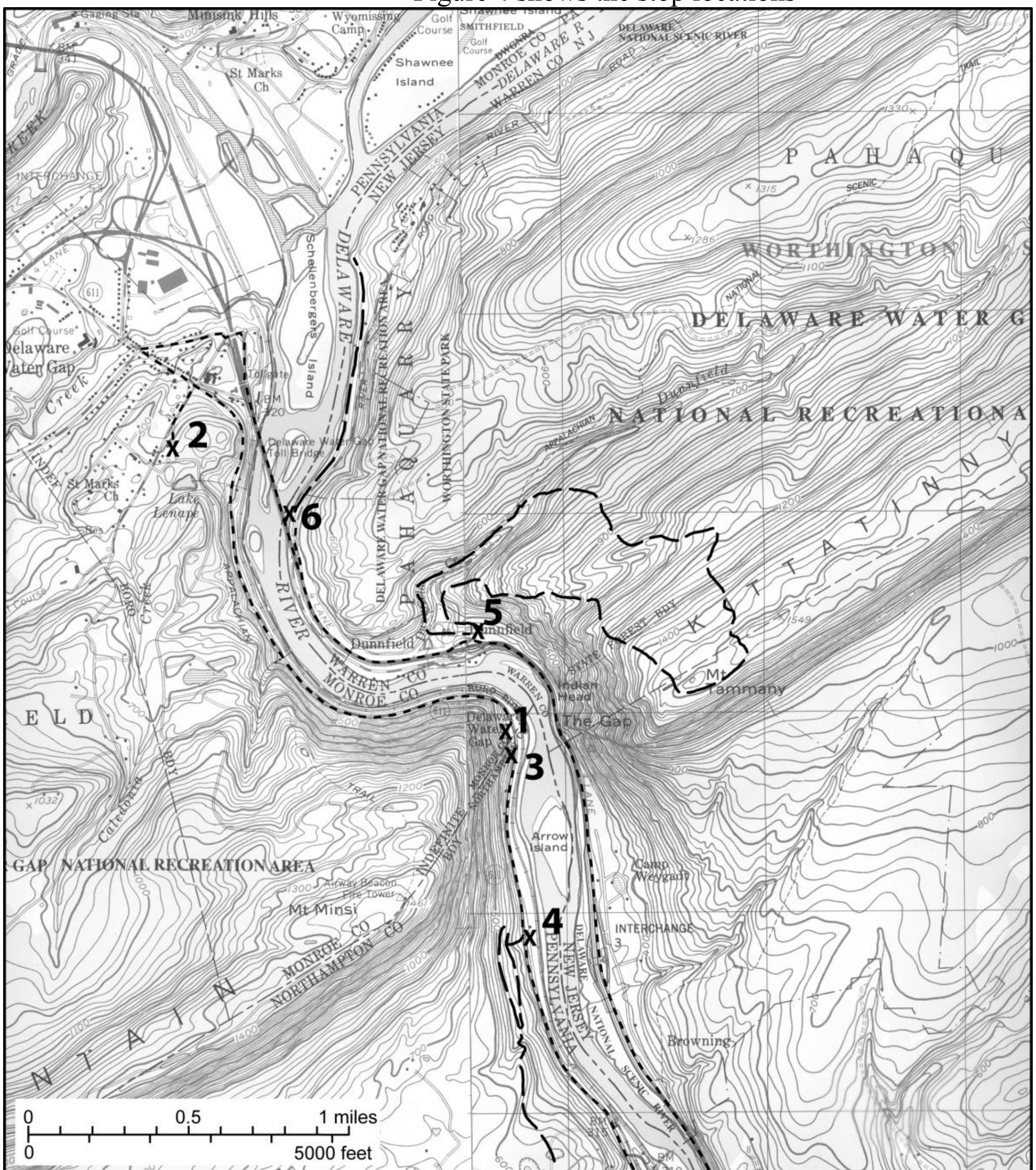


Figure 4. Topographic map of the Delaware Water Gap area showing route (short dashes), stop localities and trailheads, and trail routes (long dashes). 1, Point of Gap overlook; 2, Lake Lenape; 3. Cold Air Cave; 4. Arrow Island trail; 5, Red Dot-Blue

Blaze-Dunnfield Creek Trail; 6. Karamac trail.

00.0	0.3	Leave CUNY College of Staten Island, merge onto NY-440N at Victory Blvd
00.3	3.7	Merge onto I-278 via Exit 10W toward RT-440S/Outerbridge Crossing/Goethals Bridge. Cross into New Jersey
04.0	1.0	Take the I-95N/Turnpoike North exit on the left
05.0	6.1	Merge onto I-95N/New Jersey Turnpike North towards Cars/Trucks-Buses (portions toll)
11.1	3.0	Keep left to take I-95N/New Jersey Turnpike toward I-280/Rt-3 Meadowlands Sports Complex/Exits 15W-18W George Washington Bridge (portions toll)
14.1	17.3	Merge onto I-280 W via Exit 15W toward Newark/Kearny (portions toll)
48.7	0.5	Take the I-80 W exit on the Left toward Delaware Water Gap
49.2	1.7	Merge onto I-80 Express Lane W
50.9	39.3	I-80 Express Lane W becomes I-80 W
90.2	0.1	Take RT-94 N exit 4A-B-C toward US-46 E/Columbia/Blairstown
90.3	0.2	Take Exit 4B on Left toward US-46 E/Columbia/Pa611/Portland Pa
90.5	0.1	Turn slight left onto NJ-94W
90.6	0.1	NJ-94 W becomes US-46 E
90.7	0.2	Take the ramp toward PA-611-Portland, Pa
90.9	0.2	Turn right onto 611 ALT. Crossing into Pennsylvania. View of Delaware Water Gap on right.
91.1	0.1	Toll booth. Why does it cost money to leave New Jersey, but it's free to get in?
91.2	0.2	Exit right toward Portland, to Pa Route 611.
91.4	0.5	Turn left at bottom of ramp onto Route 611 N. Entering Portland Pa.
91.9	0.6	Exposure of fluvio-glacial outwash in road cut on left.
92.5	0.9	Road (National Park Drive) to Slateford Farm on left. Enter Delaware Water Gap National Recreation Area

94.0 0.2 Arrow Island Overlook on right. Arrow Island Trail (Stop 4 on left).

94.2 0.2 Martinsburg outcrop on left, note low dip of slaty cleavage.

94.4 0.2 Cold Air Cave on left (Stop 3).

94.6 0.1 Turn left into Point of Gap Overlook parking lot. Disembark and proceed to south on the grassy slope.

STOP 1. POINT OF GAP OVERLOOK: GEOLOGIC SYNOPSIS STRUCTURE, STRATIGRAPHY, GEOMORPHOLOGY, GLACIAL GEOLOGY

Figure 5. Entrance to Delaware Water Gap National Recreation Area as viewed from atop Kittatinny Mountain at locality 18 of the Red dot-blue-dot-Dunnfield Creek trail. View looking southward; Pennsylvania is on the right, New Jersey on the left. The Delaware River flows through the constricted gap behind the view, and as it widens into the valley beyond and as its velocity decreased, it deposited a streamlined bar, Arrow Island. Between the mountain held up by quartzite of the Silurian Shawangunk Formation, and the Precambrian metamorphic rocks of the New Jersey Highlands in the distance, lies Paulins Kill Valley, underlain by Cambrian and Ordovician limestone and slate. Coarse gravels in a Wisconsinan outwash terrace lines both sides of the valley south of the gap.

The following questions are typically asked—or at least thought about—by laymen visitors to Delaware Water Gap:

What is a water gap?
What is the layering in the rocks?
What are the rocks made of?
Why are they of different colors?
Why are they tilted and curve in different directions?
Why is the cliff irregular and not smooth?
Why does the mountain top look flat from a distance?
Why did the Delaware River cut through the mountain here?
Did it cut through anywhere else?

Possible answers to these questions will be discussed with trip participants. But first, let us go through the jargon! The NPS has prepared a curriculum guide for school grades 3–6, "The Many Faces of Delaware Water Gap" (Ferrence et al., 2003). In that guide, the Point of Gap Overlook is discussed (see Appendix 1), and the guide includes exercises in how the rocks formed, how the mountains were built, and glaciation.

Delaware Water Gap is often cited as the classic water gap in the Appalachian Mountains. Figure 6 portrays its geology. Anyone who studies the area is compelled to contemplate the history of the gap and why it is where it is. Is the structure seen in the gap (Fig. 2) as simple as it looks? What story do the satellitic folds in the Shawangunk tell? Why does the cleavage dip to the northwest within several hundred feet of the contact with the Shawangunk? These issues were summarized in the last GSA trip to this area four years ago (Epstein, 2006) and may be discussed at the stop.

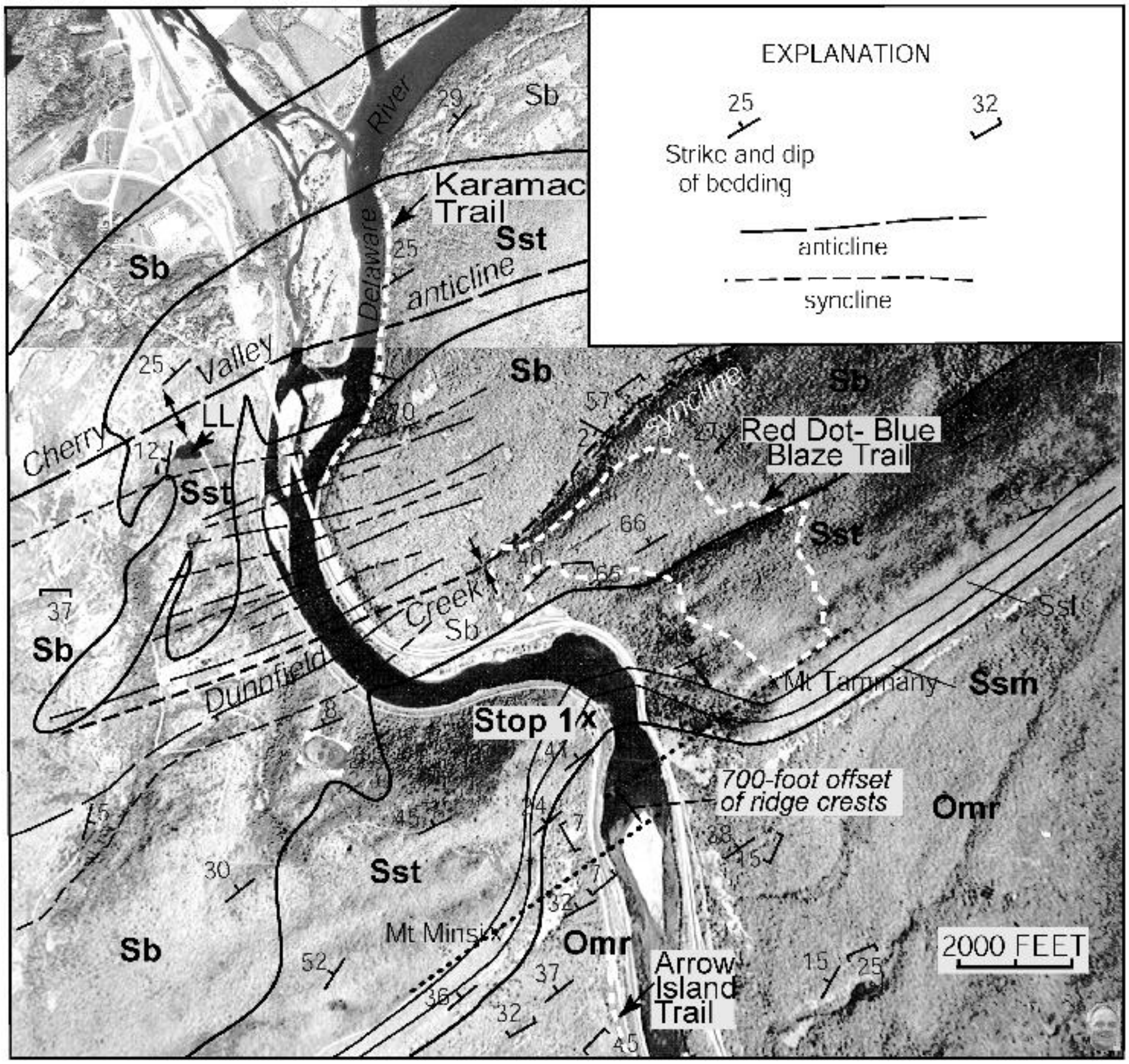


Figure 6. Aerial photograph and geologic map of Delaware Water Gap. Trails and stops are shown; LL, Lake Lenape. The 700-foot offset of the ridge crests (dotted line) on either side of Kittatinny Mountain is also shown and will be used in the discussion of the origin of the gap. Omr, Ramseyburg Member of the Martinsburg Formation; Shawangunk Formation: Ssm, Minsi Member; Ssl, Lizard Creek Member; Sst, Tammany Member; Sb, Bloomsburg Red Beds. A series of small anticlines and synclines lie between the Dunnfield Creek syncline and Cherry Valley anticline. Arrow Island is a streamlined bar that formed where the Delaware River emerges from the constricted portion of Delaware Water Gap. The unusual pattern of the Ss-Sb contact in the western area is due to the variable nature of the color boundary (Epstein, 1973).

Story of Delaware Water Gap, a Popular Version

The spectacular Delaware Water Gap has inspired people for generations, and created wonder on how this magnificent chasm through Kittatinny Mountain could have been cut by the Delaware River. Its story goes back in geologic time many hundreds of millions of years, although the actual cutting took place within the last few million years.

Our planet has had a dynamic history for much of its existence. Mountains have risen and fallen as continents have shifted and collided with each other during the earth's more than 4.5 billion year history in a process called *plate tectonics*. Kittatinny Mountain is part of the Appalachian Mountain chain that extends for more than 2000 miles from Maine to Georgia. The birth of the Appalachians dates back several hundred million years. About 550 million years ago during the Cambrian period, a carbonate bank (*A* in top of Fig. 7) lying on the shelf of the old North American continental crust (*B*) faced eastward toward an ancient ocean, Iapetus (*C*). Africa lay across the Iapetus Ocean far to the east. A volcanic island (*D*) formed in the middle of the ocean as the African continent began to drift westward, beginning the closure of Iapetus. Thick sediments were deposited in the basin ahead of the volcanic island arc that were later consolidated into the Martinsburg Formation. As the ocean basin continued to close during continued plate convergence, sediments were deformed (middle figure) and later, not shown in Figure 7, uplifted ~450 million years ago during the Ordovician period of geologic time, resulting in the buckling of the earth's crust and uplifting of the ancient Appalachian Mountains. This period of mountain building is called the *Taconic Orogeny*. Sands and pebbles from the rising mountains were shed to the northwest during Silurian time, deposited across the warps and folds in the underlying Martinsburg. These clastic rocks were to become the Shawangunk Formation that holds up the mountain at the gap. Other orogenies variously affected the rocks in the Delaware Water Gap area. During the Devonian period, the Acadian orogeny uplifted mountains to the southeast which resulted in deposition of thick sediments seen in the Pocono Mountains today, but it had little or no structural effects on the rocks here. The Alleghenian orogeny at the end of the Paleozoic era, probably beginning in the Pennsylvanian period, folded and faulted all the rocks in eastern Pennsylvania, and was responsible for the folds seen in the Gap. It was at this time that Africa finally collided with North America.

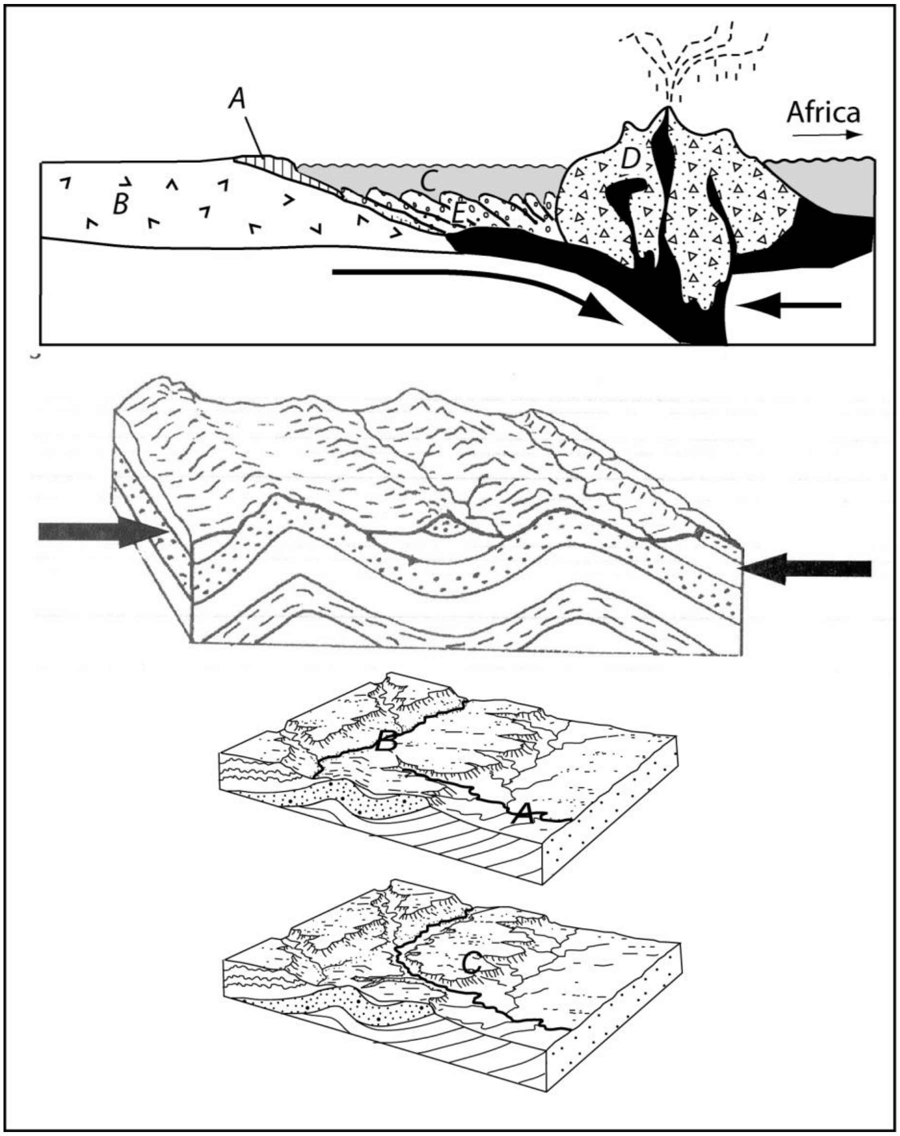


Figure 7. Formation of Delaware Water Gap, a story of plate tectonics, uplift, and erosion.

For many millions of years erosion has worn down these rocks–the harder rocks stand as mountains while the softer rocks were eroded to form valleys. The sandstones and conglomerates of the Shawangunk Formation (shown by the stippled pattern in Figure 7) are among the hardest rocks in the entire Appalachian Mountains and form some of the highest ridges, extending all the way south to Alabama.

How the Gap was Formed

Some millions of years ago, the headwaters of a southeastward flowing river that was still south and east of here (perhaps near present-day Trenton, New Jersey) eroded backward toward the Kittatinny Ridge (*A* in lower Fig. 7). The hard rocks of Kittatinny Mountain presented an obstacle to erosion, and the Delaware found a place in the ridge where the rocks are more highly fractured and less resistant to the erosive power of the river. Thus, the river eventually worked its way *headward* through the mountain (*A*), “capturing” waters on the other side (*B*) and establishing the present course of the Delaware through Kittatinny Mountain (*C*). It carved it way through the rocks that make up the Pocono Mountains to the north. Carrying sand and pebbles, which eventually wind up in the Atlantic Ocean, the Delaware River continues to downcut and widen what was originally a small cleft in the rock. That cleft, over time, has been enlarged into today’s mile-wide chasm: The Delaware Water Gap.

At this stop we may discuss the several intricacies of structural geology in this area, such as origin of slaty cleavage, pressure-shadow mechanism affecting cleavage, and ages of deformation. Maybe not!

94.7 0.7 Leave Point of Gap parking lot, turning left (north) on US 611.

95.5 0.8 Contact between the Shawangunk and Bloomsburg dipping 35°NW.

96.3 0.1 Southeast-dipping rocks in the Bloomsburg Red Beds. US 611 traverses the Bloomsburg for the next 0.8 miles in a series of small undulating, low-amplitude folds. Note the southeast-dipping cleavage.

96.4 0.1 Crest of the Cherry Valley anticline in the Shawangunk Formation at top of road. The contact between the Shawangunk Formation and Bloomsburg Red Beds is conventionally placed at the base of the lowest red bed. However, at this locality this color change migrates up and down section by as much as 700 feet, making for a peculiar map pattern (Epstein, 1973). Enter upstream side of Delaware Water Gap.

96.5 0.1 Northwest-dipping rocks of the Shawangunk Formation on left. Turn left on Mountain Road just as you enter the town of Delaware Water Gap.

96.6 0.1 Turn left onto Lake Road to parking area.

STOP 2. LAKE LENAPE

We will follow the Appalachian Trail along an abandoned blacktop road that once led to the top of Mount Minsi to a fire tower. At the north edge of the pond is a memorial to Sean Dolan, A Delaware Water Gap fire fighter who died in a nearby car accident in 2004.

Lake Lenape is near the trough of one of many small folds in Delaware Water Gap (Fig. 8). It is in an area where the contact between the Shawangunk Formation and Bloomsburg Red Beds is based on change from gray in the Shawangunk to reddish in the Bloomsburg. In this immediate area, this color change variably cuts across ~700 feet of stratigraphic thickness, making for the complex boundary seen on the map.

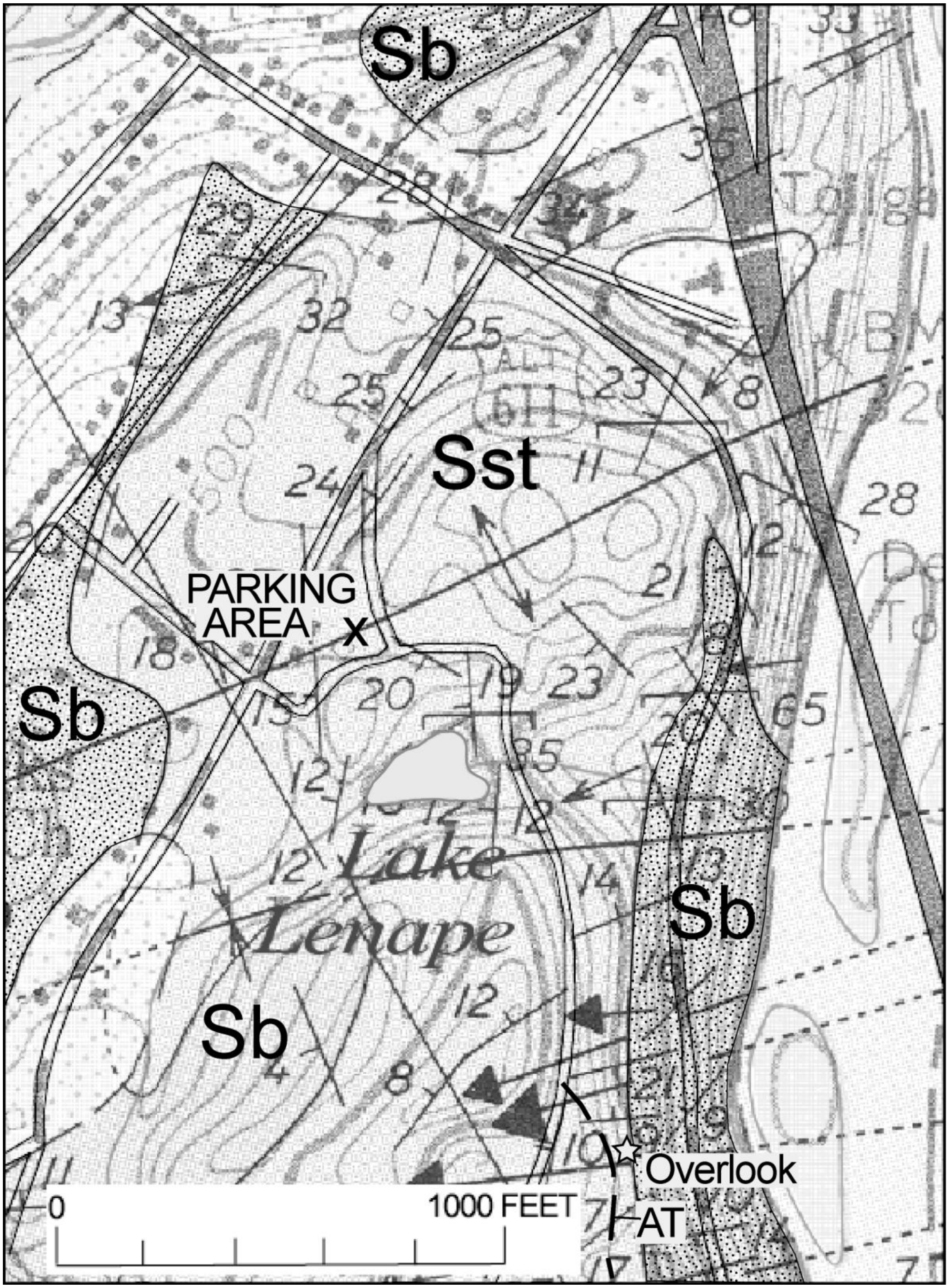


Figure 8. Geologic map of the Lake Lenape area (from Epstein, 1973). Sst, Tammany Member of the Shawangunk Formation; Sb, Bloomsburg Red Beds; AT, Appalachian Trail. Glacial deposits and alluvium in areas beyond bedrock exposures.

Lake Lenape is a marvelous area to give students an opportunity to examine several rock types, including sandstones, siltstones, and shales, to determine differences between bedding and rock cleavage, and to make a crude clinometer to prepare a map of the local gentle syncline.

The curriculum guide handout (Appendix 2) will provide instructions on how to use a compass, prepare grain size charts in order to determine sedimentary rock type, and encourage students to examine the local biota. Be wary of poison ivy, which covers the southernmost outcrop that will be measured. Here, have the students eyeball the dip with the clinometer rather than placing it on the rock. There is also an exercise that asks the geologic question, "How Do Lakes Form?"

"The Many Faces of Delaware Water Gap" is a 125-page geology curriculum (grades 3–6) available from the Delaware Water Gap National Recreation Area. It includes field trips for five locations in the park in Pennsylvania and New Jersey, as well as a 20-page introduction to general geologic topics for the park. Teachers are invited to contact the park for a copy (http://www.nps.gov/dewa/forteachers/curriculummaterials.htm).

An interesting exercise would be to have the students walk from the lowest point along the road for a little more than 100 feet up the trail to the south where the outlet is located, ~4 feet higher in altitude. Then ask them these questions: Why is the outlet not at the lowest point along the road? Did the lake formerly drain at the lowest point? (Have them look to the east of the road to determine where the most gully erosion occurred.)

Point out the location of rhododendrons–on the north slopes in the shade and on acid, organic-rich soil.

There is a splendid overlook and view of the Delaware Water Gap by following the Appalachian trail for a short distance where the trail leaves the road ~600 feet south of the pond (see Fig. 8). From here you can discuss some of the folds in the Bloomsburg Red Beds across the river.

Leave Lake Lenape parking lot. Turn right on Mountain Road

96.7 2.3 Turn right on US 611, retracing route through Delaware Water Gap.

99.0 Pull off to side of road along parking area.

STOP 3. COLD AIR CAVE

This *talus* cave (Fig. 9A) was formed by juxtaposed alignment of large talus boulders derived from conglomerate and sandstone blocks of the Shawangunk Formation above. Some of these blocks are nearly 30 feet long. A large talus floe overlies the cave (Fig. 9B) and a large cleft in the cliff above (Fig. 9C) confirms the potential for generating these blocks during periods of freeze and thaw. The cave may be as long as 70 feet, although only ~30 feet of the cave is currently accessible to normal-sized individuals. According to Snyder (1989), the cave was discovered ca. 1870 when very cold temperatures, approaching 30 °F, were reported coming from the opening. The cave became an attraction sometime thereafter with a building erected at the entrance (Fig. 9D); tours given. Beginning in September 2000, a cooperative effort to understand the phenomenon of Cold Air Cave was undertaken by the U.S. Geological Survey and National Park Service. Readings for temperature, wind speed, wind direction, and weather conditions were recorded. The data indicates significant fluctuations in cave temperature; they are warmer in summer and cooler in winter. There are significant differences in temperature depending on whether the air is blowing into or out of the cave. Air blowing out is cooler than when it is blowing in the reverse direction, indicating that there is storage of cold air in the scree system and possibly the bedrock above. The average temperature in the parking lot is only slightly cooler than underground temperatures found at this latitude—normal is 54–56 °F. The average temperature in the cave is surprising, however, and was unexpected. In conclusion, Cold Air Cave continues to attract and mystify people with its local folklore, blowing cold air, and mysterious scientific explanation.

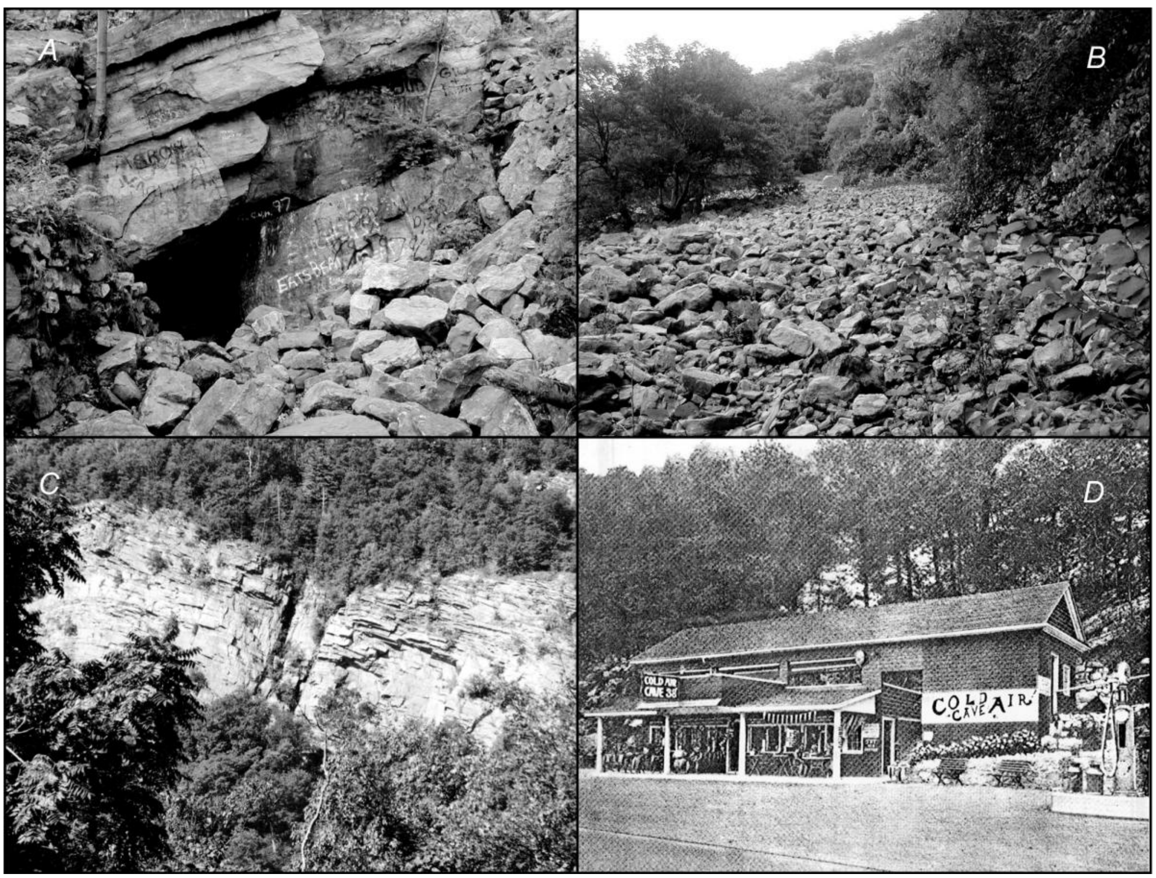


Figure 9 ***A***. "Cold Air Cave", a talus cave formed by conglomerate and sandstone blocks of the Shawangunk Formation along US 611 south of Delaware Water Gap. ***B***. Talus above Cold Air Cave. ***C***. Cleft in cliff of the Shawangunk Formation above the Cold Air Cave, the possible source for the large blocks that supplied the talus. ***D***. Concession stand at Cold Air Cave, circa 1940.

NOTE: Since the time of the preparation for this stop, this cave and others throughout DEWA, have been closed to the public because of a virulent fungus, "white nose syndrome", that threatens the local bat population (http://www.nps.gov/dewa/parknews/upload/newsBATCAVES.pdf).

Leave Cold Air Cave continuing south on US 611.

99.2 0.2 Small outcrop in the Martinsburg Formation on right (Figure 10). Note the gently northwest-dipping (5°) cleavage due to fanning in a pressure shadow of the syncline defined by the Shawangunk Formation that will be discussed at Stop 1. Bedding here dips 25° NW.

Figure 10. Outcrop of the Martinsburg Formation at mileage 99.2 along US Route 611. Bedding (solid line) dips 24^{o} northward, typical of the general homoclinal dip in this area. Cleavage (dashed line) dips ~ 5^{o} northward as part of a broad cleavage arch as the Shawangunk contact is approached.

99.4 0.4 Turn right into parking area to Stop 4.

STOP 4. ARROW ISLAND OVERLOOK AND TRAIL.

Arrow Island, a streamlined bar, is presently being modified by the Delaware River. This occurs due to a decrease in the velocity of the river as it emerges from the narrow confines of the gap to the north. Such a feature has been termed an *expansion bar*.

The trail is about one mile long (Figs 11 and 12), passing over glacial deposits and outcrops of slate and greywacke in the Martinsburg Formation, ending at a kettle hole pond near National Park Drive. There are several abandoned slate quarries and prospects along the trail, one of which we hope to visit.

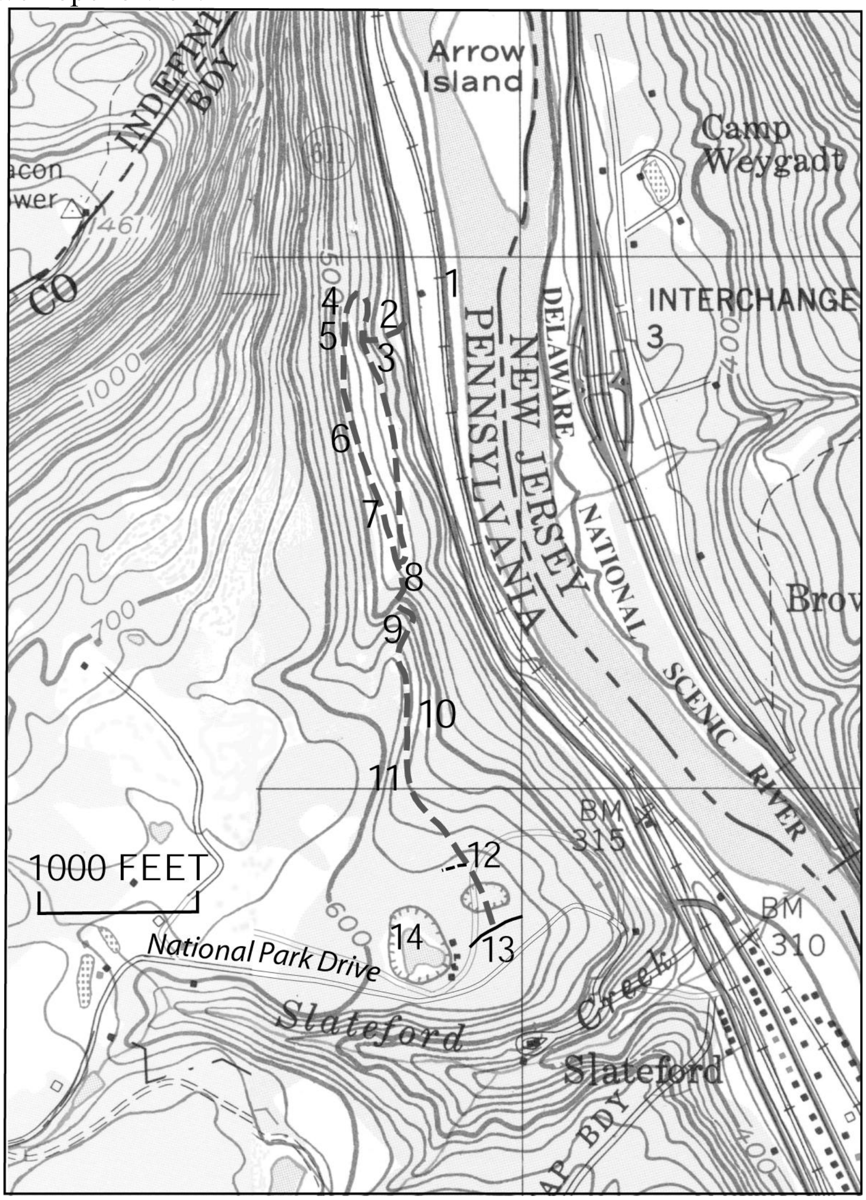


Figure 11. Map of the Arrow Island Trail

Welcome to the Arrow Island Trail.

Stop 2. Angular boulders (talus) that came off the cliff above. Compare to rounded glacial boulders nearby.

Stop 3. Slate dump of waste slate at top of trail.

Stop 3. Foundation remnants below slate quarry.

Stop 5. Slate quarry. Bedding and cleavage can be seen at arrow.

Stop 5. Original horizontal sediment layer (bedding; solid line) is now tilted. The rock breaks along cleavage (dashed line).

Stop 8. Large erratic (glacial) boulder of siltstone beneath leaves.

Stop 8. Rounded and polished glacial boulder north of creek.

Figure 12. Description of stops along the Arrow Island Trail

99.8 0.2 Very coarse gravel in late Wisconsinan outwash terrace to right exceeds 80 feet in thickness. The gravel was either laid down between the valley's wall and stagnant ice forming a kame terrace, or it is the eroded remnant of a valley train that formerly filled the Delaware Valley.

100.0 0.3 Rock fence on right is composed of blocks of the Allentown Dolomite with abundant fine sedimentary structures.

100.3 0.1 Cross Slateford Creek. In 1805, the Pennsylvania Slate Company developed a slate quarry south of the water gap near Slateford Creek. It is abandoned.

100.4 0.1 National Park Drive on right leads to Slateford Farm, an example of a National Historic Site maintained by the National Park Service.

100.5 0.5 Faulted and overturned beds in the Bushkill Member of the Martinsburg Formation in ravine to right.

101.0 0.4 Flat-lying slate in the Bushkill Member in a 100-foot-long abandoned quarry overlain by 10 feet of glacial drift in ravine to right. A dolomite concretion, characteristic of basal beds of the Martinsburg elsewhere, lies in the bottom of the quarry.

101.4 0.3 Crossing the concealed Portland fault, which juxtaposes Martinsburg against the Jacksonburg Limestone and rocks of the Beekmantown Group where it is exposed.

101.7 0.1 Enter Portland, Pennsylvania.

101.8 6.0 Traffic light. Continue straight.

101.85 0.05 Cross Jacoby Creek. Deglaciation of the Jacoby Creek valley resulted in the formation of several proglacial lakes that became progressively lower as the ice retreated from the northeast-draining valley and lower lake outlets were uncovered (Ridge, 1983).

101.9 0.15 Road passes underneath US 46.

102.0 0.1 Turn right following signs toward 1-80 and Portland Toll Bridge.

102.2 0.2 Toll Booth. (No toll leaving Pennsylvania.)

102.4 0.2 Cross Delaware River into New Jersey. Good view of Delaware Water Gap to left.

102.6 0.2 Allentown Dolomite crops out on right. Continue straight.

102.8 0.1 Road signs for 1-80 and NJ 94. Continue straight.

102.9 0.2 Cross the axis of the Ackerman anticline (Drake and others, 1969) in the Allentown Dolomite approximately at point where ramp bears off to 1-80 West. Merge left onto I-80 West.

103.1 1.8 Coarse kame-terrace (outwash) deposits on right.

104.9 1.5 Unconformable contact (< 5^o, Beerbower, 1956) between the Martinsburg Formation (Ordovician) and Shawangunk Formation (Silurian) covered by talus on right. If you are quick, lucky, and can avoid a car crash, look up about 50 feet and see mudcracks in an overhanging shale in the Shawangunk. (Fig. 13).

Figure 13. Mudcracks in shale interbedded with crossbedded sandstone of the Tammany Member of the Shawangunk Formation about 50 feet above I-80 at the entrance to Delaware Water Gap. The shales are interpreted to be overbank deposits that were dessicated in a fluvial environment

106.4 0.1 Bear right onto service road. Park in parking area for Stop 5.

If we decide to have lunch here, instead of parking, continue straight and turn left under I-80 towards Delaware Water Gap visitor's center, *Jiffy John* conveniences, and picnic tables.

After lunch, return to parking area at mileage 106.5 for Stop 5.

STOP 5. GEOLOGIC FEATURES ALONG THE RED DOT-BLUE BLAZE-DUNNFIELD CREEK TRAILS

The Red Dot–Blue Blaze–Dunnfield Creek Trail circuit takes the hiker to the top of Mount Tammany in Kittatinny Mountain, New Jersey (Fig. 14), and in less than four miles, passes through many geologic phenomena, including a variety of rock types, landforms and glacial and structural features. The points of interest along the way (Fig. 15; Table 2), provide insight into these natural elements that influenced the formation, history and composition of Delaware Water Gap. Note that the sedimentary layers that make up the cliffs in the main part of the gap dip to the left, but in Dunnfield Creek valley to the left the beds are horizontal. Trip time is ~4 hours.

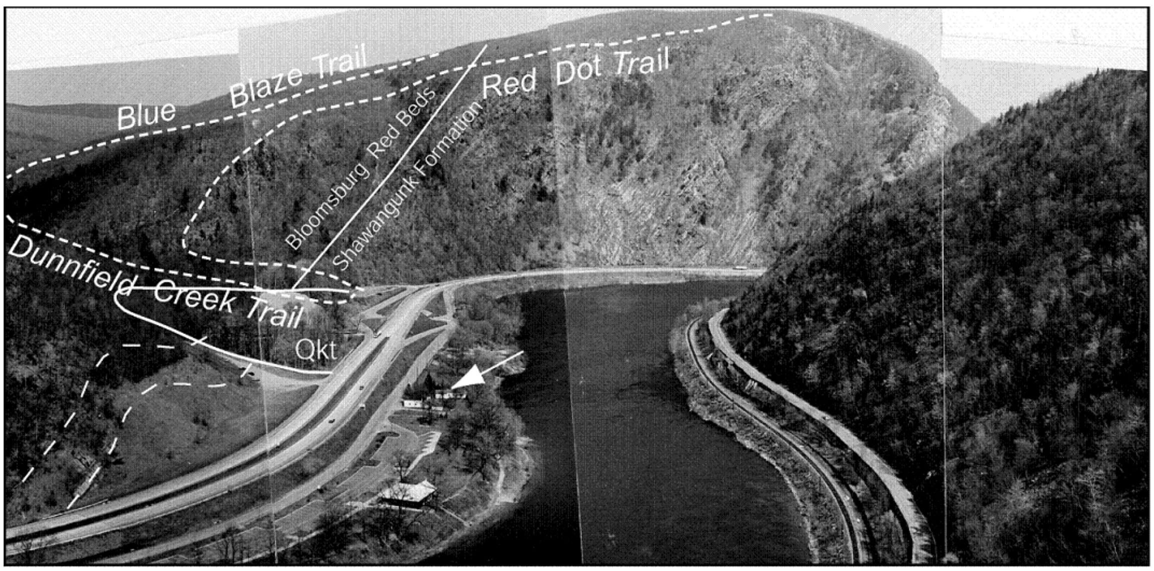


Figure 14. View of the trails at Stop 5. The Delaware River makes a sweeping bend as it heads through the world-famous Delaware Water Gap. The bend of the river mimics the underlying geology. The Shawangunk Formation of Silurian age, comprising very hard quartzite and conglomerate and holding up Kittatinny Mountain, dips moderately northward and is overlain by finer clastics of the Bloomsburg Red Beds, also Silurian in age. The dip flattens out under Dunnfield Creek beyond which the Bloomsburg is thrown into a series of small folds overlying a broader fold in the buried Shawangunk. The bend in the river mimics the form of that anticlinal fold. Approximately 20,000 years ago, Wisconsinan glacial ice occupied this valley and a kame terrace (Qkt) of sand and gravel was deposited along Dunnfield Creek. The National Park Service visitor's center is located at the arrow.

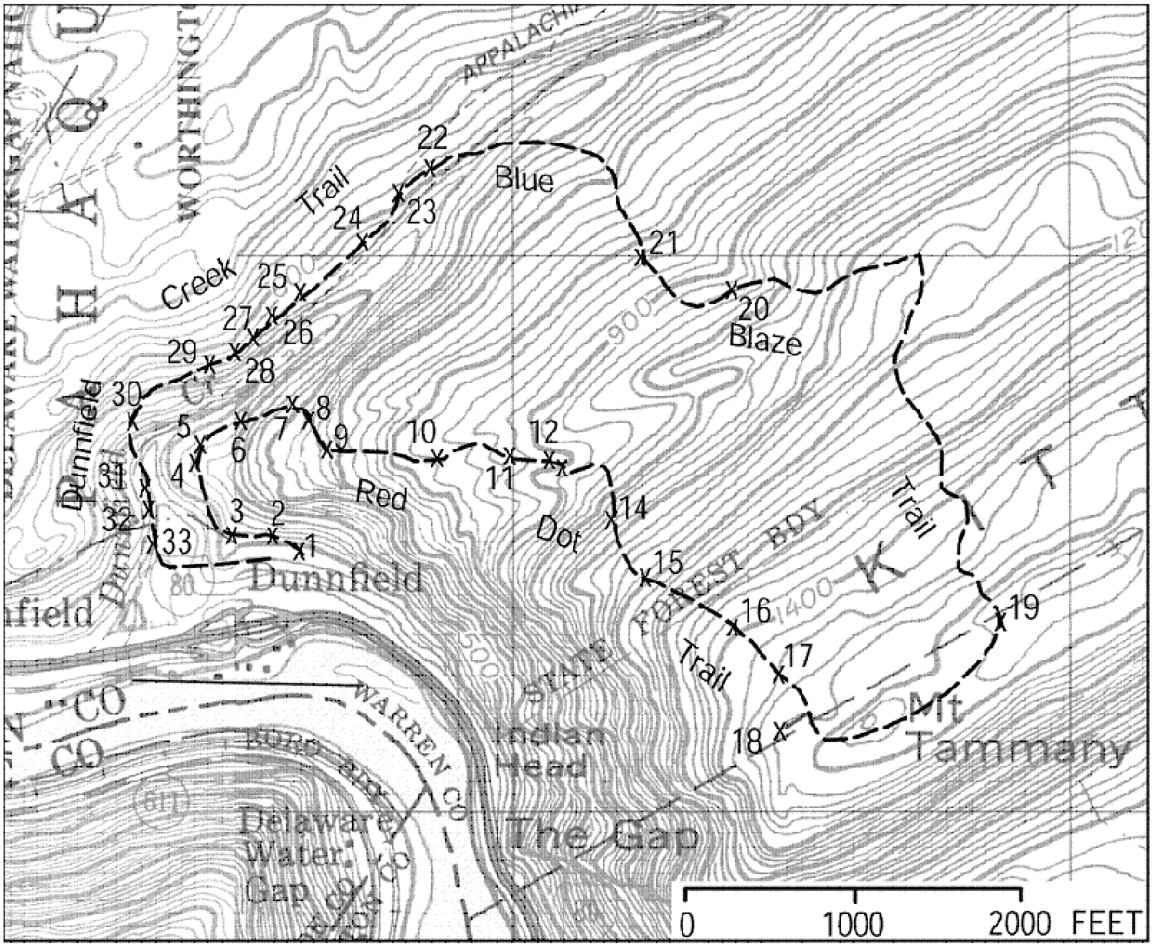


Figure 15. Trail map at Stop 5 showing places of geologic interest and are keyed to the descriptions and pictures shown on the accompanying table. A colored version of this map will be handed out to the participants.

Table 2. Localities of geologic interest along the Red Dot-Blue Dot-Dunnfield Creek trails at Stop 5.

1: Near contact between the Shawangunk Formation and Bloomsburg Red Beds.
2: Eight-foot-long boulder with slickensides.
3: Glacial kame terrace on silt, sand and gravel.
4: Glacial striae.
5: Rotted limestone glacial erratic.
6: Rib of Bloomsburg bedrock.
7: Series of greenish-gray and red siltstone, sandstone and shale of the Bloomsburg.
8: Large erratic, Schoharie Formation.
9: Overlook of the Delaware Water Gap.
10: Red sandstone and siltstone of the Bloomsburg have been polished by the

last glacier (20,000 years ago), producing glacial striae.
11: Springs.
12: Beginning of Shawangunk Formation on steep slope.
13: Talus.
14: Rib of quartzite with joints.
15: Glacial cobbles and glacial striae on Shawangunk.
16: Gentle slope underlain by some shale.
17: Forest fire.
18: Overlook, many sedimentary structures in the Shawangunk.
19: Blue Blaze Trail descends slope through laurel and blueberries.
20: Exposure of Bloomsburg bedrock with glacial striae.
21: Soil erosion by boots and rain exposing Bloomsburg bedrock with glacial striae.
22: Erosion has removed about three feet of glacial till.
23: Dunnfield Creek falls over flat beds of the Bloomsburg in bottom of syncline.
24: Three large erratic boulders of slightly cherty limestone.
25: Intersection with Appalachian trail. Several more boulders in creek.
26: Plunge pool formed where the creek drops over hard sandstone and gouges out a rounded pool in softer shale below.
27: The creek erodes along a joint surface here forming a 30-foot sluiceway.
28: Large boulders fallen from adjacent Bloomsburg outcrop have sharp edges compared to the rounded and eroded edges of erratics.
29: Cleavage present in horizontal shale layers but not in sandstone. Erratic in creek.
30: Beginning of the terrace deposit that was first seen at locality 3.
31: 25-foot-long limestone erratic limestone.
32: Bridge. Flat alluvial fan towards parking lot made up of rounded cobbles.
33: Parking lot. Note 6-foot boulder 40 feet above the creek in terrace to right.

GLOSSARY

Alluvial fan: Gently sloping mass of sediment fanning out from a river mouth.

Cleavage: Closely spaced fractures along which a rock may split.

Erratic: A rock that was carried some distance by a glacier from its place of origin.

Kame terrace: Flat-topped hill formed from sediment that was deposited along a valley wall by streams that flowed from an adjacent melting glacier.

Slickensides: Polished and striated rock surface caused by one rock mass sliding past another.

Striae: Narrow parallel scratches cut into a rock surface by rock debris embedded in the bottom of a moving glacier.

Syncline: U-shaped downfold of rock layers.

Talus: An apron of irregular rock fragments derived from and lying at the base of a cliff.
Till: Unsorted mixture of clay, sand, and boulders deposited beneath a glacier.

106.5 0.7 Retrace route out of parking area, turning right along service road, carefully merging onto I-80 west.

107.2 0.1 Turn right at Exit 1 toward Millbrook/Flatbrookville. At the ramp turnoff from I-80 to Old Mine Road there are splendid folds in the Bloomsburg Redbeds showing the relations of cleavage and bedding (fig. 16). The southeast dip of cleavage in all limbs of the folds indicate that it was rotated by bedding slip after they formed.

Figure 16. Panoramic view of folds in alternating sandstone, siltstone, and shale in the Bloomsburg Redbeds showing prominent southeast-dipping cleavage.

107.3 0.1 Turn right at Stop sign onto Old Mine Road.

107.4 1.0 Pull off onto parking area to left before traffic light. Stop 6.

STOP 6. GUIDE TO THE GEOLOGY OF THE KARAMAC TRAIL

The trail begins in the parking area at the traffic light and barrier along the Old Mine Road 400 feet north of the I-80 overpass (Fig. 17). This a level trail, nearly one mile long, along an old railroad grade, and along which a few different types of rocks (sandstone, shale, siltstone and conglomerate), two different geologic formations of Silurian age (Bloomsburg Red Beds and Shawangunk Formation), a landslide retained by a concrete structure (gabion), a few folds, joints, rock cleavage, and glacial boulders (erratics) can be seen. Terms in italics should be fully explained to students. The distance along the trail north from the parking area is shown in feet.

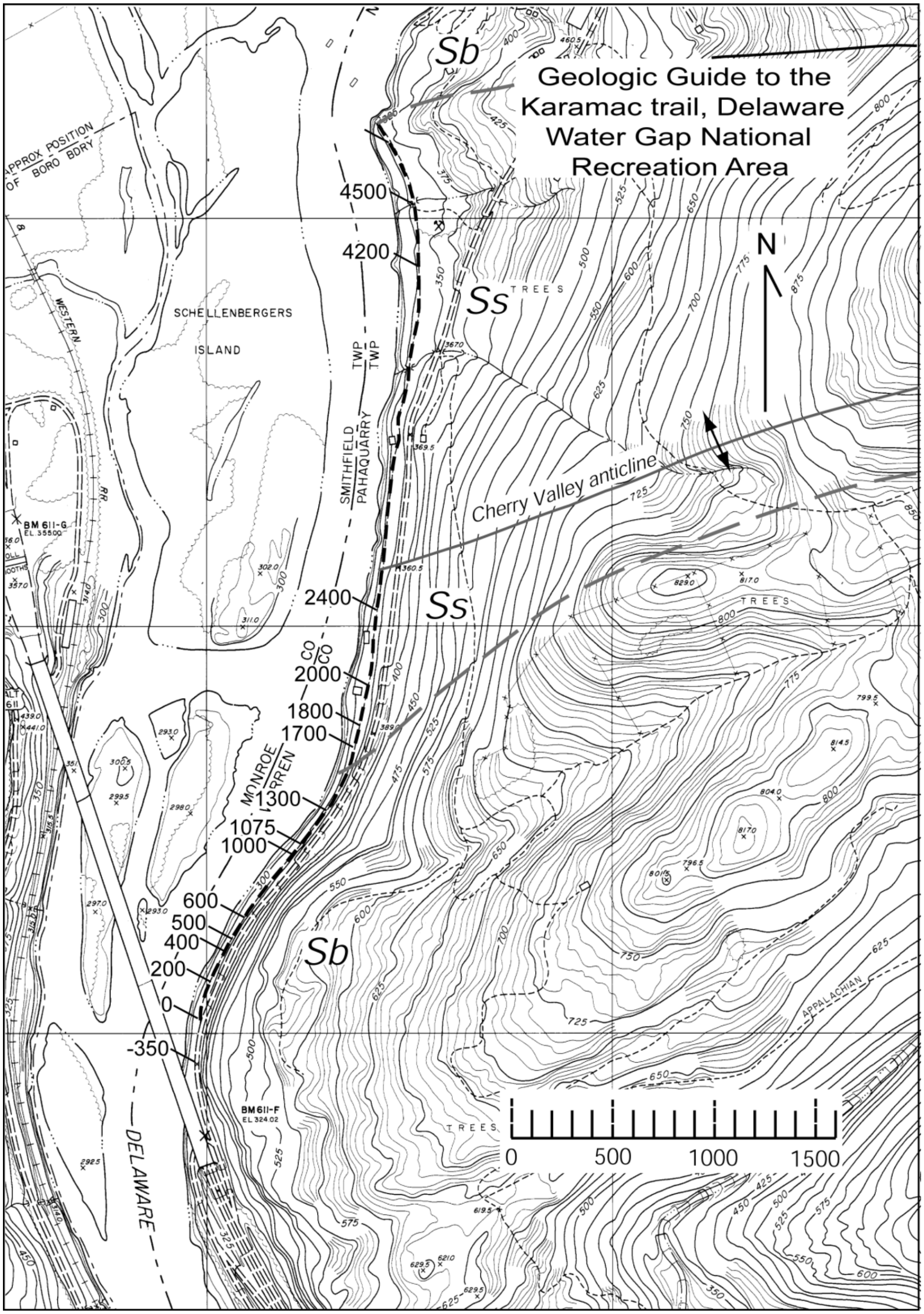


Figure 17. Detailed topographic map (contour interval: 5 feet) of the Karamac trail and distances from the parking area, keyed to descriptions in the text. *Ss*, Shawangunk Formation; *Sb*, Bloomsburg Red Beds. Consecutive figure numbers are for this guide only. A few sections of the trail are overgrown, but access to outcrops is generally good.

LOCALITY DESCRIPTION

-350 On the east side of the road is an excellent example of a small *thrust fault* and *ramp* in the Bloomsburg Red Beds (fig. 18). The floor of the thrust is at the base of a thick sandstone bed lying on top of shale and siltstone that are less resistant to erosion.

Figure 18. Bedding thrust and ramp at base of sandstone in the Bloomsburg Red Beds along Old Mine Road in New Jersey.

0 Parking area. Looking to the right up the steep slope are *joints*, the result of the tectonic stress that folded the rocks in the Delaware Water Gap area. Similar vertical fractures can be seen along the trail for the next 400 feet.

400 The smooth face in a 6-foot high bed of sandstone ahead is an excellent example of a *joint* (fig. 19). Note the cleavage in the silty beds at arrow.

Figure 19. Smooth joint surfaces in sandstones of the Bloomsburg Red Beds and cleavage in siltstone bed.

500 The closely spaced fractures in some of the greenish and reddish beds are tilted into the hill in a much gentler angle than are the joints (fig. 20). This is *cleavage* which forms in shale and siltstone by reorientation of the mineral grains in the rock.

Figure 20. Cleavage (dashed line) in shale and siltstone dipping more steeply than bedding (solid line). Be careful of the plant with “leaves of three …let it be”.

600 The interlaced segments of concrete is a 100-foot long *gabion* (fig. 21) which was constructed at this site to hold back a landslide that removed part of the Old Mine Road above the trail. Because of the slide the road is now a single lane with a three-minute traffic light to control traffic along this stretch, and it is unwise to walk along it. The concrete blocks are filled with cobbles to stabilize the road. Immediately above this site along the Old Mine Road there is the potential for another rockfall because blocks of rock are being separated along irregular joints (fig. 22).

Figure 21. Concrete gabion along abandoned railroad grade below the Old Mine Road in Worthington State Park

Figure 22. Irregular joints, trending along the Old Mine Road above the Karamac Trail. Boulders in the fractures are wedging the block apart, creating the potential for toppling.

1000 The islands in the middle of the Delaware River and terraces along its banks are made of sand that was deposited when the river flowed at a slightly higher level after retreat of the *Wisconsinan glacier* which left this area ~20,000 years ago. The sand islands are mute testimony to a former higher level of the river.

1075 Rivulet tumbling down over dark red rocks. This is the *Bloomsburg Red Beds.* Note that the rock layers are nearly flat at this spot. These rocks were deposited by rivers flowing to the sea. The soft sediment, composed of mud and sand, was hardened or *lithified* into rock and the layers are termed *bedding*. The process of *lithification* results from a variety of causes, including *cementation* (the rock is cemented by minerals that precipitated out from ground water) and *compaction* (due to the weight of thousands of feet of overlying sediments that were deposited on top).

1300 Water emitting from a drain pipe. Just beyond a rivulet the bedding is tilted much more steeply to the south than previously (fig. 23), showing that rocks have been *folded*. The rocks here are no longer red, but shades of gray. Thus, we have left the Bloomsburg Red Beds and are into another rock unit, the *Shawangunk Formation*. It would be worthwhile to discuss the fact that the Bloomsburg overlies and is younger than the Shawangunk based on the dip of the beds here (*law of superposition*).

Figure 23. Beds in the Shawangunk Formation dip ~ 40^{o} to the southeast

1700 Outcrop of interbedded sandstone and some conglomerate and siltstone. Some of the sandstones have *cross bedding* (see fig. 25). Beds dip abut 35 degrees to southwest. The *sediment* which was compacted into the hard sandstone formed by rapidly flowing streams that came off mountains that existed to the south of us ~ 425 million years ago.

1800 This small *anticline* (fig. 24) interrupts the general south dipping layers along this stretch of the trail.

Figure 24. Anticline in Shawangunk Formation. The northward-dipping beds are fairly steep.

2000 Note that the bedding layers are tilted back towards the south, indicating that there is a *syncline* between here and the previously seen northward dipping beds. This is the site of a *spring*, where water, flowing through cracks in the underground rock (termed *ground water*), appears at the surface. During the summer when rainfall is less plentiful, the spring may dry up.

2400 Bedding in the Shawangunk Formation *dips* (is tilted) gently to the south (fig. 25). Farther north along the trail and to the end of the trail it will be dipping northward, a structural configuration which is termed an upfold or *anticline*, named the *Cherry Valley anticline* because it is a major anticline that extends for several miles to the northeast and southwest. This outcrop is also a good example of *crossbedding*, which is defined by layers of rock that dip more steeply than the beds above and below and is caused by large *ripples* that formed in a rapidly moving stream during the Silurian Period.

Figure 25. Crossbedding in the Shawangunk Formation. The bedding dips fairly gently to the southeast (solid line), whereas the cross beds (dashed line) dip more steeply to the south. The crossbeds were formed by large ripples in a stream that flowed from left to right

3400 100-foot long *outcrop* through the Shawangunk Formation (fig. 26). A close look at the different beds show that some have grains the size of sand (sandstone), in others the grains are very fine (shale and siltstone), and some of the beds contain pebbles, perhaps ½-inch long, in a rock termed *conglomerate*. The beds here are tilted ~20^{o} toward the north.

Figure 26. Moderately northwest-dipping beds in the Shawangunk Formation. There is an *anticlinal axis* between here and the rocks in figure 25.

4200 The gray rocks of the Shawangunk Formation here are more steeply dipping (33°) than to the south.

4500 So far all the loose boulders we have seen are quite angular in shape, having been derived by breaking off from the adjacent bedrock. Notice the large rounded boulders here. That shape implies that they have been water worn. But water needs to be very rapidly flowing to move boulders of that size. Because these rounded boulders are made up of a variety of rock types, some of which are foreign to the immediate neighborhood, it implies that they have been carried here by *glaciers* which invaded this area ~20,000 years ago. Now, fortunately, the glaciers are gone. Could they return some time in our geologic future? This would be a good place to discuss *climate change*, both natural and anthropogenic. These boulders, because they have come from afar, perhaps several miles away, are termed *erratics*. In the ground underneath they are mixed with clay, a geologic deposit called *till*. Till forms much of the surface further northward along the trail.

4850 The rocks of the Shawangunk Formation are cut through here along this old railroad grade (fig.27). The beds are tilted more steeply than before; 62° at this spot. This is a good outcrop to show the different lithologies in the Shawangunk Formation and the cleavage in the shales that dip 85° northwest. It is also a good outcrop to illustrate the various orientations of joints. It would also be an excellent locality to demonstrate the use of a *Brunton compass* to acquire structural orientations and also to demonstrate the orientation of cleavage in folds (fig. 28) and also to discuss the relationship of cleavage to overturned beds in a fold.

Figure 27. Steeply dipping beds in the Shawangunk Formation. Various joint orientations have developed in the sandstones and cleavage has formed in the shales. This would be a good outcrop to discuss the use of cleavage and bedding orientations to determine the surrounding structure (fig. 28).

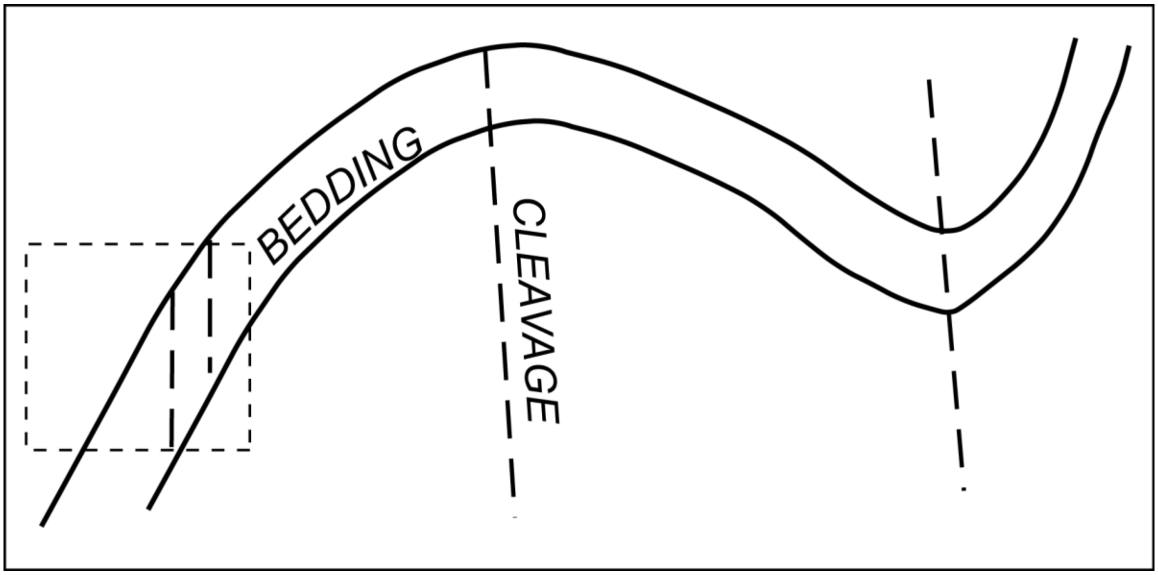


Figure 28. Using bedding-cleavage relations to show that where bedding is gentler than cleavage, the beds are right-side-up and that there should be an anticlinal axis to the right. Dashed box shows extent of outcrop in Figure 27.

4950 Trail ends at bridge abutment, part of the Lackawanna Railroad bridge that was removed during the flood of 1955 (Fig. 29). Excellent blocks of *conglomerate* at the abutment.

Figure 29. End of the Karamac trail. The bridge for the Lackawanna Railroad was taken out by the flood in 1955. That flood was one of the reasons for proposing the construction of the ill-fated Tocks Island dam about xxx miles upstream. Note the gravel bars in Delaware River and the higher stream terraces. The hill beyond is held up by various Upper Silurian and Lower Devonian sedimentary rocks.

		Turn around and leave parking area heading south on River Road towards the Delaware Water Gap visitor's center.
108.4	76.0	Pass the DEWA Visitor's Center (or stop for respite) and continue straight
108.5	184.4	Carefully merge onto I-80E. Retrace route back to CUNY College of Staten Island

END OF TRIP-HAVE A SAFE JOURNEY HOME.

REFERENCES CITED

Beerbower, J.R., 1956, The Ordovician-Silurian contact, Delaware Water Gap, New Jersey: Pennsylvania Academy of Science Proceedings, v. 30, p. 146-149.

Davis, W.M., 1889, The rivers and valleys of Pennsylvania: National Geographic Magazine, p. 183-253.

Drake, A.A., Jr., and Epstein, J.B., 1967, The Martinsburg Formation (Middle and Late Ordovician) in the Delaware Valley, Pennsylvania-New Jersey: U.S. Geological Survey Bulletin 1244-H, p. H2-H16.

Drake, A. A., Jr., Epstein, J. B., and Aaron, J. M., 1969, Geologic map and sections of parts of the Portland and Belvidere quadrangles, New Jersey-Pennsylvania: U.S. Geological Survey Miscellaneous Geologic Investigations Map 1-552, scale 1:24,000.

Epstein, J.B., 1966, Structural control of wind gaps and water gaps and of stream capture in the Stroudsburg area, Pennsylvania and New Jersey, in Geological Survey Research, 1966: U.S. Geological Survey Professional Paper 550-B, p. B80-B86.

Epstein, J.B., 1969, Surficial geology of the Stroudsburg quadrangle, Pennsylvania-New Jersey: Pennsylvania Geological Survey Bulletin G-57, 67 p.

Epstein, J.B., 1973, Geologic map of the Stroudsburg quadrangle, Pennsylvania-New Jersey: U.S. Geological Survey Quadrangle Map GQ-1047. Scale, 1:24,000.

Epstein, J.B., 1974a, Map showing slate quarries and dumps in the Stroudsburg quadrangle, PA-NJ, with a description of their environmental significance: U.S. Geological Survey Miscellaneous Field Studies Map MF-578. Includes interpretive text.

Epstein, J.B., 1974b, Metamorphic origin of slaty cleavage in eastern Pennsylvania [abs.]: Geological Society of America Abstracts with Programs, v. 6, Miami Beach, Fla., p. 724.

Epstein, J.B., 1989, Geologic map of Cherry and Godfrey Ridges, Saylorsburg, Stroudsburg, and East Stroudsburg quadrangles, Pennsylvania: U.S. Geological Survey Miscellaneous Geologic Investigations Map 1-1422.

Epstein, J.B., 2006, Geology of Delaware Water Gap National Recreation Area New Jersey-Pennsylvania: *in* Pazzaglia, F.G., ed., Excursions in Geology and history: Field trips in the Middle Atlantic states: Geological Society of America, Field Guide 8, p. 47-63.

Epstein, J.B., 2010, Teachers guide to geologic trails in Delaware Water Gap National Recreation Area, Pennsylvania–New Jersey, *in* Fleeger, G.M., and Whitmeyer, S.J., eds., The Mid-Atlantic Shore to the Appalachian Highlands: Field Trip Guidebook for the 2010 Joint Meeting of the Northeastern and Southeastern GSA Sections: Geological Society of America Field Guide 16, p. 127–147

Epstein, J.B., and Epstein, A.G., 1967, Field conference of Pennsylvania geologists, 32nd Geology in the region of the Delaware to Lehigh Water Gaps: Harrisburg, Pa., Pennsylvania Geological Survey, 89 p.

Epstein, J.B., and Epstein, A.G., 1969, Geology of the Valley and Ridge province between Delaware Water Gap and Lehigh Gap, Pennsylvania, in Subitzky, Seymour, ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, N.J., Rutgers University Press,p.132-205.

Epstein, J.B., and Epstein, A.G., 1972, The Shawangunk Formation (Upper Ordovician(?) to Middle Silurian) in eastern Pennsylvania: U.S. Geological Survey Professional Paper 744, 45p.

Epstein, J.B., and Lyttle, P.T., 1987, Structure and stratigraphy above, below, and within the Taconic unconformity, southeastern New York, *in* Waines, Russell H., ed., New York State Geological Association, 59th annual meeting, Kingston, N.Y., November 6-8, 1987, Field Trip Guidebook: New Paltz, N.Y., State University of New York, College at New Paltz, p. C1-C78.

Epstein, J.B., and Lyttle, P.T., 2001, Structural relations along the Taconic unconformity between New York, New Jersey, and Pennsylvania, *in*, 2001: A Delaware River Odyssey: Guidebook for the Annual Field Conference of Pennsylvania Geologists, v. 66, p. 22-27.

Epstein, J.B., Sevon, W.D., and Glaesser, J.D., 1974, Geology and mineral resources of the Lehighton and Palmerton 7 1/2-minute quadrangles, Pennsylvania: Pennsylvania Geological Survey 4th Ser., Atlas 195 cd, 460 p.

Ferrence, Jeanine, Ellis, Susan, Griffin, Jeaneen, Monteverde, Don, Witte, Ron, Epstein, Jack, Anderson, Lorraine, Leja, Marge, and Martin, Lisa, 2003, The Many Faces of Delaware Water Gap, A curriculum Guide for Grades 3-6; An integrated Guide to the Geology of Delaware Water Gap National Recretion Area: National Park Service, Delaware Water Gap National Recreation Area, 123 p.

Inners, J.D., and Fleeger, G.M., eds., 2001: A Delaware River Odyssey: Guidebook for the Annual Field Conference of Pennsylvania Geologists, v. 66, 314 p.

Johnson, D.W., 1931, Stream sculpture on the Atlantic slope, a study in the evolution of Appalachian rivers: New York, Columbia University Press, 142 p.

Meyerhoff, H.A., and Olmstead, E.W., 1936, The origins of Appalachian drainage: American Journal of Science, 5th Series, vol. 32, p. 21-42.

Ridge, J. C., 1983, The surficial geology of the Great Valley section of the Valley and Ridge Province in eastern Northampton Co., Pennsylvania, and Warren Co., New Jersey: unpublished M.S. thesis, Lehigh University, 234 p.

Snyder, D.H., ed., 1989, The Caves of Northampton County, Pennsylvania: Mid-Appalachian Region of the National Speleological Society, Bulletin #16, 46p.

Strahler, A.N., 1945, Hypothesis of stream development in the folded Appalachians of Pennsylvania: Geological Society of America Bulletin, v. 56, p. 45-88.

Thompson, H.D., 1949, Drainage evolution in the Appalachians of Pennsylvania: New York Academy of Science Annals, v. 52, art. 2, p. 31-62.

Van Houten, F. B., 1954, Sedimentary features of Martinsburg slate, northwestern New Jersey: Geological Society of America Bulletin, v. 65, p. 813-818.

Wintsch, R.P., Kunk, M.J., and Epstein, J.B., 1996, $^{40}Ar/^{39}Ar$ whole-rock data constraints on Acadian diagenesis and Alleghanian cleavage in the Martinsburg Formation, eastern Pennsylvania: American Journal of Science, v. 296, p. 766-788.

Witte, R.W., 2001, Late Wisconsinan Deglaciation and Postglacial History of the Minisink Valley, *in,* Inners, J.D., and Fleeger, G.M., eds., 2001 - a Delaware River Odyssey, Guidebook, 66th Annual Conference of Pennsylvania Field Geologists, Shawnee-on-Delaware, p. 99-118.

GLACIAL GEOLOGY AND GEOMORPHOLOGY OF THE PASSAIC, HACKENSACK, AND LOWER HUDSON VALLEYS, NEW JERSEY AND NEW YORK

Scott D. Stanford
New Jersey Geological Survey
P. O. Box 427
Trenton, NJ 08628
Scott.Stanford@dep.state.nj.us

Introduction

The New York City area is a gateway between glaciated terrain to the north and preglacial terrain and the Atlantic Ocean to the south. The Hudson valley is one of the three principal conduits into the Atlantic for meltwater draining from interior North America. Meltwater and sediment discharges from the Hudson have been invoked as the source of glaciogenic deposits and erosional features on the New Jersey continental shelf. Freshwater outflows from glacial lakes in the Ontario basin discharged down the Hudson during the late Wisconsinan deglaciation. Those outflows may have triggered the Younger Dryas and earlier climate coolings by suppressing ocean circulation in the North Atlantic (Donnelly and others, 2005; Rayburn and others, 2007; Obbink and others, 2010). More locally, glacial deposits are productive aquifers in places. Their strength properties and permeability are important features to consider when engineering infrastructure and foundations, modeling groundwater flow, and remediating groundwater and soil contamination.

Deposits of at least three glaciations are present in the New York City area. The earliest glaciation, known as the pre-Illinoian, may have occurred in the late Pliocene, between 2 and 2.5 Ma. A second glaciation is likely of late Illinoian age (~150 ka, oxygen-isotope stage 6), although it may be somewhat older. For ease of reference, it will be referred to as the Illinoian glaciation in this paper, although an Illinoian age is not proven. The most recent glaciation is of late Wisconsinan age and reached its maximum position, as dated by radiocarbon, between 21 and 20 ka (all dates related to the late Wisconsinan glaciation are stated in radiocarbon years). Fluvial deposits south of the glacial limit, and fluvial erosional features preserved within glaciated terrain, provide evidence for preglacial and interglacial routes of the Hudson and its tributaries. This paper will survey the geomorphic and glacial history of the lower Hudson, Hackensack, and Passaic valleys in northeastern New Jersey and adjacent New York, and will discuss lake-drainage events in the upper Hudson valley that affected the lower valley and shelf during the late Wisconsinan deglaciation. A final section will discuss the glacial stratigraphy of the buried valley aquifer system in the upper Passaic basin.

The observations presented here are based on surficial geologic mapping at 1:100,000 scale in New Jersey conducted in the 1980s and early 1990s as part of a cooperative mapping program by the N. J. Geological Survey and the U. S. Geological Survey (Stone and others, 2002), and subsequent 1:24,000 quadrangle mapping by the NJGS in the 1990s and 2000s conducted in part with funding from the Statemap component of the National Cooperative Geologic Mapping Program, administered by the USGS. All of the quadrangles covering the New Jersey part of the study area have been mapped and published (pdfs for all maps can be viewed and downloaded at (http://www.njgeology.org/pricelst/geolmapquad.htm). These mapping efforts built on the excellent early work of R. D. Salisbury, C. E. Peet, and H. B Kummel (in Salisbury, 1895, 1902; Merrill and others, 1902; Bayley and others, 1914). Observations in New York also draw from mapping by Woodworth (1901) and Fuller (1914) on Long Island, followed by subsurface mapping conducted as part of groundwater-resource investigations in Brooklyn and Queens (Perlmutter and Geraghty, 1963; Soren, 1978; Buxton and Shernoff, 1999), and the regional map compilation of Cadwell (1989). Lacustrine and fluvial features in the upper Hudson valley discussed in the section on Lake Albany were compiled by Stanford (2010) from the works listed in the caption for fig. 6.

Preglacial Fluvial Drainage

Nonglacial fluvial deposits outside the glacial limit, and wind gaps and buried valleys within the limit, document preglacial and interglacial routes of the Hudson and other rivers in the New York City area. For the purposes of this discussion, drainage before the pre-Illinoian glaciation will be referred to as "pre-early", that before the presumed Illinoian glaciation will be referred to as "pre-intermediate", and that before the late Wisconsinan glaciation will be referred to as "pre-Wisconsinan".

Pre-early drainage is documented by the Pensauken Formation (fig. 1). The Pensauken is a fluvial braidplain deposits of arkosic quartz sand and quartz-quartzite-chert gravel that forms a 10 to 12-mile (15 to 20-km) wide plain along the inner edge of the Coastal Plain from the glacial limit southwestward to the head of the Delmarva Peninsula, where the plain broadens and turns southward (Owens and Minard, 1979). The surface of the plain slopes from an altitude of 165 feet (50 m) near the glacial limit to 65 feet (20 m) in southern Delaware (fig. 1A), where it grades into a marginal-marine sand known as the Beaverdam Formation. Paleoflow direction measured at numerous locations on the tabular, planar cross beds that typify the Pensauken demonstrate southwesterly flow, in agreement with the surface slope. North of the glacial limit the Pensauken is glacially eroded but remnants of it occur beneath till in New Jersey and Staten Island for several miles north of the limit, and distinctive well-rounded white to yellow-stained quartz pebbles from the Pensauken are common in till in the Hackensack lowland and on Long Island. The Manetto Gravel in central Long Island (fig. 1A) is lithically similar to the Pensauken (Suter and others, 1949) and is preserved on uplands outside the glacial limit at elevations similar to those expected if the Pensauken plain were projected northeastward. While the Manetto has been interpreted as a glacial deposit (Fuller, 1914; Sirkin, 1986; Stone and Borns, 1986), its lithology and topographic position suggest that it is a Pensauken equivalent and marks former extension of the plain to the northeast.

A series of wind gaps in the Palisades and Watchung ridges provide further evidence of drainage alignment during Pensauken time. A gap in the Palisades at Sparkill, NY (1 on fig. 1 B and C) and a series of six aligned wind gaps through the three Watchung ridges at, from north to south, Paterson, Little Falls, Beaufort, Livingston, Short Hills, and Millburn (2 through 7 on fig. 1 B and C), all are of similar width and have preglacial rock floors (notched in places by later glacial and local fluvial erosion) that decline from north to south and grade to the Pensauken plain. These gaps mark the route of the Hudson during deposition of the Pensauken (Johnson, 1931; Stanford, 1993). The multiple crossings of resistant basalt and diabase is probably the result of superposition of the Hudson on a covering of Miocene sand (probably inland extensions of the Kirkwood or Cohansey formations of southern New Jersey) in the late Miocene.

A much broader gap is cut into the Palisades Ridge just north of Staten Island (8 on fig. 1 B and C). The 60 and 25 m (200 and 80 ft) elevation lines in this gap bracket the top and bottom of the thalweg of the Pensauken deposit as projected upvalley from the glacial limit. This projection fits the gap and also fits the Manetto remnants to the east. It marks the main trunk of the Pensauken river system, which likely included drainage from southern New England, including precursors to the modern Connecticut and Housatonic rivers. The Hudson was a tributary to this trunk Pensauken river, as was the Delaware, which joined the river to the south at Trenton. An upland in the Coastal Plain to the southeast separated the Pensauken valley from the Atlantic north of Delmarva.

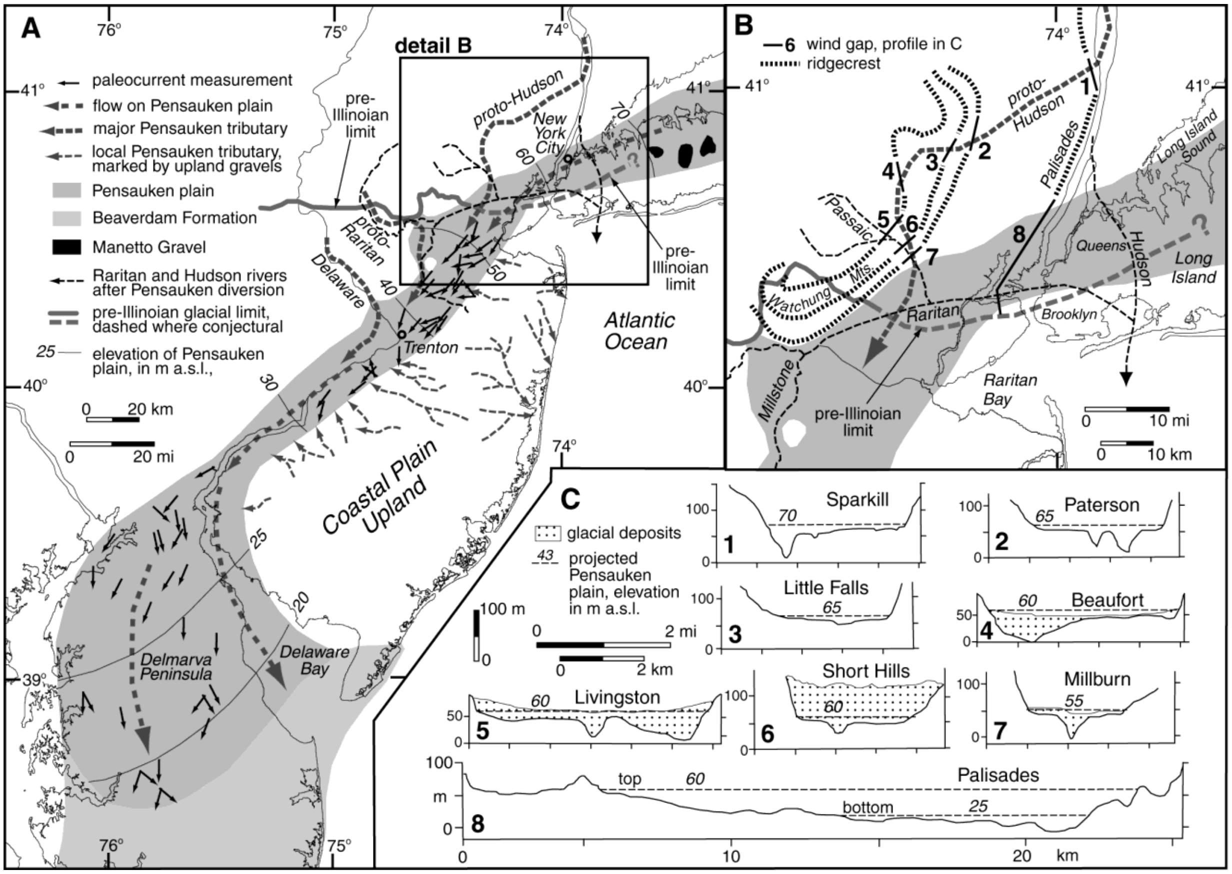


Figure 1. Preglacial and interglacial fluvial drainage, modified from Stanford (2010). A. Pensauken fluvial plain, with paleocurrent measurements (Owens and Minard, 1979; Martino, 1981; Stanford and others, 2002). B. Detail of the New York City area showing location of wind gaps (numbers correspond to profiles on C) and pre-intermediate route of the Hudson, Raritan, and Passaic rivers, marked today by buried valleys, established after the pre-Illinoian glaciation. Location of buried shelf valley from Schwab and others (2003). C. Profiles of wind gaps. Note scale change on profile 8.

The Pensauken in central New Jersey contains pollen indicating a Pliocene age (Stanford and others, 2001), and contains temperate-climate plant fossils (Berry and Hawkins, 1935) that indicate a nonglacial origin. The Beaverdam Formation, with which the Pensauken interfingers downvalley, is also of Pliocene age based on its pollen content (Groot and Jordan, 1999). The Pensauken plain probably aggraded during the mid-Pliocene eustatic highstand centered around 3.5 Ma, although the incised valley in which it aggraded had likely been in place since the latest Miocene, given the depth and extent of erosion into middle and late Miocene upland deposits bordering the Pensauken valley (Stanford and others, 2002).

During or shortly after the pre-Illinoian glaciation in the late Pliocene or early Pleistocene, the Pensauken river was diverted southeasterly in the New York City area, breaching the Coastal Plain upland and initiating direct access of the Hudson to the Atlantic. Details of the diversion are unknown because the depositional record has been removed by subsequent fluvial, marine, and glacial erosion. However, in the Raritan valley just west of the study area, the southernmost deposits of pre-Illinoian till overlie Pensauken Formation remnants and show erosional preservation and weathering intensity similar to the Pensauken, indicating that they are not widely separated in age. It is also clear from the distribution of pre-Illinoian till west of the late Wisconsinan glacial limit that the pre-Illinoian glacier advanced across the Pensauken plain in the New York City area (fig. 1B, see also Stop 3 of field trip).

Following this diversion a new drainage network became established. The Hudson drained southward, probably along the route of what is now the Harlem River and then across Queens by way of a now-buried valley (fig. 1B) (deLaguna, 1948; Soren, 1978; Buxton and Shernoff, 1999). This buried valley extends eastward onto the continental shelf off of Rockaway Beach (Schwab and others, 2003). The lower Raritan, and its tributaries the Passaic and Millstone, established an easterly course on the abandoned Pensauken plain as a tributary to the Hudson, crossing Brooklyn by way of another now-buried valley. Incision of these valleys, to depths of as much as 150 feet below the Pensauken plain, was accomplished in the early and middle Pleistocene, prior to the Illinoian glaciation, because Illinoian glacial deposits, including the Jameco Gravel on Long Island and Illinoian glaciolacustrine deposits and till in the upper Passaic basin, partly fill them. This network constitutes the “pre-intermediate” drainage that was in place before the Illinoian glaciation.

Some of this pre-intermediate drainage was no doubt altered by erosion and deposition during the Illinoian glaciation, although the only record of this alteration is in part of the upper Passaic basin where late Wisconsinan deposits are stacked on top of Illinoian deposits (see discussion of Lake Passaic, below, and fig. 8). Here, the pre-intermediate valleys were overdeepened in places by Illinoian glacial erosion and were then filled, or nearly filled, with Illinoian glacial deposits, indicating that post-Illinoian streams were locally dislocated from their previous routes. The main exit point for the Passaic through Second Watchung Mountain at the Short Hills gap was filled with Illinoian till to an elevation of about 200 feet (60 m), which was sufficient to divert the Passaic northward, in a manner similar to that of today, through the gap at Little Falls, which has a floor elevation of about 180 feet (55 m). Elsewhere, late Wisconsinan glacial erosion has removed all evidence of the post-Illinoian land surface.

Final modifications of the river network were made during the late Wisconsinan glaciation. The present Hudson channel between the Palisades and the Bronx and Manhattan was carved by deep glacial erosion along the outcrop belt of soft Stockton sandstone sandwiched between the resistant Palisades diabase and schist and gneiss of the Manhattan Prong. This erosion created an overdeepened trough more than 350 feet deep. The valley across Queens was filled with till and outwash and blocked by deposition of the terminal moraine. Similarly, the lower Raritan valley in New Jersey and Brooklyn was filled with outwash and blocked by the terminal moraine. The Raritan was diverted southeastward across a low shale divide and cut a narrow, gorge-like valley into the shale between Bound Brook and New Brunswick. East of New Brunswick, the rerouted Raritan entered, and broadened and deepened, a pre-existing valley in Cretaceous deposits that drained eastward to Perth Amboy and then into what is now Raritan Bay (Stanford, 1993).

Pre-Illinoian Glaciation

Pre-Illinoian till, and a few glaciolacustrine deposits, occur west of the late Wisconsinan glacial limit on the Watchung Mountains and intervening valleys (fig. 2). They have been entirely removed by glacial and fluvial erosion to the east, and have not been definitively identified on Long Island, although they surely were present there at one time. In their outcrop area, pre-Illinoian sediments are preserved in erosional remnants on flat hilltops and divides. They are absent from similar flat terrain within valleys below divide levels. This pattern indicates that the valleys were cut into bedrock since the pre-Illinoian glaciation, to depths of between 50 and 150 feet below the pre-Illinoian land surface. Pre-Illinoian deposits are also deeply weathered. Gneiss and arkosic sandstone gravel clasts are fully saprolitized or have thick (>0.5 inch) weathering rinds. Quartzite, quartz, and chert clasts are typically intact but stained; some have thin weathering rinds and are easily fractured with a hammer. Siltstone, quartz sandstone, and shale clasts are intermediate in weathering intensity between the gneisses and quartzites. The till matrix contains much secondary clay from alteration of feldspar in the sand fraction to clay minerals, and is reddish-yellow in color due to accumulation of iron oxides and hydroxides from weathering of mafic minerals.

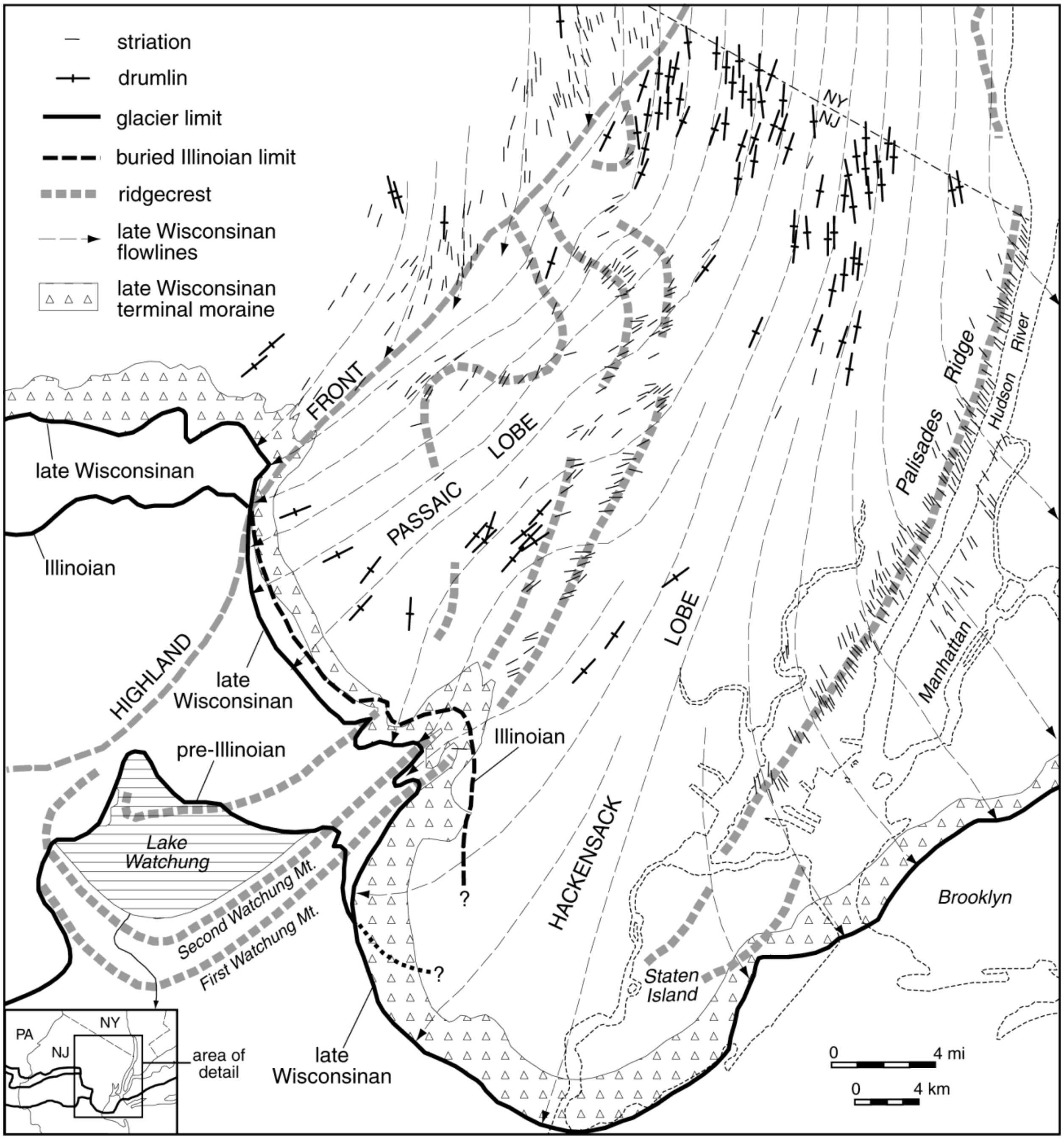


Figure 2. Glacial limits, late Wisconsinan striations and drumlin axes, and late Wisconsinan glacial flowlines. Striations and drumlins are from 1:24,000 surficial geologic maps available at http://www.njgeology.org/pricelst/geolmapquad.htm. Some striations on the Palisades are from Salisbury and Peet (1895). Striations on Manhattan are from Baskerville (1994).

Paleomagnetic measurements on pre-Illinoian lacustrine and fluvial sediments in New Jersey and eastern Pennsylvania yield both normal and reversed signatures, and some reversals are on weathering products (Sasowsky, 1994; Ridge, 2004). While there are many ways to interpret the paleomagnetic data, they at the very least prove a glaciation before 800 ka, the age of the last magnetic reversal. Pollen recovered from basal lake sediments in Budd Lake, a glacially dammed upland basin in the Highlands in western New Jersey, included pre-Pleistocene taxa (Harmon, 1968). Budd Lake lies just outside the

Illinoian and late Wisconsinan glacial limits, and so deposits in the lake basin were not eroded during these glaciations, and long accumulation is possible there. The basin was likely first dammed by deposition of glacial sediments during pre-Illinoian retreat. The Budd Lake pollen, and the similar erosional preservation and weathering properties of the pre-Illinoian till and Pensauken Formation, suggest a Pliocene age for the pre-Illinoian glaciation. The earliest Laurentide glacial deposits are dated by magnetic polarity and volcanic ash in the Missouri River valley to between 2 and 2.5 Ma (Boellstorff, 1978; Roy and others, 2004). The pre-Illinoian deposits here may be an eastern correlate of these earliest mid-continent tills.

The pre-Illinoian limit is mapped based on the well-defined extent of pre-Illinoian till (fig. 2). In addition to the preserved till patches, pre-Illinoian erratics occur within basal parts of the basalt colluvium at the foot of First Watchung Mountain within, but not outside, the limits shown on fig. 2. The limit defines a lobe in the broad Raritan lowland west of the Watchungs and a somewhat less extensive lobe in the Passaic and Hackensack lowlands. The Passaic and Hackensack lowlands, judging from the elevation of the wind gaps discussed in the previous section, were not as deeply eroded in pre-Illinoian time as they are now, or as they were in the middle and late Pleistocene, and so did not serve as effectively as channels for ice as they did in later glaciations. The upper Raritan lowland, in contrast, was largely as it is today and so provided little impediment to pre-Illinoian ice flow.

At its maximum extent the pre-Illinoian glacier enclosed a basin behind Second Watchung Mountain to create glacial Lake Watchung (fig. 2). With the Moggy Hollow gap (which served as the late Wisconsinan lake outlet, see section below on Lake Passaic) closed off by ice, Lake Watchung was controlled by a higher gap to the south that directed lake outflows southward into the Raritan River (Stanford, 2008). A sizable deposit of sand and gravel on the Passaic-Raritan divide at Bernardsville, NJ, is an erosional remnant of a delta deposited in this lake, and smaller deposits of lake clay and lacustrine-fan gravel in the Bernardsville-Basking Ridge area were also laid down in Lake Watchung.

Illinoian Glaciation

Illinoian deposits crop out in a belt several miles wide south of the late Wisconsinan limit west of the Highland Front, and occur beneath late Wisconsinan deposits in the valley fill in the upper Passaic basin between the Highland Front and Second Watchung Mountain (fig. 2, see also section below on Lake Passaic). They have not been observed to the east in the Hackensack lowland but are present in the cores of drumlins to the north near the state line. On Long Island, the Jameco Gravel and Montauk Till, both of which occur beneath Wisconsinan deposits, are likely of Illinoian age.

In contrast to the pre-Illinoian, Illinoian till in the outcrop belt is preserved on gentle and moderate slopes and is eroded only from steep slopes. Illinoian deposits fill modern valley bottoms, and form subdued but recognizable moraines, deltas, and fluvial plains. There has been no incision into bedrock beneath the Illinoian deposits. Gneiss and arkosic sandstone clasts have weathering rinds that are generally <0.5 inch thick and may be partially weathered but are not saprolitized. Matrix color is brown to yellowish-brown rather than reddish-yellow, and the matrix does not contain significant amounts of pedogenic clay. However, in a few places where soil B horizons in Illinoian deposits have been preserved, they are sufficiently well developed to indicate exposure during an interglacial rather than an interstadial (Ridge and others, 1990).

The only radiometric age control for the Illinoian advance is a coral fragment dated by U-Th to 130 ka from a marine sand overlying till on Nantucket (Oldale and others, 1982). This till may be equivalent to the Montauk Till on Long Island, and if so would indicate a pre-Sangamon age for that till, that is, Illinoian (oxygen-isotope stage 6) or older. The degree of weathering also indicates a pre-Sangamon age. On the other hand, the much greater degree of erosion and weathering of the pre-Illinoian deposits indicates a much greater period of time between the pre-Illinoian and Illinoian than between the Illinoian and late Wisconsinan, perhaps an order of magnitude difference.

Late Wisconsinan Glaciation

Advance of Ice

Striation and drumlin orientations show that late Wisconsinan ice advanced across the study area in two lobes, both part of the much larger Hudson-Champlain lobe (fig. 2). The largest mass, the Hackensack lobe, flowed south-southwesterly down the broad Hackensack lowland underlain by shale and sandstone between the Palisades Ridge to the east and First Watchung Mountain to the west. In the axis of the lowland, unimpeded by topography, it extended southward to a terminus at Perth Amboy, NJ, and the southern tip of Staten Island. At 40°30' this is the southernmost glaciated point in North America east of the Ohio River valley. The east sector of the lobe moved southeasterly over Manhattan, Brooklyn, and Queens to the terminal position. The western sector of the lobe flowed southwesterly over First and Second Watchung mountains and into the upper Passaic lowland, underlain by shale, to form the Passaic lobe. Impeded by the Watchungs, the Passaic lobe did not reach as far south as the Hackensack lobe. A third lobe, flowing out of the Wallkill valley, advanced southward across the Highlands to the west of the Highland Front. Impeded by the topographic barrier of the Highlands, it did not advance as far as the other two and did not enter the Passaic lowland (except at the mouth of the Wanaque valley, see Stop 4 of field trip).

During advance the glacier eroded elongate overdeepened troughs into shale and sandstone bedrock in the Passaic and Hackensack lowlands and Hudson valley. These troughs show as much as 300 feet of overdeepening and extend to more than 350 feet below sea level in the New York City area and, in the Hudson valley, as much as 700 feet below sea level at and north of the Tappan Zee (Newman and others, 1969). To the south they shallow and transition into buried pre-Wisconsinan fluvial topography within about 10 miles of the glacial limit. The troughs in the Passaic and Hackensack are parallel to ice flow, suggesting they were produced by scour. The trough beneath the Hudson, which is excavated chiefly in sandstone, is nearly perpendicular to ice flow over the Palisades, suggesting that it may have been produced by plucking like the Palisades cliffs themselves, or perhaps by scour from local basal ice channeled southwesterly along strike of the sandstone belt.

Terminal Moraine

At its maximum extent, the ice front stabilized and built a prominent terminal moraine in a belt averaging about 2 miles (3 km) wide and consisting of till between 40 and 200 feet (12 and 60 m) thick. The moraine belt consists of ridges, knolls, and basins with as much as 100, but generally less than 50, feet (30 and 15 m) of relief. By volume of till, the terminal moraine is at least an order of magnitude larger than the few recessional moraines mapped in the northeastern United States.

Varves deposited in the southern basin of Lake Passaic (fig. 3, portion of Lake Passaic south of margin TM) provide an estimate of residence time at the moraine. This lake basin formed when ice arrived at the terminus and was isolated from glacial sediment input when the ice front retreated from the moraine, because the moraine itself and its fronting delta formed a continuous barrier above lake level along the north edge of the basin. Reimer (1984) counted 750 glacial varves in the basin, overlain by at least 450 much thinner "microvarves" that likely represent postglacial sedimentation. A 750-year duration for residence at the terminal moraine is consistent with regional bracketing radiocarbon dates (fig. 4).

Looping of the moraine at the margin of the Hackensack lobe around the Todt Hill upland underlain by serpentinite bedrock on Staten Island indicates that the glacier surface rose at a rate of about 300 feet per mile, which is the same gradient observed along terminal moraine loops in the New Jersey

Highlands. This rate of rise indicates that, at maximum advance, the late Wisconsinan glacier was between 1500 and 2000 feet thick over Manhattan, somewhat less than claimed in disaster movies but enough to bury the Empire State Building.

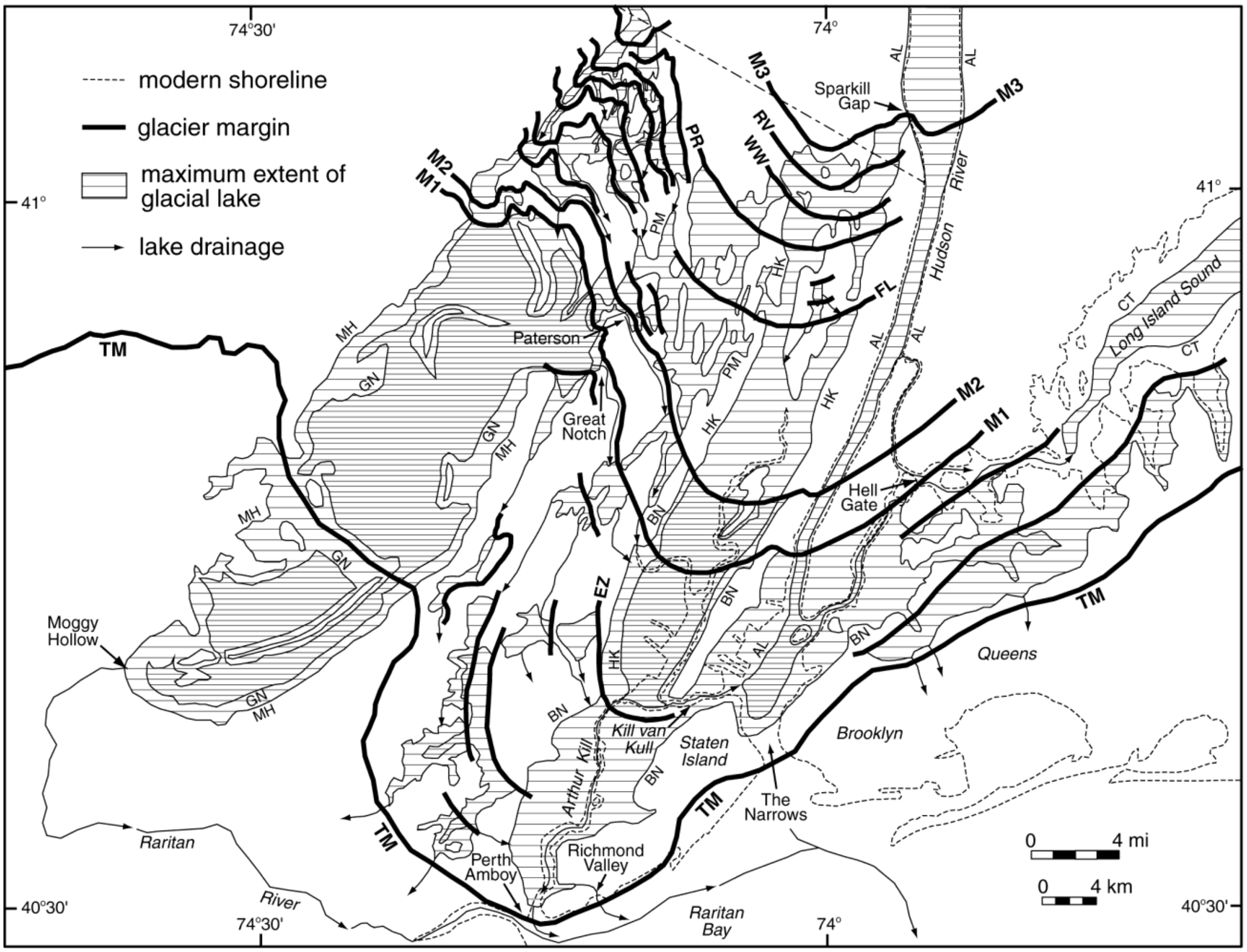


Figure 3. Glacial lakes and ice margins during late Wisconsinan retreat. Lakes are identified by the following abbreviations on their shorelines: AL=Albany, BN=Bayonne, CT=Connecticut, HK=Hackensack, PM=Paramus, MH=Passaic, Moggy Hollow stage, GN=Passaic, Great Notch stage. Ice margins are: TM=terminal moraine, M1=last ice margin before Lake Bayonne lowers to form Lake Albany, Hell Gate stage, in the Hudson valley, and before Lake Passaic lowers from the Moggy Hollow stage to the Great Notch stage, M2=last ice margin before Lake Passaic, Great Notch stage drains, and before spillway erosion establishes stable Lake Hackensack, M3=last ice margin before Lake Hackenssack lowers through Sparkill Gap into Lake Albany. Recessional ice margins marked by large deltas or glaciofluvial plains are: EZ=Elizabeth, FL=Fair Lawn, PR=Paramus, WW=Westwood, RV=Rivervale.

Retreat

Retreat from the moraine is documented by lacustrine stratified deposits, and a few glaciofluvial and ice-contact deposits, that mark ice-margin positions or that record lake stages that in turn require certain ice-margin geometries (fig. 3). Lake-stage elevations are marked primarily by delta-plain elevations or, ideally, by the elevation of the contact of fluvial topset beds and lacustrine foreset beds within deltas, although these contacts are rarely exposed. Discontinuous beach features have been observed in Lake Passaic (Salisbury and Kummel, 1895), and were likely present in other broad lake basins with sufficient fetch like lakes Bayonne and Hackensack, but today all such fine topographic details have been erased by urbanization.

Lakes formed in valleys that drained toward, and so were dammed by, the glacier margin, or in valleys and lowlands dammed by earlier glacial deposits. In the latter class are Lake Bayonne (BN on Fig. 3) and the Hell Gate stage of Lake Albany (AL on fig. 3), which were dammed by the terminal moraine, and Lake Paramus (PM on fig. 3) which was dammed by an earlier delta that clogged a narrow reach of the lower Passaic valley. In the former class are the many small lakes along the west edge of the Hackensack lobe, which occupied east- or northeast-draining tributary valleys in the Rahway, lower Passaic, and Saddle river valleys that were dammed by the retreating ice front. Lake Passaic is a combination of both types, with an early stage of the lake forming when ice blocked the northeast-draining upper Passaic valley at Paterson, and a maximum (MH on fig. 3) and recessional (GN on fig. 3) stage held in by the terminal moraine dam in the Short Hills gap.

During retreat the terminal moraine dammed the Arthur Kill lowland between Perth Amboy and Staten Island and also the Narrows between Staten Island and Brooklyn, forming the basin occupied by Lake Bayonne. The earliest level of Lake Bayonne was controlled by a spillway across the moraine at Richmond Valley on Staten Island but this was soon succeeded by an eroding spillway across the moraine at Perth Amboy. Lake Bayonne expanded northward to the Newark area and eastward into lower Manhattan as the ice front retreated. When ice uncovered a till upland in northwestern Queens (M1 on fig. 3), a lower spillway draining eastward into the Long Island Sound lowland opened and Lake Bayonne drained. This lower spillway soon stabilized on gneiss bedrock at what is now Hell Gate in the East River at an elevation of -30 feet (-9 m). At this time the Long Island Sound lowland was occupied by glacial Lake Connecticut, which was controlled by a moraine dam at The Race off the east end of Long Island (fig. 5) (Lewis and Stone, 1991). At its west end, the level of Lake Connecticut at this time was -70 feet (-21 m) or lower, providing at least 40 feet (12 m) of drop for the Hell Gate spillway to function.

As Lake Bayonne lowered and drained it was replaced in the Hudson valley by the Hell Gate stage of Lake Albany, controlled by the stable spillway on gneiss in Hell Gate. In the Hackensack valley to the west of the Palisades, outflows from the remnant of Lake Bayonne eroded the Arthur Kill and Kill van Kull channels to uncover Palisades diabase bedrock at Tremley Point in the Arthur Kill at -30 feet (-9 m) and at the west end of the Kill van Kull at -20 feet (-6 m). These became the stable spillways for Lake Hackensack, which was 40 feet (12 m) higher than the Hell Gate stage of Lake Albany.

The uncovering of Hell Gate also coincides with the uncovering of Great Notch in First Watchung Mountain, which caused Lake Passaic to lower 80 feet (24 m) from its highest (Moggy Hollow, MH on Fig. 3) stage to the Great Notch (GN on fig. 3) stage (M1 on fig. 3). This lowering released about 2.5 mi^3 (10 km^3) of water into Lake Bayonne via the Third River sluice, which provided an erosional pulse that deepened the Arthur Kill and Hell Gate channels. With continued retreat, the Paterson gap in First Watchung Mountain (M2 on fig. 3, Stop 5 on field trip) was uncovered, and the Great Notch stage itself drained, releasing about 1.2 mi^3 (5 km^3) of water down the Weasel Brook sluice into the lower Passaic valley. This outflow provided a final erosional pulse in the Arthur Kill and Kill van Kull channels, uncovering the spillways for Lake Hackensack and, possibly, completing erosion down to gneiss bedrock at Hell Gate.

Continued retreat uncovered Sparkill gap, a broad wind gap containing a narrow inner notch with a floor at an altitude of 30 feet (9 m) in the Palisades ridge just north of the state line (M3 on fig. 3). Lake Hackensack lowered 40 feet (12 m) to the level of Lake Albany, Hell Gate stage, through the notch. After the lowering, a shallow postglacial lake remained in the northern half of the Hackensack valley. The north end of this postglacial lake was filled with fluvial terrace sand fed by the Passaic and Saddle rivers, which flowed northeasterly across the Lake Paramus basin to Sparkill gap for several thousand years after deglaciation, until isostatic rebound shifted them to their present southerly flow at about 13 ka (Stanford and Harper, 1991).

During this early postglacial period, before isostatic rebound, the floor of lakes Hackensack and Bayonne in the southern part of the Hackensack lowland was exposed and desiccated. Desiccation created an overconsolidated horizon in the uppermost lake clays, beneath post-rebound alluvial sand and Holocene salt-marsh peat, which is well known from engineering studies (Lovegreen, 1974). Depth of the desiccated zone is as much as 60 feet (18 m) below sea level at Newark and 30 feet (9 m) below sea level at Secaucus which, when adjusted for rebound at the rate of 3.5 feet per mile (0.7 m/km) recorded by delta elevations in stable Lake Hackensack and Lake Albany, Hell Gate stage, approximates the pre-rebound Sparkill gap baselevel at 30 feet (9 m)(Stanford and Harper, 1991).

Chronology

Chronologic control for the late Wisconsinan advance and retreat in the New Jersey-Long Island area is provided by radiocarbon dates and varve counts (fig. 4). All the dates younger than 13 ka are on freshwater peat or terrestrial plant material, except the bone dates for the Sparta mastodon. The older dates are on concretions in varves (the Great Swamp date), and disseminated organic matter in sediment, since no peat or plant macrofossils have been recovered from these deposits, except for the older Long Island date, which is on peat. The two Long Island dates shown on fig. 4 are the youngest of 29 total dates on shells, peat, wood, and organic silt from deformed beds under late Wisconsinan till at Port Washington, Long Island. The other 27 older dates range from 25 to >43.8 ka (Sirkin and Stuckenrath, 1980). Bulk organic sediment and concretions may incorporate old geologic carbon that produces older-than-actual dates for the enclosing deposit. Consistency of the dates, and intervening varve counts, suggests that contamination with old carbon is not a significant problem here. In addition, the dates are consistent with the New England varve chronology. The oldest varves in this chronology date the recessional ice margin at Newburgh, NY to 15 ka (Ridge, 2004, 2008). A minimum of 3200 additional years are recorded by varves in Lake Hackensack and the lower Hudson valley that are not matched to the chronology but precede it (Antevs, 1928). These additional varves indicate that Little Ferry, NJ, the site of the earliest varves, was deglaciated at or before 18.6 ka. This date is consistent with the 19-19.2 ka deglaciation for Little Ferry inferred from the radiocarbon chronology (fig. 4).

Ice arrived at its maximum position at about 21 ka, remained at the terminal moraine for 750-1000 years, based on the glacial varve count in the southern basin of Lake Passaic, and then retreated rather rapidly (an average of 50 m per year, using the smooth recessional curve on fig. 4) up the Hackensack lowland, if the Tappan Zee date of Weiss (1971) is accurate. This date is in organic sediment underlain by 108 feet of varves resting on till. The varves represent a minimum of 200-300 years of elapsed time from deglaciation to deposition of the organic sediment, assuming they are thick (3 to 6 inches or 7 to 14 cm) proximal varves. This elapsed interval is within the one sigma error on the date.

Lake Hackensack varves counted in clay pits at Little Ferry, NJ (Reeds, 1926) show an abrupt thickening at varve year 1097 (Reeds' varve 0) above the basal till. This thickening is likely the result of the lowering of Lake Hackensack when Sparkill gap was deglaciated. At this time, pre-rebound routing of the Passaic and Saddle rivers into the shallow postglacial lake increased sediment inflow, accounting for the thicker varves. The 1097 glacial varves thus measure the duration of retreat from Little Ferry to Sparkill gap, indicating a slower overall retreat rate in the northern Hackensack lowland compared to that further south. This slowed retreat is consistent with the greater volume of ice-contact deltaic sediment laid down in lakes Paramus and Hackensack (from ice margins FL, WW, RV, and M3 on fig. 3) during this interval.

Residence time at these and earlier recessional positions can be estimated from the deglaciation chronology tied to the Tappan Zee date. Large deltas at Newark (in Lake Bayonne), Fair Lawn (in Lake

Paramus), Westwood, Rivervale, and Tappan (all in Lake Hackensack), a glaciofluvial plain and underlying lacustrine valley fill at Elizabeth, and a small recessional till moraine (the Bloomfield moraine) associated with the Newark delta and with other deltaic deposits northwest of Newark, all appear to represent 100- to 300-year intervals of ice-margin stability. Outcrops and test borings showing till over stratified deposits record short (<1 mile) readvances at ice margins EZ (Stanford, 2002) and M1, especially along M1 in the Lake Passaic basin (Salisbury, 1902; Stanford, 2003), where deep water buoyed the ice and increased margin mobility. There is no evidence of readvance elsewhere.

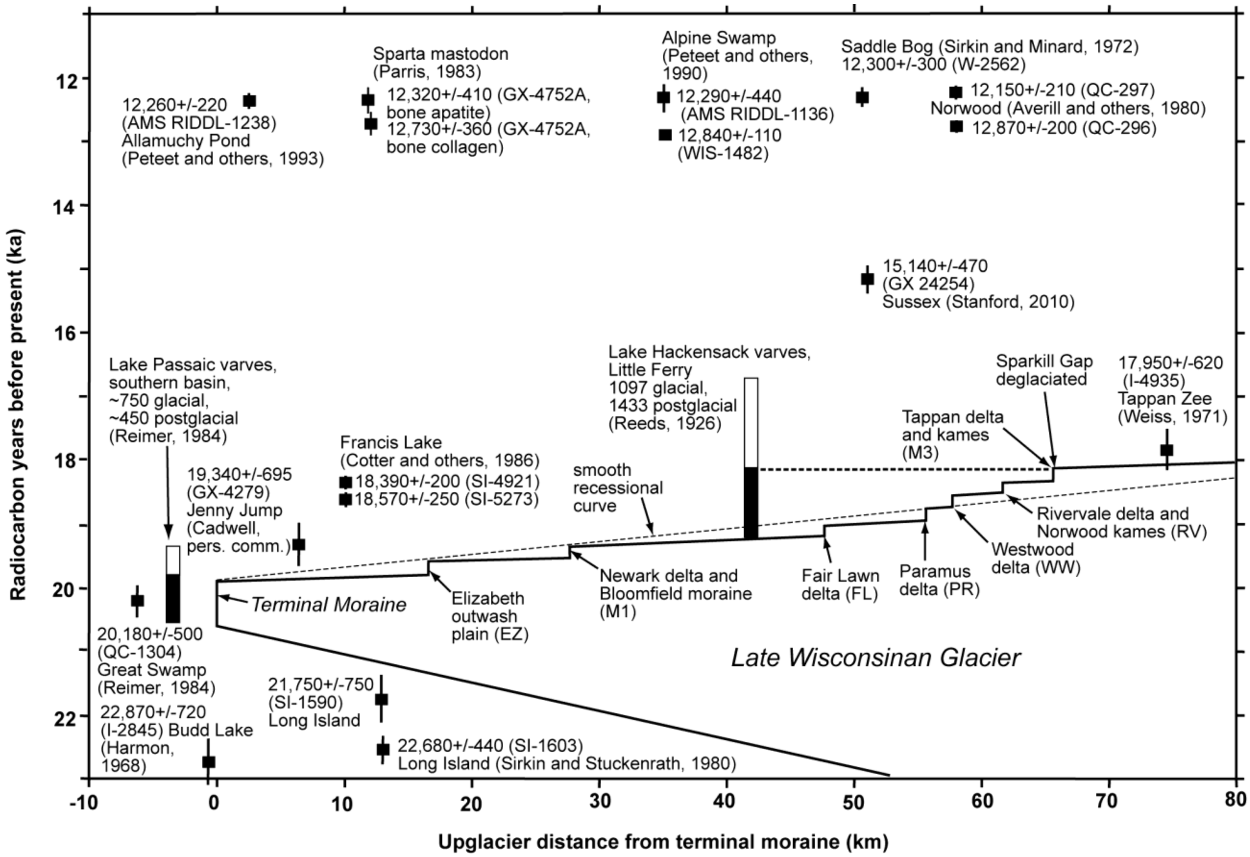


Figure 4. Time-distance plot of late Wisconsinan glaciation of study area. Radiocarbon dates are from the New Jersey-Long Island area. Recessional positions and ages are for the Hackensack lobe. Vertical lines at dates are one-sigma errors. Bold dotted line connecting top of glacial varves at Little Ferry and Sparkill dates the deglaciation of the gap and lowering of Lake Hackensack.

Lake Albany

The Hell Gate stage of Lake Albany extended up the Hudson valley to just north of the Hudson Highlands, where delta elevations (fig. 6) show the onset of an unstable phase of the lake. This unstable phase records lowering lake level, which indicates that the Narrows moraine dam was breached and an eroding spillway on till in the Narrows had replaced Hell Gate as the spillway. Rapid drainage of water from the Augusta stage of Lake Wallkill (fig. 5) at about 15.5 ka is the cause of the Narrows dam breach. The Augusta stage filled the ice-dammed north-draining Wallkill valley and was controlled by a spillway on the Wallkill-Delaware divide that drained southward into the Delaware basin (Stone and others, 2002). When the Moodna Creek valley at the north end of Schunnemunk Mountain, and Storm King in the main Hudson valley to the east, was deglaciated, the Augusta stage dropped 230 feet (70 m), releasing 6 mi^3 (25 km^3) of water into Lake Albany (Stanford, 2010). At this time Lake Albany covered an area of 79 mi^2 (205 km^2), about the same as the present-day Hudson River downstream from Moodna Creek, and the sudden inflow would have temporarily raised the level of Lake Albany by 390 feet (120 m) if added all at once. Topography of the Moodna Creek and Wallkill basins upstream of the outburst point suggests that perhaps 100 feet (30 m) of the 230 foot (70 m) fall occurred suddenly, which would have raised Lake Albany by 160 feet (49 m). Since the Hell Gate stage water level projects to the Narrows dam at -70 feet (-21 m) and the height of the moraine in the Narrows was around 50 feet (15 m), judging from the headlands bordering the Narrows today, the 120 feet (36 m) of freeboard on the dam would have been overtopped by the 160 foot (49 m) rise. Alternatively, if the outburst was not sudden, the added

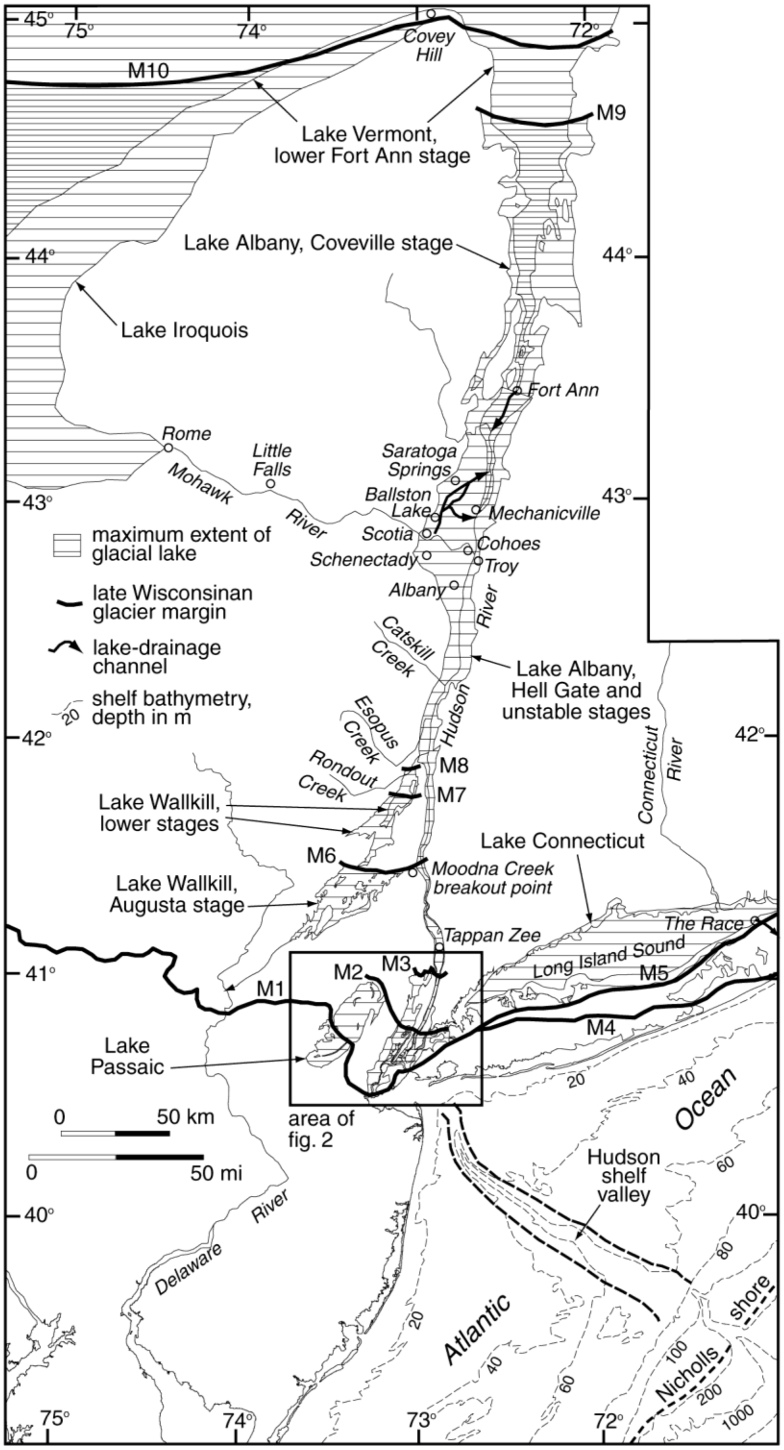


Figure 5. Lakes Wallkill, Albany, Iroquois, Vermont, and Connecticut, and the Hudson shelf valley. Ice margin abbreviations are: M1=terminal moraine in New Jersey and Pennsylvania, M2= last ice margin before Hell Gate

stage of Lake Albany was established, M3=last ice margin before Lake Hackensack lowers to Lake Albany, M4=Ronkonkoma Moraine, the late Wisconsinan terminus on Long Island, M5=Harbor Hill Moraine, M6=last ice margin before the Moodna Creek breakout flood, M7 and M8=last ice margins before draining of two lower stages of Lake Wallkill, M9=approximate last margin before stable Coveville stage of Lake Albany begins to lower, M10=last margin before Lake Iroquois lowers to Lake Vermont.

volume in Lake Albany could have caused seepage failure of the dam. If Storm King rather than Schunnemunk Mountain was the site of the outburst, which may have been the case depending on local ice-margin geometry, then there would be a slight increase in the released volume due to extension of Lake Wallkill into the Woodbury Creek valley between Schunnemunk Mountain and Storm King.

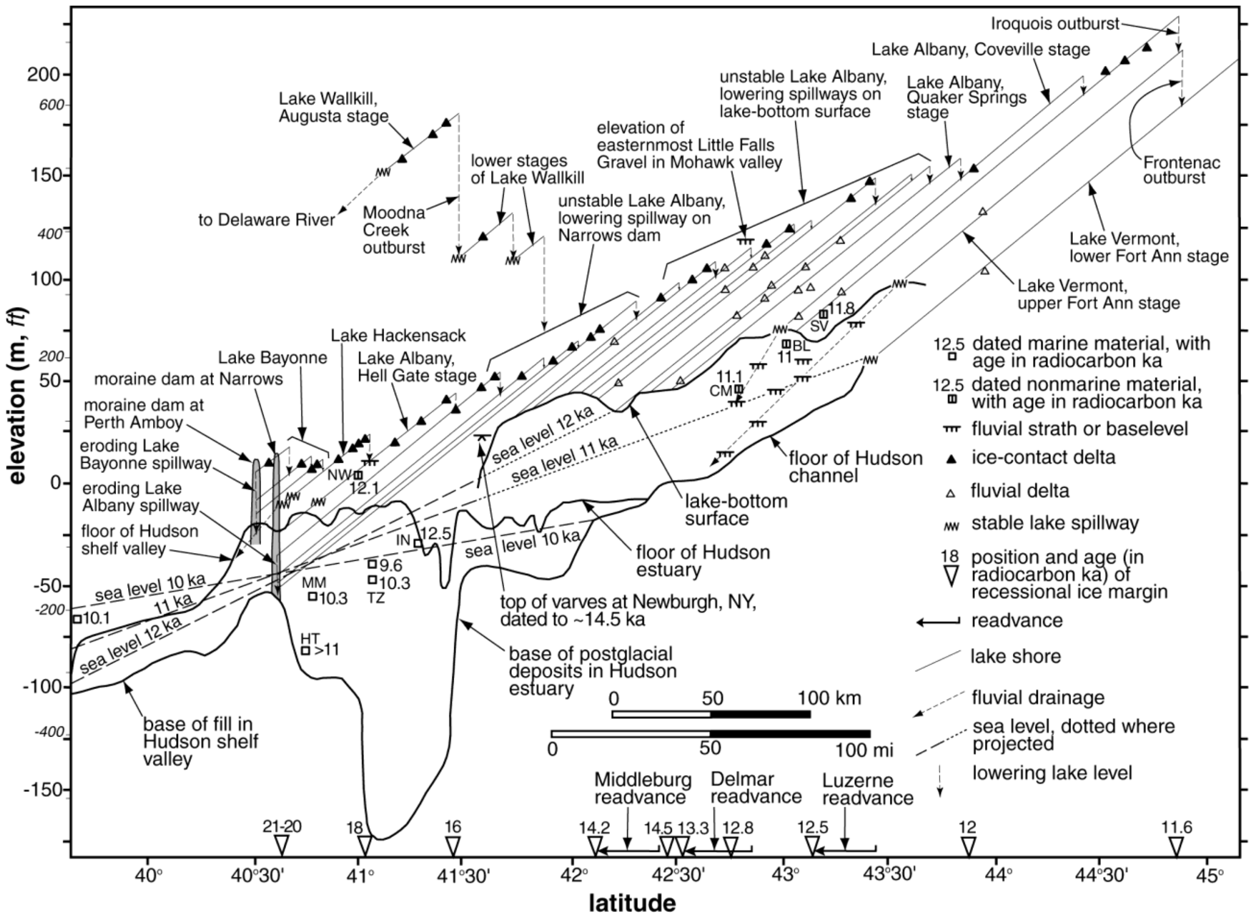


Figure 6. Present elevation of glacial-lake shorelines, lake-bottom surface of glacial Lake Albany, floor of the Hudson channel cut into the lake-bottom surface, top and bottom of postglacial fill in the Hudson estuary, and top and bottom of fill in the Hudson shelf valley. From Stanford (2010). Position and age of recessional ice margins and radiocarbon dates that bracket marine incursion are also shown. All elevations are projected to a north-south line perpendicular to isobases. Deltas, spillways, and fluvial features are compiled from: Woodworth (1905), Fairchild (1919), LaFleur (1965), Connally (1973), Connally and Sirkin (1970, 1973, 1986), Dineen and Hanson (1983), DeSimone and LaFleur (1986), Dineen (1986), Cadwell and Dineen (1987), Cadwell (1989), Stanford and Harper (1991), Wall and LaFleur (1994), Ridge (1997) and DeSimone and others (2008). Shorelines for the Coveville and Fort Ann stages north of 43°40' are from Rayburn and others (2005). Dates of recessional ice margins are from Ridge (2004, 2008), with minor additions in the lower valley. Base of postglacial fill in the estuary is from Worzel and Drake (1959), Newman and others (1969), Weiss (1974), and Lovegreen (1974). Base of fill in the shelf valley is from Thieler and others (2007). Radiocarbon dates are from Newman and others (1969), Averill and others (1980), Weiss (1974), Toney and others (2003), Donnelly and others (2005), Dineen and Miller (2006), and Miller (2008). Abbreviations for locations of dates are: NW=Norwood, NJ, MM=midtown Manhattan, HT=Holland Tunnel, TZ=Tappan Zee, IN=Iona Island, CM=Cohoes mastodon, BL=Ballstone Lake, SV=Schuylerville.

With the Narrows dam breached, Hudson valley drainage could now exit directly to the continental shelf in the New York Bight rather than around the east end of Long Island. This drainage cut the Hudson shelf valley, which leads from the Narrows to the edge of the continental shelf (fig. 5). The shelf in this region was subaerially exposed until about 10 ka, and the shelf valley itself was not flooded by marine incursion until 12 ka, leaving about 3500 years for fluvial incision of the shelf valley by Lake Albany outflows after dam breaching (fig. 6).

Erosion of the Narrows dam and the shelf valley was accelerated when Great Lakes outflows entered the Hudson via the Mohawk valley. During the Erie Interstade between 14.5 and 14.2 ka the Mohawk valley was temporarily deglaciated and conducted outflows from the Ontario and, possibly, the Erie basin into the Hudson (Ridge, 1997). This period of Great Lakes outflow corresponds to final erosion of the Narrows dam, to an elevation of about -200 feet (-60 m), because the altitude of deltas deposited at this time in the Albany area shows that the threshold for Lake Albany had shifted northward onto the emerging lake bottom north of the Hudson Highlands (fig. 6). From 14.2 to 12.7 ka readvances blocked the Mohawk valley but from its final deglaciation at 12.7 ka to 11 ka the Mohawk again conducted Great Lakes outflows into the Hudson valley, this time from Lake Iroquois in the Ontario basin (Wall and LaFleur, 1994). The threshold for Lake Albany continued to incise into emerging lake-bottom sediments downvalley, and migrated northward as isostatic rebound progressively raised the lake bottom to the south. The outflow from the threshold cut the channel that now contains the Hudson River in the mid-Hudson valley.

The rising sea had entered the lower Hudson valley by 12 ka (Newman and others, 1969), and from 12 to 10 ka, when rebound had elevated the upper valley above marine influence, high relative sea level in the lower valley limited the depth of fluvial incision in the Hudson channel upvalley. The inability of lake outflows to incise into the lake bottom during this interval created quasi-stable thresholds for the Quaker Springs and Coveville stages of Lake Albany and the upper Fort Ann stage of Lake Vermont (fig. 6).

When the north end of the Adirondack upland at Covey Hill was deglaciated at 10.9 ka, Lake Iroquois lowered in two stages (the "Iroquois outburst" and the "Frontenac outburst" on fig. 6) to the lower Fort Ann stage of Lake Vermont (Rayburn and others, 2005). These were enormous releases of 170 mi^3 (700 km^3) and 600 mi^3 (2500 km^3), respectively (Rayburn and others, 2005), which were the largest, and last, floods to discharge down the Hudson valley. High relative sea level in the Hudson valley at this time, however, prevented deep fluvial scour. By 10.6 ka, retreat opened the St. Lawrence valley and glacial sedimentation and meltwater discharge in the Hudson valley ceased.

History of Lake Passaic

The arcuate trace of the west-dipping cuesta ridges of the Watchung Mountains, formed on resistant basalt flows, reflects the canoe-shaped geometry of the Watchung syncline. The Ramapo Fault on its west side brings Proterozoic gneiss against the syncline. Erosion of the soft shale and sandstone enclosed between the basalts and the gneiss excavated a 30 mile long by 10 mile (48 by 16 km) wide basin between Second Watchung Mountain and the Highland Front. This basin is punctured by only two sets of gaps: the paired gaps at Little Falls and Paterson and at Short Hills and Millburn. These gaps are inheritances from the Pliocene course of the Hudson (see discussion on preglacial fluvial drainage above). Pre-intermediate drainage (fig. 7A) of the basin was established in the early Pleistocene after the pre-Illinoian glaciation and diversion of the Pensauken-Hudson river system. It exited the basin through the Short Hills-Millburn pair of gaps.

Illinoian glacial deposits in and south of the Short Hills gap indicate that Illinoian ice sealed this gap and thereby created an Illinoian version of Lake Passaic, probably similar to the late Wisconsinan Moggy Hollow stage. Illinoian delta and lacustrine-fan sand and gravel, and lake-bottom silt and clay, occur beneath late Wisconsinan deposits in the central and southern sections of the basin ("overramp zone' on fig. 8), where they fill the pre-intermediate valleys and also extend over the low interfluves between the valleys. The Illinoian fill in the Short Hills gap rises to about 200 feet (60 m) in elevation (section AA', fig. 8), so during Illinoian retreat a lake was maintained at that level until the gaps at Little Falls and Paterson were uncovered. This lake was much shallower than the late Wisconsinan recessional stages and limited the vertical accretion of Illinoian valley fill sediments north of the Short Hills gap. Erosion of the Illinoian valley fill during the Sangamon interglacial and early and middle Wisconsinan was minimal, because basalt in the Little Falls gap at an elevation of 180 feet (55 m) established a high base level for the basin, limiting fluvial incision.

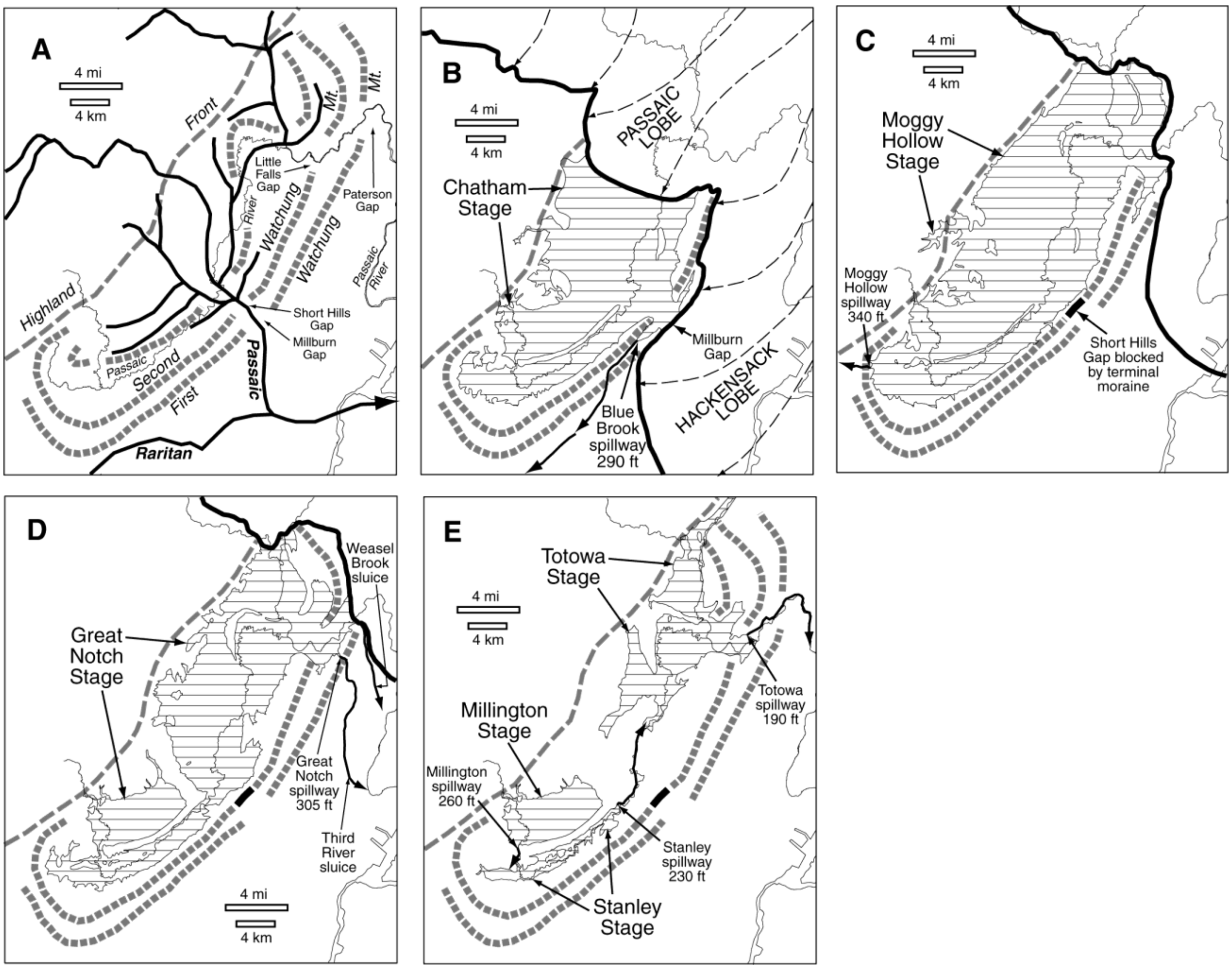


Figure 7. History of Lake Passaic. A. Fluvial drainage before the Illinoian glaciation. B. Advancing Hackensack lobe of late Wisconsinan glacier blocks Millburn gap, establishing the Chatham stage. Continued advance of Hackensack lobe blocks Short Hills gap, establishing the Moggy Hollow stage. C. Maximum extent of Moggy Hollow stage, just before uncovering of Great Notch. D. Maximum extent of Great Notch stage. Uncovering of Great Notch released 2.5 mi ^{3} (10 km^3) of water down the Third River sluice. Uncovering of Garrett Mountain released 1.2 mi^3 (5 km^3) of water down the Weasel Brook sluice. E. Postglacial stages.

The late Wisconsinan history of the lake includes three glacial lake stages (fig. 7B, C, D) and three postglacial lakes (fig. 7E). When the advancing Hackensack lobe, which extended further south than the Passaic lobe, blocked Millburn gap, the Chatham stage of the lake formed (fig. 7B), controlled by a spillway at the head of the Blue Brook valley at an elevation of 290 feet (88 m) (Stop 3 of field trip). Filling of this lake stage buoyed the Passaic lobe ice and allowed the Passaic lobe to ramp over rather than erode pre-existing sediments as it advanced to the terminal position. These pre-existing sediments included Illinoian deposits and delta, fan, and lake-bottom sediments laid down in the Chatham stage in front of advancing ice (fig. 8, sections AA', BB'). An earlier, shallow, advance-phase lake probably formed when the Hackensack lobe blocked the Paterson gap. At this time the Passaic River was ponded to the level of the Illinoian fill in the Short Hills gap. However, since this fill was only about 20 feet (6 m) higher than the Little Falls base level, the lake was shallow, accumulated little sediment, and did not buoy the Passaic lobe.

Continued advance of the Hackensack lobe blocked the Blue Brook valley and then sealed the Short Hills gap. This caused the Chatham stage to rise 50 feet (15 m) to the Moggy Hollow stage (fig. 7C), which was controlled by a spillway into the Raritan basin across Second Watchung Mountain near Far Hills. The 50-foot (15 m) rise in lake level further buoyed the Passaic lobe, allowing it to ramp over the back end of a large fan-delta complex built into the lake at the terminal position. Till of the terminal moraine was deposited along the back end of this large delta deposit (fig. 8, sections AA', BB').

Deposition of till of the terminal moraine in the Short Hills gap filled the gap to an elevation of more than 400 feet (122 m), about 50 feet (15 m) higher than the Moggy Hollow stage, allowing this lake to expand northward as the ice front retreated. When Great Notch, a gap through First Watchung Mountain east of Little Falls, was uncovered, the Moggy Hollow stage dropped 80 feet (24 m) to the Great Notch stage (fig. 7D). This drop released about 2.5 mi^3 (10 km^3) of water down the Third River sluice downhill from Great Notch into the lower Passaic. The configuration of the glacier margin at this time (M1 on fig. 3, also on fig. 7C) is fixed by the last ice-contact deltas deposited in the Moggy Hollow stage: one along the Highland Front near Riverdale (Stop 4 of field trip) and one in the north end of the Preakness valley, where the ice front was lodged along the crest of Second Watchung Mountain.

Further retreat of the Hackensack lobe uncovered the north end of First Mountain at the Paterson gap (Stop 5 of field trip), allowing the Great Notch stage to drain down the Weasel Brook sluice. This flood released about 1.2 mi^3 (5 km^3) into the lower Passaic valley. Again, the position of the glacier margin at this time is fixed by Great Notch-stage deltas, including a delta at Riverdale (Stop 4 of field trip) and one at the north end of the Preakness valley. Both of these deltas are reworked from adjacent Moggy Hollow deltas by meltwater draining from local lakes adjacent to Lake Passaic, held in by ice deployed as shown in fig. 7D.

After the Great Notch stage drained, sediment dams held in three postglacial lakes. At Totowa, a large lacustrine fan blocked the Passaic valley in a narrow reach downstream from Little Falls, forming the dam for the Totowa stage. This lake extended up the Pompton and Ramapo valleys, where several large ice-contact deltas were deposited in it, including the Pompton plain, which nearly filled the northern bay of the Totowa lake, and a large delta which fills the Ramapo valley further north. Sand terraces were deposited by the Rockaway, Whippany, and Passaic rivers in the western and southern bays of the lake.

The terminal moraine formed a dam across the Passaic at Stanley, forming the Stanley stage. The Passaic and Dead rivers deposited silt and fine sand terraces in this shallow lake. The Great Swamp basin north of Lake Stanley was dammed by the terminal moraine-delta complex to the northeast. Postglacial Lake Millington filled this basin and was controlled by a spillway at Millington in a gap on Long Hill, the basalt ridge that forms the southern rim of the basin. Sand, reworked from the deltas to the north, was deposited in terraces in the northeast end of Lake Millington by Loantaka and Great Brooks.

Each of the postglacial lakes was controlled by spillways on erodible material (sand and gravel for Totowa, till for Stanley, and weathered and fractured basalt for Millington). The spillways were gradually lowered by erosion and the lakes eventually drained to leave the broad floodplains and marshes that now occupy their floors. Today, these floodplains and marshes are valuable open spaces that provide thousands of acres of flood storage and wildlife habitat in an otherwise fully built environment.

Hydrogeology of the Lake Passaic Basin

The valley-fill deposits of the Lake Passaic basin produce over 15,000 million gallons per year and are the most productive glacial aquifers in New Jersey (Hoffman and Quinlan, 1994). They are the principal or sole water source for several municipal systems and so are classified by the EPA as a sole-source aquifer. The "overramp zone" (fig. 8), where Illinoian and Chatham-stage lacustrine sand and gravel are thick, extensive, and fill the pre-intermediate buried fluvial valleys, is the most productive sector of the aquifer system (see distribution of production wells on fig. 8). The Illinoian and Chatham-stage sands and gravels are overlain by low-permeability silt and clay lake-bottom deposits and till. Till matrix is chiefly silty sand to sandy silt but is of low permeability because it has been highly consolidated in most places by the weight of overlying ice. These materials act as confining or semi-confining layers. Some wells bored through these sediments into underlying sand and gravel flowed at the surface when first drilled, indicating artesian conditions, although pumping in recent decades has greatly lowered the piezometric surface and wells no longer flow (Meisler, 1976; Hoffman and Quinlan, 1994). Gradients on the piezometric surface in the overramped sand and gravel indicate flow toward the Short Hills gap, along the original pre-intermediate fluvial valley gradient, although there is much perturbation of the natural gradient by pumping-induced cones of depression. Recharge to the confined, overramped sand and gravel is by infiltration of precipitation on the delta fronting the terminal moraine, which connects to the overramped sand and gravel in the subsurface (fig. 8, section BB'), by stream loss from the Passaic and Whippany rivers where they flow across the delta-moraine complex, and from vertical leakage of surface water through the lake-bottom sediment and till capping the overramped sand and gravel at and north of the moraine.

Neither Illinoian nor late Wisconsinan ice advanced south of the moraine complex (fig. 8, late Wisconsinan limit), and pre-Illinoian deposits in this area are thin, patchy erosional remnants largely above the level of the valley fill. Thus, no sand and gravel occurs in the valley fills south of the moraine complex. These fills instead consist of late Wisconsinan lake clays and silts deposited atop Illinoian lake clays and silts. These deposits confine the underlying bedrock but provide no water themselves.

North of the overramp zone, Passaic lobe ice was not buoyed by Chatham-stage lake water, since the Millburn gap was not yet blocked by Hackensack ice. The Passaic lobe thus eroded rather than overran any Illinoian stratified deposits on the landscape during advance. In fact, Passaic-lobe ice was so erosive in this sector that it scoured overdeepenings as much as 300 feet deep into shale bedrock in the lowlands between the basalt ridges (fig. 8, late Wisconsinan glacial overdeepenings). The valley fill in this northern sector of Lake Passaic thus consists exclusively of recessional lacustrine sediments on a basal till on bedrock, with no buried pre-advance materials (fig. 8, section BB', north end). The recessional lacustrine deposits are chiefly silt and clay lake-bottom sediments but in places, particularly in the valley fill along the Highland Front, permeable lacustrine-fan sand and gravel underlies the lake-bottom material. These fan deposits supply several municipal well fields in this area (Stop 4 of field trip). The fan gravels were deposited at the mouths of subglacial tunnels as the ice front retreated, and were later buried by lake-bottom sediment that accumulated vertically as fine sediment settled out of the lake. In places the fans rise above the lake-bottom fill and crop out as knolls and ridges (Stop 4 of field trip). In places along the Highland Front they built up far enough to reach the lake surface, where they prograded into the lake to form deltas. Other deltas along the Highland Front were deposited where ice-lateral meltwater channels draining down the Front entered the lake. The outcropping deltas and fans act as recharge conduits for the buried fans, where the deposits are physically connected.

Shallow-water delta sands deposited in the postglacial lakes by meteoric or meltwater streams, especially in the Totowa stage, lie above the thick deposits of lake clay laid down in the Moggy Hollow and Great Notch stages (fig. 8, north end of BB'). These sands may be unconfined aquifers if sufficiently thick, particularly adjacent to rivers, which are in hydraulic connection with the sands. However, no large-capacity wells tap these deposits in the Lake Passaic basin.

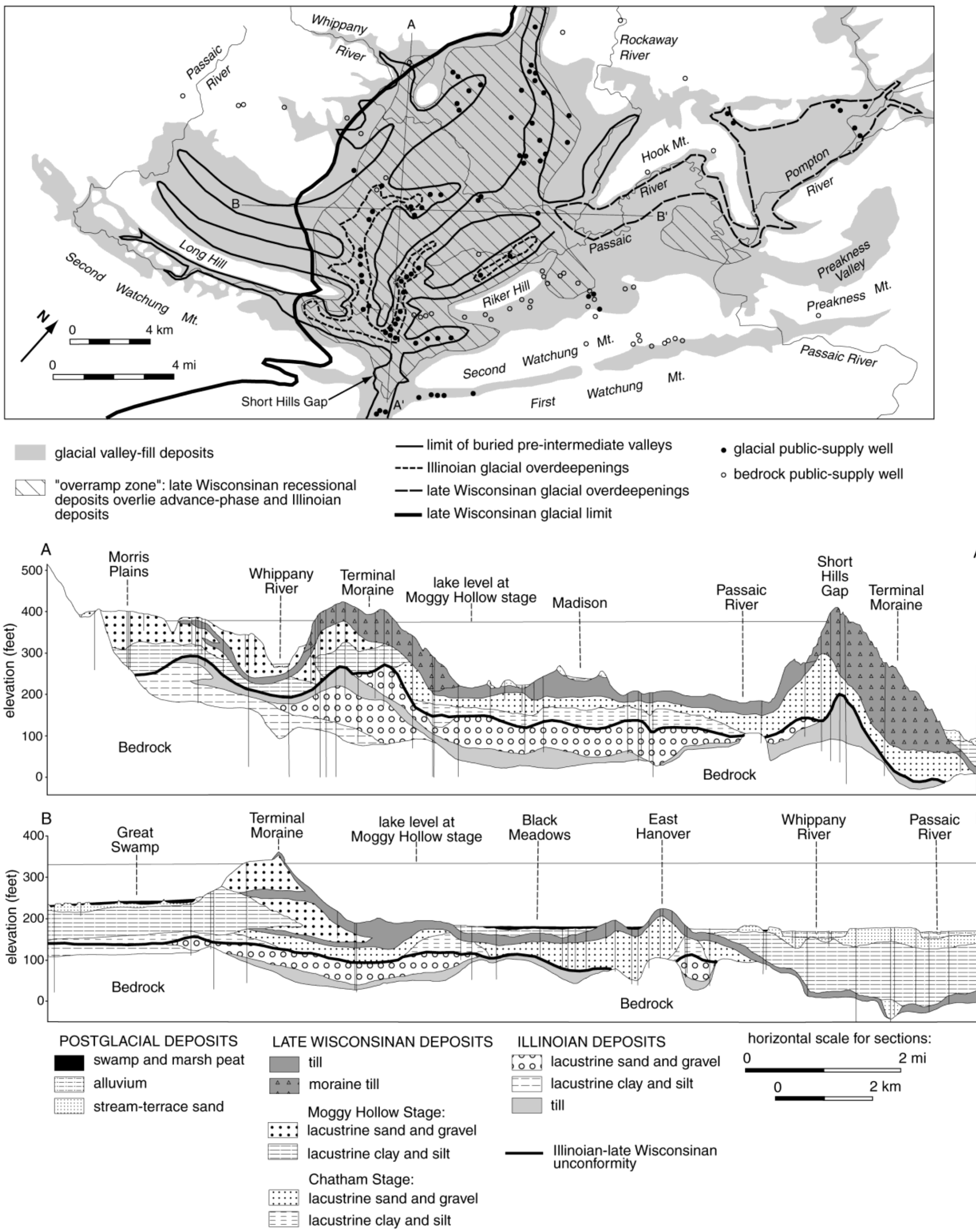


Figure 8. Stratigraphy and geomorphology of the valley-fill aquifer system in the Lake Passaic basin.

Summary

In the Pliocene the Hudson River, and a trunk river from southern New England, drained southwesterly from the New York City area to the Delmarva Peninsula. Drainage dislocations during the pre-Illinoian glaciation in the late Pliocene between 2 and 2.5 Ma redirected the Hudson seaward in the New York City area. Stabilization and incision of this drainage in the early Pleistocene established the lower Raritan and Passaic as tributaries to the Hudson, which exited to the shelf via a now-buried valley in Queens. An intermediate glaciation of probable Illinoian age (150 ka) reached the New York City area. Till and lacustrine and fluvial sediments in the lower parts of valley fills in the upper Passaic basin and on Long Island are deposits of this glacier. The late Wisconsinan glacier, centered as a southwesterly flowing lobe in the Hackensack lowland, arrived at its terminal position slightly after 21 ka and remained there for about 750 years, where it deposited a prominent moraine. It began to retreat from this position at 20 ka. Lake Passaic occupied the upper Passaic basin west of Second Watchung Mountain. It included an advance stage, a maximum stage, and a recessional stage. These stages spanned advance and retreat of the glacier. It drained in two stages eastward into Lake Bayonne. Stacked valley-fill sediments in Lake Passaic are important glacial aquifers. To the east, the terminal moraine formed a dam for glacial lakes during retreat. Lake Bayonne, the earliest lake, was controlled by an eroding spillway on the moraine dam at Perth Amboy. It was succeeded by Lake Hackensack and Lake Albany. Lake Hackensack drained eastward into Lake Albany when Sparkill Gap in the Palisades Ridge was deglaciated. Lake Albany was dammed by the moraine at the Narrows and extended up the Hudson valley as ice retreated. The Narrows dam was breached at 15.5 ka when Lake Wallkill flooded into the Hudson valley. Breaching initiated an unstable phase of the lake, with an eroding spillway first on the moraine, then on the emerging lake bottom in the mid-Hudson valley. Rising sea level flooded the lower Hudson valley after 12 ka, limiting fluvial incision and so providing quasi-stable spillways for the Quaker Springs and Coveville stages of Lake Albany, north of the Albany area, and the upper Fort Ann stage of Lake Vermont, in the Champlain lowland. Outburst floods from draining of Lake Iroquois at 10.9 ka discharged down the Hudson valley and were the last glacial events to affect the New York City area.

References Cited

Antevs, E., 1928, The last glaciation, with special reference to the ice sheet in North America: American Geographical Society, Research Series 17, 292 p.

Averill, S. P., Pardi, R. R., Newman, W. S., and Dineen, R. J., 1980, Late Wisconsinan-Holocene history of the lower Hudson region: new evidence from the Hackensack and Hudson river valleys: *in* Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: field guide of the 52nd annual meeting of the New York State Geological Association, p. 160-186.

Baskerville, C. A., 1994, Bedrock and engineering geologic maps of the New York County and parts of Kings and Queens counties, New York, and parts of Bergen and Hudson counties, New Jersey: U. S. Geological Survey Miscellaneous Investigations Series Map I-2306, scale 1:24,000.

Bayley, W. S., Salisbury, R. D., and Kummel, H. B., 1914, Description of the Raritan quadrangle, New Jersey: U. S. Geological Survey Geologic Atlas, folio 191, 72 p.

Berry, E. W., and Hawkins, A. C., 1935, Flora of the Pensauken Formation in New Jersey: Geological Society of America Bulletin, v. 46, p. 245-252.

Boellstorff, J., 1978, North American Pleistocene stages reconsidered in light of probable Pliocene-Pleistocene continental glaciation: Science, v. 202, p. 305-307.

Buxton, H. T., and Shernoff, P. K., 1999, Ground-water resources of Kings and Queens counties, Long Island, New York: U. S. Geological Survey Water-Supply Paper 2498, 113 p.

Cadwell, D. H., 1989, Surficial geology map of New York, lower Hudson sheet: New York State Museum Map and Chart Series 40, scale 1:250,000.

Cadwell, D. H., and Dineen, R. J., 1987, Surficial geology map of New York, Hudson-Mohawk sheet: New York State Museum Map and Chart Series 40, scale 1:250,000.

Connally, G. C., 1973, Surficial geology of the Glens Falls region, New York: New York State Museum Map and Chart Series 23, scale 1:48,000.

Connally, G. G., and Sirkin, L. A., 1970, Late glacial history of the upper Wallkill valley, New York: Geological Society of America Bulletin, v. 81, p. 3297-3306.

Connally, G. G., and Sirkin, L. A., 1973, Wisconsinan history of the Hudson-Champlain lobe, *in* Black, R. F., Goldthwait, R. P., and Willman, H. B., eds., The Wisconsinan Stage: Geological Society of America Memoir 136, p. 47-69.

Connally, G. G., and Sirkin, L. A., 1986, Woodfordian ice margins, recessional events, and pollen stratigraphy of the mid-Hudson valley, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 50-72.

Cotter, J. F. P., Ridge, J. C., Evenson, E. B., Sevon, W. D., Sirkin, L. A., and Stuckenrath, R., 1986, The Wisconsinan history of the Great Valley, Pennsylvania and New Jersey and the age of the "Terminal Moraine", *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 22-50.

deLaguna, W., 1948, Geologic correlation of logs of wells in Kings County, New York: State of New York, Department of Conservation, Water Power and Control Commission, Bulletin GW 17, 35 p.

DeSimone, D. J., and LaFleur, R. G., 1986, Glaciolacustrine phases in the northern Hudson lowland and correlatives in western Vermont: Northeastern Geology, v. 8, p. 218-229.

DeSimone, D. J., Wall, G. R., Miller, N. G., Rayburn, J. A., and Kozlowski, A. L., 2008, Glacial geology of the northern Hudson through southern Champlain lowlands: Guidebook for the 71st annual reunion of the Northeastern Friends of the Pleistocene, 58 pp.

Dineen, R, J., 1986, Deglaciation of the Hudson valley between Hyde Park and Albany, New York, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York: New York State Museum Bulletin 455, p. 89-108.

Dineen, R. J., and Hanson, E. L., 1983, Bedrock topography and glacial deposits of the Colonie channel between Saratoga Lake and Coeymans, New York: New York State Museum Map and Chart Series 37, 39 p.

Dineen, R. J., and Miller, N. G., 2006, Age and paleoecology of plant fossils associated with the Quaker Springs stage of Lake Albany, and a chronology of deglacial events in the Hudson-Champlain lowlands, New York: Geological Society of America Abstracts with Programs, v. 38, p. 8.

Donnelly, J. P., Driscoll, N. W., Uchupi, E., Keigwin, L. D., Schwab, W. C., Thieler, E. R., and Swift, S. A., 2005, Catastrophic meltwater discharge down the Hudson valley: a potential trigger for the Intra-Allerød cold period: Geology, v. 33, p. 89-92.

Fairchild, H. L., 1919, Pleistocene marine submergence of the Hudson, Champlain, and St. Lawrence valleys: New York State Museum Bulletin 209, 76 p.

Fuller, M. L., 1914, The geology of Long Island, New York: U. S. Geological Survey Professional Paper 82, 231 p.

Groot, J. J., and Jordan, R. R., 1999, The Pliocene and Quaternary deposits of Delaware: palynology, ages, and paleoenvironments: Delaware Geological Survey Report of Investigations 58, 36 p.

Harmon, K. P., 1968, Late Pleistocene forest succession in northern New Jersey: Ph.D. dissertation, Rutgers University, 203 p.

Hoffman, J. L., and Quinlan, J., 1994, Ground-water-withdrawal and water-level data for the central Passaic River basin: N. J. Geological Survey Geologic Report Series GSR 34, 78 p.

Johnson, D. W., 1931, Stream sculpture on the Atlantic slope: Columbia University Press, New York, 142 p.

LaFleur, R. G., 1965, Glacial geology of the Troy, N. Y., quadrangle: New York State Museum and Science Service, Map and Chart Series 7, 22 p.

Lewis, R. S., and Stone, J. R., 1991, Late Quaternary stratigraphy and depositional history of the Long Island Sound basin: Connecticut and New York: Journal of Coastal Research Special Issue 11, p. 1-23.

Lovegreen, J. R., 1974, Paleodrainage history of the Hudson estuary: M.S. thesis, Columbia University, 152 p.

Martino, R. L., 1981, The sedimentology of the late Tertiary Bridgeton and Pensauken formations in southern New Jersey: Ph.D. dissertation, Rutgers University, 299 p.

Meisler, H., 1976, Computer simulation model of the Pleistocene valley-fill aquifer in southwestern Essex and southeastern Morris counties, New Jersey: U. S. Geological Survey Water Resources Investigations 76-25, 76 p.

Merrill, F. J. H., Darton, N. H., Hollick, A., Salisbury, R. D., Dodge, R. E., Willis, B., and Pressy, H. A., 1902, Description of the New York City quadrangle: U. S. Geological Survey Geologic Atlas, folio 83, 19 p.

Miller, N. G., 2008, The Cohoes mastodon and Younger Dryas in eastern New York, in DeSimone, D. J., Wall, G. R., Miller, N. G., Rayburn, J. A. & Kozlowski, A. L., eds., Glacial geology of the northern Hudson through southern Champlain lowlands: guidebook for the 71st annual reunion of the Northeastern Friends of the Pleistocene, p. 19-25.

Newman, W. S., Thurber, D. H., Zeiss, H. S., Rokach, A., and Musich, L., 1969, Late Quaternary geology of the Hudson River estuary: a preliminary report: Transactions of the New York Academy of Sciences, v. 31, p. 548-570.

Obbink, E. A., Carlson, A. E., and Klinkhammer, G. P., 2010, Eastern North American freshwater discharge during the Bølling-Allerød warm periods: Geology, v. 32, p. 171-174.

Oldale, R. N., Cronin, T. M., Valentine, P. C., Spiker, E. C., Blackwelder, B. W., Belknap, D. F., Wehmiller, J. F., and Szabo, B. J., 1982, The stratigraphy, absolute age, and paleontology of the upper Pleistocene deposits at Sankaty Head, Nantucket Island, Massachusetts: Geology, v. 10, p. 246-252.

Owens, J. P., and Minard, J. P., 1979, Upper Cenozoic sediments of the lower Delaware valley and northern Delmarva Peninsula, New Jersey, Pennsylvania, Delaware, and Maryland: U. S. Geological Survey Professional Paper 1067D, 47 p.

Parris, D. C., 1983, New and revised records of Pleistocene mammals of New Jersey: The Mosasaur, v. 1, p. 1-21.

Perlmutter, N. M., and Geraghty, J. J., 1963, Geology and ground-water conditions in southern Nassau and southeastern Queens counties, Long Island, N.Y.: U. S. Geological Survey Water Supply Paper 1613A, 205 p.

Peteet, D. M., Vogel, J. S., Nelson, D. E., Southon, J. R., Nickmann, R. J., and Heusser, L. E., 1990, Younger Dryas climate reversal in northeastern USA? AMS ages for an old problem: Quaternary Research, v. 33, p. 219-230.

Peteet, D. M., Daniels, R. A., Heusser, L. E., Vogel, J. S., Southon, J. R., and Nelson, D. E., 1993, Late-glacial pollen, macrofossils, and fish remains in northeastern USA—the Younger Dryas oscillation: Quaternary Science Reviews, v. 12, p. 597-612.

Rayburn, J. A., Knuepfer, P. L. K., and Franzi, D. A., 2005, A series of large, late Wisconsinan meltwater floods through the Champlain and Hudson valleys, New York state, USA: Quaternary Science Reviews, v. 24, p. 2410-2419.

Rayburn, J. A., Franzi, D. A., and Knuepfer, P. L. K., 2007, Evidence from the Lake Champlain valley for a later onset of the Champlain Sea and implications for late glacial meltwater routing to the North Atlantic: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 246, p. 62-74.

Reeds, C. A., 1926, The varved clays at Little Ferry, New Jersey: American Museum Novitates, no. 209, p. 1-16. .

Reimer, G. E., 1984, The sedimentology and stratigraphy of the southern basin of glacial Lake Passaic, New Jersey: M. S. thesis, Rutgers University, 205 p.

Ridge, J. C., 1997, Shed Brook discontinuity and Little Falls Gravel: evidence for the Erie Interstade in central New York: Geological Society of America Bulletin, v. 109, p. 652-665.

Ridge, J. C., 2004, The Quaternary glaciation of western New England with correlations to surrounding areas, *in* Ehlers, J. & Gibbard, P. L., eds., Quaternary glaciations—extent and chronology, Part II: Elsevier, Amsterdam, p. 169-199.

Ridge, J. C. 2008: The North American glacial varve project (http://ase.tufts.edu/geology/varves), sponsored by the National Science Foundation and the geology department of Tufts University. Calibrated New England varve timescale as of April 2008.

Ridge, J. C., Braun, D. D., and Evenson, E. B., 1990, Does Altonian drift exist in Pennsylvania and New Jersey?: Quaternary Research, v. 33, p. 253-258.

Roy, M., Clark, P. U., Barendregt, R. W., Glasmann, J. R., and Enkin, R. J., 2004, Glacial stratigraphy and paleomagnetism of late Cenozoic deposits of the north-central United States: Geological Society of America Bulletin, v. 116, p. 30-41.

Salisbury, R. D., 1895, Surface geology: report of progress: N. J. Geological Survey Annual Report for 1894, p. 1-149.

Salisbury, R. D., 1902, The glacial geology of New Jersey: N. J. Geological Survey Final Report, v. 5, 802 p.

Salisbury, R. D., and Kummel, H. B., 1895, Lake Passaic: an exinct glacial lake: Journal of Geology, v. 3, p. 533-560.

Salisbury, R. D., and Peet, C. E., 1895, Drift phenomena of the Palisades ridge: N. J. Geological Survey Annual Report for 1894, p. 157-224.

Sasowsky, I. D., 1994, Paleomagnetism of glacial sediments from three locations in eastern Pennsylvania, *in* Braun, D. D., ed., Late Wisconsinan to pre-Illinoian (G?) glacial and periglacial events in eastern Pennsylvania: guidebook for the 57th Friends of the Pleistocene field conference, p. 21-23.

Schwab, W. C., Denny, J. F., Foster, D. S., Lotto, L. L., Allison, M. A., Uchupi, E., Danforth, W. W., Swift, B. A., Thieler, E. R., and Butman, B., 2003, High-resolution Quaternary seismic stratigraphy of the New York Bight continental shelf: U. S. Geological Survey Open-File Report 02-152, http://pubs.usgs.gov/of/2002/of02-152/.

Sirkin, L., 1986, Pleistocene stratigraphy of Long Island, New York, *in* Cadwell, D. H., ed., The Wisconsinan stage of the First Geological District, eastern New York. New York State Museum Bulletin 455, 6-21.

Sirkin, L. A., and Minard, J. P., 1972, Late Pleistocene glaciation and pollen stratigraphy in northwestern New Jersey: U. S. Geological Survey Professional Paper 800D, p. D51-D56.

Sirkin, L. A., and Stuckenrath, R., 1980, The Portwashingtonian warm interval in the northern Atlantic coastal plain: Geological Society of America Bulletin, v. 91, p. 332-336.

Soren, J., 1978, Subsurface geology and paleogeography of Queens County, Long Island, New York: U. S. Geological Survey Water Resources Investigations 77-34, 17 p.

Stanford, S. D., 1991, Surficial geology of the Roselle quadrangle, Union, Essex, and Morris counties, New Jersey: N. J. Geological Survey Open-File Map OFM 8, scale 1:24,000.

Stanford, S. D., 1993, Late Cenozoic surficial deposits and valley evolution of unglaciated northern New Jersey: Geomorphology, v. 7, p. 267-288.

Stanford, S. D., 2002, Surficial geology of the Elizabeth quadrangle, Essex, Hudson, and Union counties, New Jersey: N. J. Geological Survey Open-File Map OFM 42, scale 1:24,000.

Stanford, S. D., 2003, Surficial geology of the Paterson quadrangle, Passaic, Bergen, and Essex counties, New Jersey: N. J. Geological Survey Open-File Map OFM 54, scale 1:24,000.

Stanford, S. D., 2007a, Surficial geology of the Chatham quadrangle, Morris, Union, and Somerset counties, New Jersey: N. J. Geological Survey Open-File Map OFM 69, scale 1:24,000.

Stanford, S. D., 2007b, Surficial geology of the Pompton Plains quadrangle, Morris, Passaic, Essex, and Bergen counties, New Jersey: N. J. Geological Survey Open-File Map OFM 68, scale 1:24,000.

Stanford, S. D., 2008, Surficial geology of the Bernardsville quadrangle, Morris and Somerset counties, New Jersey: N. J. Geological Survey Open-File Map OFM 74, scale 1:24,000.

Stanford, S. D., 2010, Onshore record of Hudson River drainage to the continental shelf from the late Miocene through the late Wisconsinan deglaciation, USA: synthesis and revision: Boreas, v. 39, p. 1-17.

Stanford, S. D., and Harper, D. P., 1991, Glacial lakes of the lower Passaic, Hackensack and lower Hudson valleys, New Jersey and New York: Northeastern Geology, v. 13, p. 271-286.

Stanford, S. D., Ashley, G. M., and Brenner, G. J., 2001, Late Cenozoic fluvial stratigraphy of the New Jersey piedmont: a record of glacioeustasy, planation, and incision on a low-relief passive margin: Journal of Geology, v. 109, p. 265-276.

Stanford, S. D., Ashley, G. M., Russell, E. W. B., and Brenner, G. J., 2002, Rates and patterns of late Cenozoic denudation in the northernmost Atlantic coastal plain and piedmont: Geological Society of America Bulletin, v. 114, p. 1422-1437.

Stone, B. D., and Borns, H. W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: Quaternary Science Reviews, v. 5, p. 39-53.

Stone, B. D., Stanford, S. D., and Witte, R. W., 2002, Surficial geologic map of northern New Jersey: U. S. Geological Survey Miscellaneous Investigations Series Map I-2540-C, scale 1:100,000.

Suter, R., deLaguna, W., and Perlmutter, N. M., 1949, Mapping of geologic formations and aquifers of Long Island, New York: New York State Water Power and Control Commission Bulletin GW-18, 212 p.

Thieler, E. R., Butman, B., Schwab, W. C., Allison, M. A., Driscoll, N. W., Donnelly, J. P., and Uchupi, E., 2007, A catastrophic meltwater flood event and the formation of the Hudson shelf valley: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 246, p. 120-136.

Toney, J. L., Rodbell, D. T., and Miller, N. G., 2003, Sedimentologic and palynologic records of the last deglaciation and Holocene from Ballston Lake, New York: Quaternary Research, v. 60, p. 189-199.

Volkert, R. A., 2010, Bedrock geologic map of the Pompton Plains quadrangle, Passaic, Morris, Bergen, and Essex counties, New Jersey: N. J. Geological Survey Geologic Map Series GMS 10-1, scale 1:24,000.

Wall, G. R., and LaFleur, R. G., 1994, Post-glacial paleochannel development in the Hudson/Mohawk lowlands, New York: Northeastern Geology, v. 16, p. 116-122.

Weiss, D., 1971, Late Pleistocene stratigraphy and paleoecology of the lower Hudson River estuary: Ph.D. dissertation, New York University, 160 p.

Weiss, D., 1974, Late Pleistocene stratigraphy and paleoecology of the lower Hudson River estuary: Geological Society of America Bulletin, v. 85, p. 1561-1570.

Woodworth, J. B., 1901, Pleistocene geology of a portion of Nassau County and Borough of Queens: N. Y. State Museum Bulletin 48, p. 617-668.

Woodworth, J. B., 1905, Ancient water levels of the Champlain and Hudson valleys: New York State Museum Bulletin 84, 265 p.

Worzel, J. L., and Drake, C. L., 1959, Structure section across the Hudson River at Nyack, New York, from seismic observations: Annals of the New York Academy of Sciences, v. 80, p. 1092-1105.

Road Log and Field Stops

0.0	0.0	Turn right on Victory Boulevard from College of Staten Island gate.
0.1	0.1	Turn right on South Gannon Ave. toward I-278 east.
0.4	0.3	Exit left onto I-278 east.
4.5	4.1	Exit right at Exit 15, last before Verrazano Bridge.
4.8	0.3	Left at light at end of ramp onto Lily Pond Ave. toward Bay Street. Lily Pond Ave. becomes School Road around bend.
5.3	0.5	Straight at light into parking lot for Arthur Von Briesen Park. STOP 1.

Stop 1. The Narrows

Walk up park paths over knoll and ridge topography of the terminal moraine to the viewpoint at the bluff top of the Narrows and Verrazano Bridge.

We are standing on the late Wisconsinan terminal moraine, a 1 to 2 mile-wide (1.6 to 3 km) belt of till between 50 and 150 feet (15 and 45 m) thick, with up to 100 feet (30 m) of ridge, knoll, and basin surface relief. Ice arrived at the moraine just after 21 ka (all dates stated in radiocarbon years), and varves counted in the southern basin of Lake Passaic indicate that the ice front stayed in the moraine belt for about 750 years (Reimer, 1984). The height of the headlands here and in Brooklyn indicates that the moraine was once continuous across the mile width of the Narrows at about 50 feet (15 m) in elevation. During late Wisconsinan retreat the Narrows dam and a moraine dam across the Arthur Kill at Perth Amboy, NJ held in glacial Lake Bayonne in the Arthur Kill-Kill van Kull-New York Bay lowland. Lake Bayonne was controlled by an eroding spillway on till at the Perth Amboy dam, and gradually lowered a total of about 50 feet (15 m) as ice retreated. When the retreating ice margin uncovered Hell Gate in the East River (M1 on fig. 3), Lake Bayonne drained eastward into the Long Island Sound lowland and Hell

Gate became the outlet for Lake Albany. The Hell Gate spillway is on gneiss at -30 feet (-9 m). The Hell Gate stage of Lake Albany was thus a stable lake and the shoreline, reconstructed from ice-contact deltas (text fig 6), projects to the Narrows at -70 feet (-21 m), indicating about 120 feet (36 m) of freeboard on the Narrows dam during the Hell Gate stage. The Hell Gate stage expanded northward up the Hudson valley to north of the Highlands, where delta elevations record the onset of an unstable phase of Lake Albany, with declining lake levels recording an eroding spillway (fig. 6). This indicates breaching of the Narrows dam and replacement of the stable Hell Gate spillway with an eroding spillway on moraine till.

Dam breaching was caused by an outburst flood at 15.5 ka from the Augusta stage of Lake Wallkill into Lake Albany. When the north end of Schunnemunk Mountain (and Storm King in the Hudson valley) was deglaciated, the Augusta stage lowered 230 feet (70 m), releasing 6 mi^3 (25 km^3) of water into Lake Albany. About 40% of this volume was probably added suddenly, producing a rapid rise in Lake Albany of about 160 feet (49 m), sufficient to overtop the Narrows dam.

Dam breaching also initiated fluvial erosion of the Hudson shelf valley, which leads from the Narrows to the edge of the continental shelf. The shelf valley was subaerially exposed until about 12 ka, providing 3500 years for fluvial incision of this prominent feature.

The moraine served as the spillway for Lake Albany until ice had retreated upvalley to just south of the Albany area, where delta elevations indicate that the spillway had shifted onto the emerging lake bottom in the Newburgh area, about 60 miles (100 km) up the valley from the Narrows. During its operation, the spillway on the moraine had eroded the dam to an elevation of -200 feet (-60 m). This depth allowed the rising sea to enter the Hudson valley at about 12 ka. From 12 to 10 ka high relative sea level in the lower Hudson valley limited fluvial incision of the lake bottom in the upper valley, providing quasi-stable spillways for the Quaker Springs and Coveville stage of Lake Albany, and the upper Fort Ann stage of Lake Vermont (fig. 6).

Major lake outflows from the Ontario basin discharged down the Mohawk valley and into the Hudson during the Erie Interstade between 14.5 and 14.2 ka, when the Mohawk valley was briefly deglaciated (Ridge, 1997), and again from 12.7 to 11 ka when the Mohawk valley was the outlet for Lake Iroquois (Wall and LaFleur, 1994). These outflows exited to the shelf through the Narrows and contributed to its erosion. When the Covey Hill threshold north of the Adirondacks was deglaciated, Lake Iroquois dropped in two stages to the lower Fort Ann level of Lake Vermont in the Champlain lowland, releasing 170 mi^3 (700 km^3) and 600 mi^3 (2500 km^3) of water at 10.9 ka (Rayburn and others, 2005). These megafloods also exited through the Narrows, and may have triggered the Younger Dryas stadial in the North Atlantic region between 10 and 11 ka. They were the last glacial events to affect the Hudson valley.

5.3	0.0	Proceed straight through light onto School Rd. from parking lot of Von Briesen Park.
5.5	0.2	Straight (on Lincoln Ave.) to I-278 west.
5.6	0.1	Straight on ramp onto I-278 west.
8.0	2.4	Serpentinite outcrop to left in roadcut.
11.2	3.2	Move right for 440 south.
11.4	0.2	Exit right to 440 south (exit 5).
13.8	2.4	Salt marsh along Arthur Kill visible to right. This is the floor of Lake Bayonne.

14.8	1.0	Enter Fresh Kills landfill, cells on both sides of 440. Former offload dock for garbage barges to right at bridge over Fresh Kills tidal creek. Landfill now closed and garbage is exported by truck and train.
16.6	1.8	Exit landfill area.
17.8	1.2	Climb onto an upland on Cretaceous deposits, veneered with till.
18.9	1.1	Exit right to stay on 440 south.
19.5	0.6	Exit right to Arthur Kill Road, exit 1.
19.8	0.3	Turn right at light at end of ramp onto Veterans Road.
20.0	0.2	Turn right at light onto Tyrellan Ave.
20.2	0.2	Turn right at light onto Boscombe Ave.
20.4	0.2	Proceed straight through light onto Page Ave.
21.0	0.6	Move to right lane so as to stay on Page through light.
21.1	0.1	Proceed straight through light on Page Ave.
21.6	0.5	Left at light onto Hylan Blvd.
22.3	0.7	Right into parking lot for Mount Loretto State Unique Area. STOP 2.

Stop 2. Red Bank Bluffs, Mount Loretto State Unique Area

NO SHOVELS, PICKS, OR HAMMERS

This stop requires a 1.25 mile walk along a beach and the possible fording of a shallow (ankle-deep) tidal creek. If you do not wish to do this, please remain on the bus, which will proceed 0.5 mile down Hylan Blvd. to meet us at Lemon Creek Park, at the end of the beach walk.

Walk south on driveway from parking area across the terminal moraine (fig. 9) to the beach on Raritan Bay, then walk northeastward along the beach to the parking lot for Lemon Creek Park, viewing bluff exposures along the way.

These bluffs provide an excellent exposure of the till that forms the terminal moraine. The bluffs were formed by wave erosion that has cut back about 1000 feet (300 m) into the outer edge of the moraine (fig. 9), which is estimated by fitting a smooth curve to the front of the moraine where it comes ashore to the north on Staten Island and to the west in New Jersey. The till exposed here is a compact, reddish-brown silty fine sand to silty fine-to-medium sand, containing 5-15% by volume pebbles and cobbles, and very few boulders. Pebble composition, counted on a total of 622 pebbles aggregated from five sites in the till along the bluff, is 54% red siltstone, 23% white to yellow well-rounded quartz pebbles, 13% gray siltstone and sandstone, 7% gneiss, 1% chert, and 1% purple conglomerate. The red siltstone is from the Passaic Formation in the Newark Basin. The quartz pebbles are from the Pensauken Formation, a

Pliocene fluvial gravel that was widespread in the New York City area before glaciation and is locally preserved in erosional remnants beneath till on Staten Island and in adjacent New Jersey. The gray siltstone and sandstone are from gray beds in the Newark Basin and from Paleozoic sedimentary rock in the Green Pond outlier and Wallkill valley. The gneiss is from Proterozoic bedrock in the Hudson Highlands. The chert is from Paleozoic carbonate rock in the Wallkill valley. The purple conglomerate is from the Green Pond and Skunnemunk formations in the Green Pond outlier, although some may be from Newark Basin fanglomerates. The reddish matrix color is derived from erosion of the red siltstone and sandstone of the Newark Basin. Notably absent or rare are basalt, diabase, and serpentinite, all local bedrock types found within 5 to 20 miles of the bluffs. Glacial flowlines, however, show that ice moving down the Hackensack lowland to this location did not cross those lithologies (fig. 2).

Also observed in the bluff (locations 2 and 4, fig. 9) are lenses or blocks of thinly bedded white and gray silt, fine sand, and coarse sand. The lenses at location 4 are within a deformed body of glaciofluvial sand and gravel, the one at location 2 appears to be within till, although it is situated behind an old sea wall and may have been artificially emplaced. These are glacially thrust blocks of Cretaceous sand and silt of the Magothy Formation, which, along with the underlying Raritan Formation, onlaps the bedrock in the southern half of Staten island and directly underlies the till. The undeformed bedding in the thrust blocks perhaps indicates that they were frozen when glacially entrained.

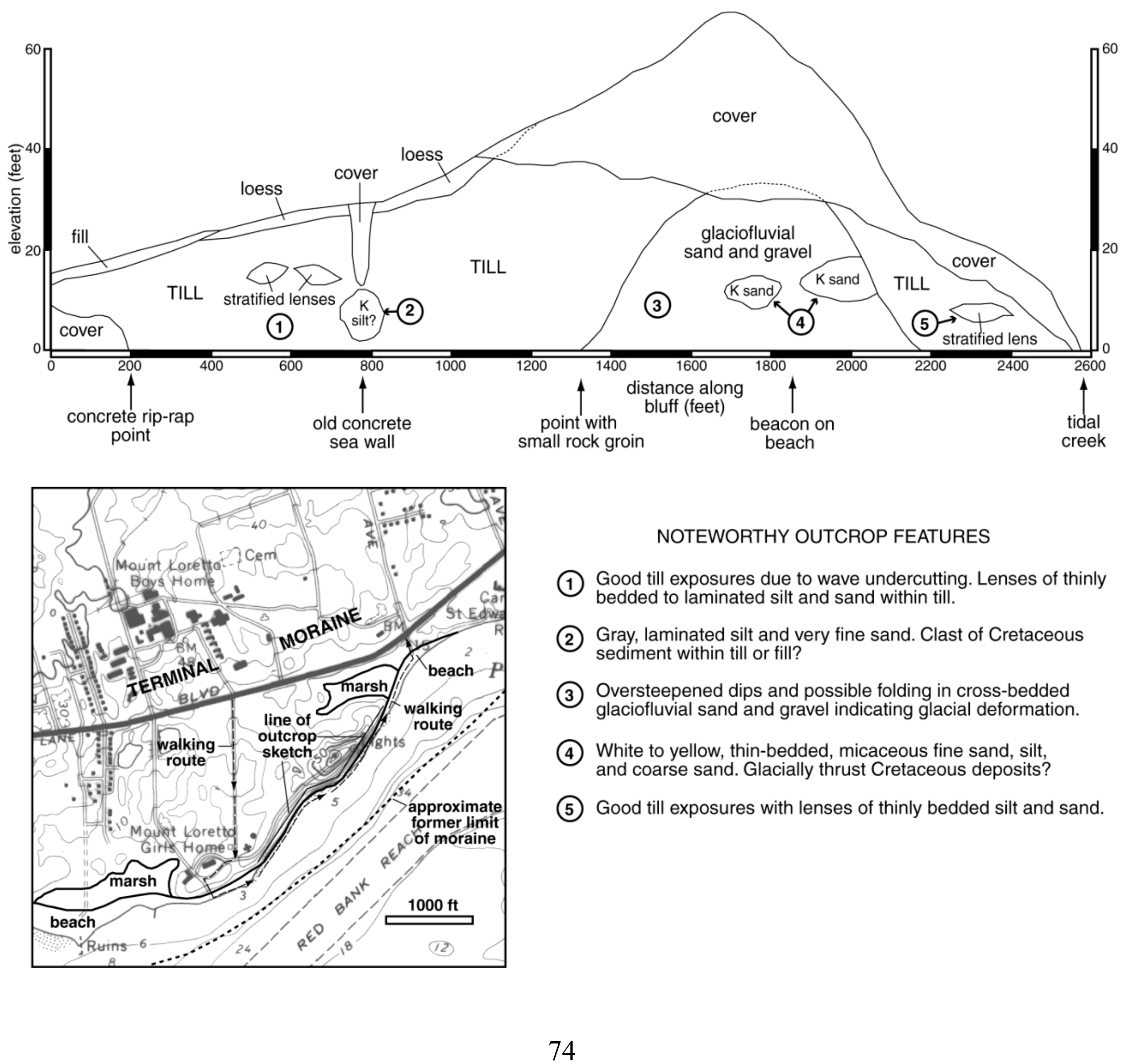


Figure 9. Map and outcrop sketch of Red Bank bluffs, Mount Loretto State Unique area, Stop 2. Base map from USGS Arthur Kill 7.5-minute quadrangle.

The 70-foot hill is cored with light gray to pinkish-gray to very pale brown, cross bedded, clean sand and pebble-to-small cobble gravel, overlain by till. Dips on the cross-bed sets are oversteepened in places (location 3), and recumbent fold axes are also visible on the bluff face in some of the finer-grained beds, indicating that the sand and gravel has been deformed. This gravel is proglacial glaciofluvial outwash that was overrun and deformed by the advancing glacier. The deformation thickened the gravel here to form the core of the hill. Glaciofluvial plains front the moraine when it comes ashore further north on Staten Island, on Long Island, and at Perth Amboy, New Jersey. This deposit is of similar origin. Pebble composition in this unit, counted on a total of 366 pebbles aggregated from three sites in the exposure, is 41% red siltstone, 41% quartz pebbles, 13% gray siltstone and sandstone, 4% gneiss, and 1% purple conglomerate. The greater quartz-pebble content in the gravel compared to the till is from meltwater reworking of Pensauken Formation gravel that was likely widespread on the land surface in front of advancing ice.

Lenses of laminated silt and sand occur within the till at several places (locations 1 and 5). These lenses are about 20 to 50 feet long and generally less than 5 feet thick. They are approximately the same dimensions as the shallow basins on the moraine surface (visible in the meadows on the walk in) and may be swale-pond fills on the moraine surface that were later overrun by readvances as the ice front moved back and forth in the moraine belt. Again, lack of deformation suggests the fills were frozen when overrun.

22.3	0.0	Turn left onto Hylan Blvd. from parking lot.
23.0	0.7	Turn right at light onto Page Ave.
23.6	0.6	Move to right to stay on Page at light.
23.8	0.2	Cross Mill Creek, here a tidal creek. The head of this valley to the east (right) was an early spillway for Lake Bayonne, before it shifted to the moraine dam at Perth Amboy.
24.3	0.5	Turn left at light onto 440 north.
24.6	0.3	Bear right to 440 north.
32.1	7.5	Move left for I-278 west and Goethals Bridge.
32.3	0.2	Exit left for I-278 west and Goethals Bridge.
33.9	1.6	Cross the Arthur Kill. Railroad bridge to right is the Arthur Kill lift bridge, longest center-lift bridge in the world. It was built by the B&O railroad in 1959. Piers for it and the Goethals Bridge are on red shale at a depth of 30 to 40 feet.
34.4	0.5	Bayway refinery on left.
34.6	0.2	Exit left onto NJ Turnpike.
35.3	0.7	Bear left onto NJ Turnpike north after tolls.
38.4	3.1	Newark Airport to left, Port Newark to right. Both are built on former salt marsh atop

		Lake Bayonne lake-bottom clays.
40.2	1.8	Exit right at Exit 14 for I-78 west.
40.7	0.5	Bear left for I-78 west.
41.6	0.9	Bear left onto I-78 west after toll.
41.9	0.3	Bear right to I-78 west.
43.2	1.3	Avoid 2-lane left exit for Clinton Ave., Newark. Rise from lake bottom onto a till on sandstone upland.
46.6	3.4	Cross Garden State Parkway.
50.1	3.5	Move left to stay on I-78 west.
50.6	0.5	Exit left to stay on I-78 west.
51.2	0.6	Roadcut in Orange Mountain basalt, of Lower Jurassic age. This basalt holds up First Watchung Mountain.
52.1	0.9	Roadcut through terminal moraine (behind soundwalls).
53.2	1.1	Cross Blue Brook spillway for Chatham stage of Lake Passaic, here buried by a small outwash plain laid down after the spillway became inactive.
53.5	0.3	Roadcut on right (to 55.6) in Preakness basalt, holding up Second Watchung Mountain.
55.6	2.1	Exit right onto Diamond Hill Road (exit 43).
56.2	0.6	Turn right at light onto McMane Ave.
57.0	0.8	Turn left at light onto 527 north (Glenside Ave.)
57.1	0.1	Proceed straight through light, staying on Glenside.
58.3	1.2	Turn right on 645 south (W. R. Tracy Drive).
58.7	0.4	Cross Chatham-stage sluiceway, occupied by a man-made lake here.
59.1	0.4	Turn right into Watchung Loop picnic area.
59.2	0.1	Comfort station and picnic site. STOP 3.

Stop 3. Watchung Reservation. Lunch and Pre-Illinoian Till.

NO SHOVELS OR PICKS

Before lunch at the picnic area, walk down trail towards Tracy Drive to the head of a gully with a rock-gabion floor, descend the gully to view exposures of till, then exit left at bottom of gully and return uphill on adjoining gravel trail to the picnic area.

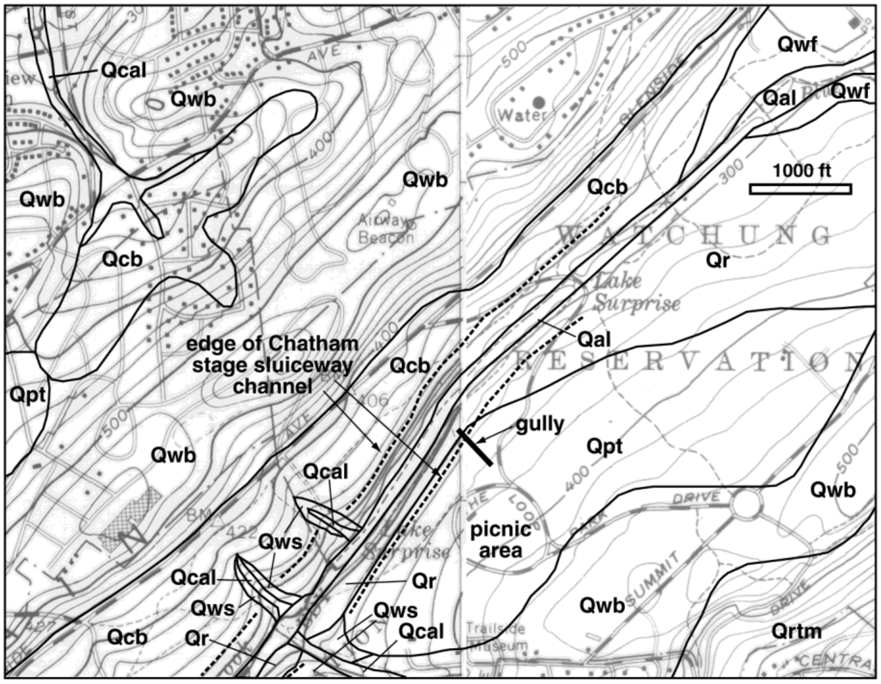


Figure 10. Map of area around Stop 3. Map units are: Qpt=pre-Illinoian till, Qr=late Wisconsinan Rahway Till, Qrtm=Rahway Till of the late Wisconsinan terminal moraine, Qwf=late Wisconsinan glaciofluvial sand and gravel, Qcb=basalt colluvium, Qal=postglacial alluvium, Qcal=alluvium and colluvium, undivided, Qwb=weathered basalt, Qws=weathered shale. Geology from Stanford (1991, 2007a). Base map from USGS Chatham and Roselle 7.5-minute quadrangles.

This gully is eroded into the easternmost deposit of pre-Illinoian till in New Jersey. This deposit is on the gentle dip slope of First Watchung Mountain, just outside the late Wisconsinan limit. Late Wisconsinan till (Qr on fig. 10), laid down by a brief advance beyond the terminal moraine, onlaps and overprints the pre-Illinoian till just to the north. The pre-Illinoian till here is a reddish-brown to reddish-yellow sandy clayey silt with deeply weathered to saprolitized pebbles and cobbles of red and gray siltstone, basalt, and gneiss. Gray quartzite, purple quartzite-conglomerate, black to brown chert, and well-rounded white to yellow quartz pebbles are unweathered. Large cobbles and boulders of gneiss are also intact, perhaps because the outermost weathered parts have spalled off. Fresh exposures observed in 1991 showed a thin tongue of late Wisconsinan till with lightly weathered to unweathered siltstone, gneiss, and basalt clasts overlying weathered micaceous siltstone of the Feltville Formation, at the downstream end of the gully. These materials are no longer well exposed here but can be seen in cuts along the valley side to the south. These observations show that the pre-Illinoian till rests on a gently sloping bedrock bench about 30 feet above the valley floor.

This landscape position is somewhat anomalous for pre-Illinoian till. Generally it is preserved only on flat surfaces on divides or hilltops. The anomaly is explained by the unusual history of this valley. The Blue Brook valley was deepened and widened around 21 ka because it served as the sluiceway (dotted line on fig. 10) for the Chatham stage of Lake Passaic (fig. 7). Before this incision, bedrock-surface elevation contours (Stanford, 1991) indicate that this was a broad divide in the intermontane valley between First and Second mountains, with a much smaller Blue Brook draining southwestward from the divide and another small stream draining northeastward in a valley now buried by the terminal moraine and a small fronting outwash plain. The divide area was protected from erosion, preserving the pre-Illinoian till. The valley here is thus only 21,000 years old, not >800,000 years old like the other valleys in the pre-Illinoian glacial terrain. When the Hackensack lobe of the late Wisconsinan glacier blocked the Short Hills gap, the Blue Brook spillway shut down and the ice front here, no longer held at bay by the rushing lake outflow, briefly advanced into the now-empty sluice, depositing the thin tongue of till observed within the channel.

59.2	0.0	Proceed to right around one-way loop road.
59.5	0.3	Bear left at end of loop onto 645 north (W. R. Tracy Drive).
60.4	0.9	Turn left at stop sign onto 527 south (Glenside Ave.)
61.7	1.3	Turn right at light onto I-78 east.
65.0	3.3	Move right into local lanes of I-78 east.
65.7	0.7	Move left to exit for NJ 24 west.
66.0	0.3	Exit left to NJ 24 west (exit 48).
67.2	1.2	Climb up terminal moraine plugging the former Short Hills gap.
67.8	0.6	Crest of moraine in Short Hills gap. Enter Lake Passaic basin.
69.4	1.6	Cross Passaic River, on floor of Lake Passaic.
71.7	2.3	Roadcuts from hereabouts to 72.2 are in lacustrine-fan deposits laid down in Moggy Hollow stage of Lake Passaic.
73.1	1.4	Black Meadows marsh to right, on floor of Lake Passaic. Terminal moraine forms rise in front. Training facility for NY [sic] Jets football team to left.
75.0	1.9	Roadcut through moraine. Till of the moraine hereabouts caps a section of stacked Illinoian and late Wisconsinan lacustrine sediment about 350 feet thick, one of the thickest glacial sections in NJ.
75.6	0.6	Exit right onto I-287 north.
79.7	4.1	Cross under I-80. Buried pre-intermediate Passaic-Rockaway valley here, filled with 200 feet of stacked late Wisconsinan and Illinoian till and lacustrine deposits.
82.3	2.6	Cross Rockaway River, redirected in postglacial time from the valley at 79.7.
83.6	1.3	Top of Boonton delta, deposited in Moggy Hollow stage of Lake Passaic.
87.5	3.9	Roadcut in another Moggy Hollow stage delta at Brook Valley Road overpass.
87.7	0.2	Outcrops of gneiss on left along Highland Front.
89.9	2.2	Roadcut in yet another Moggy Hollow stage delta, to be discussed at Stop 4.

90.5 0.6 Exit right to NJ 23 south (exit 52A).

91.2 0.7 Exit right for alternate 511 (Boulevard) and then another right onto West Parkway.

91.8 0.6 Turn right into parking lot for Foothills Park. STOP 4.

Stop 4. Lake Passaic Fan-Delta Complex, Foothill Park, Pequannock

Walk from parking lot to the southwest corner of the park lawn, near wall of former gravel pit, to the left of the striated gneiss outcrop. When returning to parking lot, visit the outcrop, which is interlayered white, buff, and light green quartz-plagioclase gneiss and dark gray amphibolite of Middle Proterozoic age. The gneiss and amphibolite show ductile deformation and mylonitic fabric because this outcrop is very close to the Ramapo Fault, the bounding normal fault on the west edge of the Newark Basin (Volkert, 2010).

Foothill Park and the office complex to the east over to NJ 23 are within a former gravel pit. This pit was dug into a lacustrine fan deposited in the Moggy Hollow stage of Lake Passaic. The fan formerly rose as much as 50 feet above the present land surface. The excavated part of the fan is the top of a deposit more than 200 feet thick (fig. 11, section AA'), extending down to a thin basal layer of till resting on bedrock. The pit exposures, and logs of test and production wells adjacent to and within the former pit, show that the fan is pebble-to-cobble gravel and sand with some interbeds of finer sediment, for its entire thickness.

Pebble composition of the fan gravel, counted on a total of 737 pebbles aggregated from 5 sites in the former pit, is 64% gneiss, 31% gray sandstone and siltstone, and 5% purple conglomerate. Gneiss is from the Highlands, adjacent to the deposit to the north and west. Gray sandstone and siltstone are from Paleozoic rock in the Wallkill valley and Green Pond outlier to the north. Purple conglomerate is from the Green Pond outlier. Notable is the absence of red siltstone and basalt from the Newark basin, which underlies most of the fan complex and the broad lowland to the east. Ice flow during glacial advance here, as recorded by striations in the vicinity (fig. 11), including those on the outcrop in the park, was southward from the Highlands into the Newark Basin. Most ice flow from the Highlands into the Newark Basin was blocked by earlier-arriving ice of the Passaic lobe flowing southwesterly in the Basin (fig. 2), so the pattern here is somewhat anomalous. We are near the mouth of the broad, deep, south-trending Wanaque River valley that cuts the Highlands just to the north. This valley helped channel Highlands ice out into the Newark Basin in this area. The Highlands ice deposited gneiss-rich till (Netcong Till, Qn on fig. 11) on the red siltstone bedrock hereabouts. This till also contains significant amounts of gray siltstone and purple conglomerate. Subglacial meltwater feeding the fan complex during deglaciation eroded this till, and gneiss bedrock just to the north in the Wanaque valley, rather than the local red siltstone, accounting for the Highlands provenance of the fan gravel.

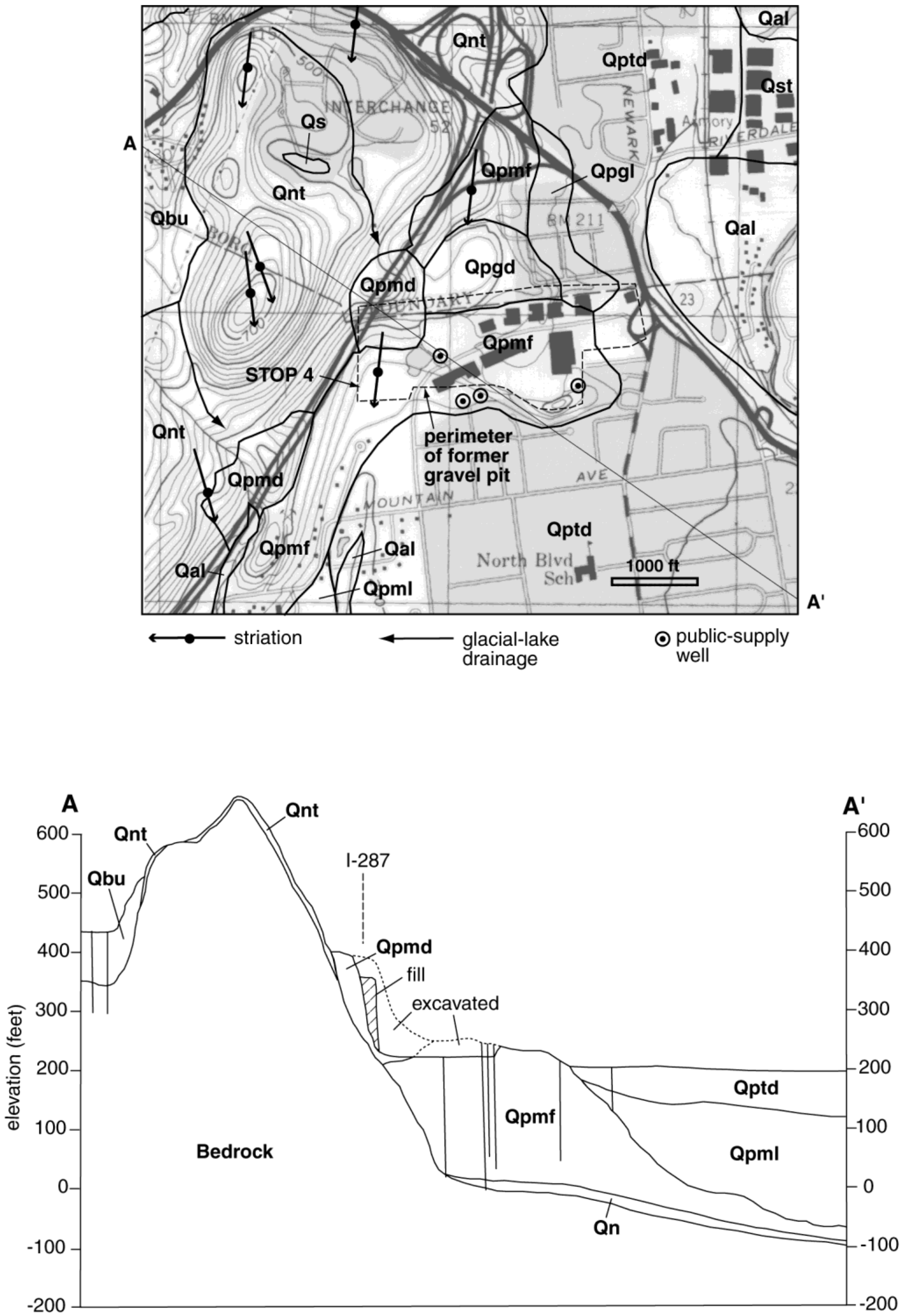


Figure 11. Map and section of area around Stop 4. Map units are: Qpmd=deltaic sand and gravel deposited in the Moggy Hollow stage, Qpmf=Lacustrine-fan sand and gravel deposited in the Moggy Hollow stage, Qpml=lake-bottom silt and clay deposited in the Moggy Hollow stage, Qpgd=deltaic sand and gravel deposited in the Great Notch stage, Qpgl=lake-bottom silt and clay deposited in the Great Notch stage, Qptd=deltaic sand and gravel deposited in the Totowa stage, Qbu=deltaic sand and gravel deposited in Lake Butler, Qnt=patchy Netcong Till and bedrock outcrop, Qn=Netcong Till, Qal=postglacial alluvium, Qst=postglacial stream-terrace sand. Geology from Stanford (2007b). Base map from USGS Pompton Plains 7.5 minute quadrangle.

The northwest edge of the pit was dug into a small Moggy Hollow stage delta. This delta is now largely gutted by the I-287 cut and fill (at mile 89.9 of road log) but a fragment west of I-287 marks its original surface at about 420 feet (128 m) in elevation, indicating that Lake Passaic here was about 500 feet (152 m) deep before deposition of recessional lacustrine deposits (fig. 11, section AA'). The north wall of the pit is dug into a Great Notch stage delta with a top surface at 320 feet (97 m), visible as the treeline on the hill across the park to the north. The pit wall here shows delta foreset sand overlying pebble-to-cobble fan gravel, indicating that the delta prograded over the older fan deposit. Fan deposits laid down in the Moggy Hollow stage also crop out north of the delta (fig. 11), indicating that this delta was not deposited in contact with the glacier. Rather, as the ice front retreated northward from the fan complex toward the present position of NJ 23, an outlet channel for Lake Butler, a local lake that occupied a north-draining valley just to the west of, and higher than, Lake Passaic, was opened at the 500-foot (152 m) elevation on the hill to the west of the fan complex (arrowed line on fig. 11). At the same time, Hackensack lobe ice on the east side of the Lake Passaic basin uncovered Great Notch, and Lake Passaic lowered 80 feet (24 m) from the Moggy Hollow to the Great Notch stage. This allowed the Butler outflow to erode the north end of the Moggy Hollow delta and adjoining fans and redeposit the sediment in the Great Notch stage delta. These relationships fix the geometry of ice margin M1 (fig. 3) between Great Notch and the Highlands.

The thick, permeable fan deposits here are a prolific aquifer. Production wells for Pequannock Township (circled dots on fig. 11) are screened between 120 and 200 feet (37 and 61 m) in depth and, when first drilled, yielded between 700 and 800 gpm. The fan deposits here comprise the full thickness of the valley fill and so are unconfined, but fan gravels are more commonly buried, and confined, by lake clays (Qpml on fig. 11). Such buried, confined fan gravels are the principal valley-fill aquifers in New Jersey.

91.8	0.0	Turn left onto West Parkway from Foothills Park lot.
92.4	0.6	Turn right at stop sign, then left onto turn lane for Newark-Pompton Turnpike, then right onto NJ 23 south at light (requires crossing Newark-Pompton Turnpike).
92.5	0.1	Turn right at light onto NJ 23 south, as described above.
93.2	0.7	Flat plain here is top of a broad shallow-water delta deposited in the Totowa stage of Lake Passaic (fig. 7). Late Wisconsinan overdeepening in this area extends to more than 100 feet (30 m) below sea level; the overdeepening is filled with up to 300 feet (91 m) of recessional lacustrine deposits.
95.7	2.5	Climb from delta onto till hill.
98.0	2.3	Descend onto a stream terrace deposited in the Totowa stage, on top of lake clay.
98.5	0.5	Move left for I-80 east.
98.9	0.4	Proceed straight for I-80 east.
99.1	0.2	Exit left to I-80 east.
101.3	2.2	Roadcut through Great Notch-stage lacustrine fan which was the dam for the Totowa stage.
101.9	0.6	Cross Passaic River, much larger than before because it now includes the Pompton River.

102.4	0.5	Exit right onto Squirrelwood Road (exit 56A).
102.6	0.2	Ramp joins Squirrelwood through light.
103.1	0.5	Turn left, then left again, onto Mountain Ave.
103.4	0.3	Turn right into Garrett Mountain Reservation.
103.5	0.1	Bear right around one-way park loop road.
104.2	0.7	Bear around to left on loop road.
105.1	0.9	Turn right into parking lot for overlook. STOP 5.

Stop 5. Preglacial Drainage, Paterson Gap Overlook, Garrett Mountain Reservation

The Paterson gap in First Watchung Mountain is one of six paired wind gaps through the three Watchung ridges that, together with Sparkill gap in the Palisades ridge to the northeast (visible on the horizon from the overlook), define the preglacial route of the Hudson River in the Pliocene (fig. 1). These gaps are all between 1 and 1.5 mile (2 and 3 km) wide and have preglacial rock floors that decline evenly from 230 feet (70 m) at Sparkill to 180 feet (55 m) at Millburn, the southernmost gap. These rock-floor elevations are on grade with the Pensauken plain, tying them into a dated depositional chronology. This chronology indicates that, in the Pliocene before the pre-Illinoian glaciation, the Hudson was a tributary to the trunk Pensauken River that flowed southwesterly along the inner edge of the Coastal Plain, and discharged to the Atlantic across what is now the Delmarva Peninsula (fig. 1). The pre-Illinoian glacier diverted the Hudson and the Pensauken rivers to the Atlantic across western Long Island, and the gaps were abandoned. After the diversion, in the early and middle Pleistocene, the Passaic established a course that exited the Watchung crescent via the Short Hills and Millburn gaps (fig. 8A). The Illinoian glacier filled the Short Hills gap with till to a height sufficient to divert the Passaic northward through the Little Falls and Paterson gaps, which were not filled, in post-Illinoian time. The Passaic reestablished this route after the late Wisconsinan glaciation, which had further filled the Short Hills gap. Passaic Falls, a 77-foot drop over basalt in the center of Paterson, formed after the late Wisconsinan deglaciation as the river reestablished its course across a new escarpment between the resistant basalt floor of the wind gap and a glacially scoured siltstone lowland to the east. Hydropower from the falls was the original source of industrial development in Paterson, starting in 1791 with the "Society for the Establishment of Useful Manufactures" founded by Alexander Hamilton, which was one of the earliest industrial developments in the United States. The original design of water-power canals and city streets was by Pierre L'Enfant, who also designed the layout of Washington, D. C. Electricity is still generated at the falls.

105.1	0.0	Turn right from parking lot onto park loop road.
105.6	0.4	Turn right to park exit.
105.7	0.1	Turn left at stop sign onto Mountain Ave.
106.1	0.4	Turn left (actually, go straight) at stop sign onto Rifle Camp Road.
108.0	1.9	Turn left onto ramp for US 46 east.

108.4	0.4	Area of the Great Notch spillway, now defaced by road and railroad cuts.
108.7	0.3	Bear right to NJ 3 east.
110.1	1.4	Cross Garden State Parkway.
111.1	1.0	Cross sluiceway eroded by drainage of Great Notch stage around Garrett Mountain.
112.7	1.6	Descend onto flat surface of delta deposited in Lake Delawanna, a local lake held in by a sediment dam in the lower Passaic valley.
113.1	0.4	Cross NJ 21.
113.6	0.5	Cross Passaic River, tidally influenced here.
114.4	0.8	Roadcuts here are in a till on sandstone upland.
115.0	0.6	Cross NJ 17. Descend to the Meadowlands salt marsh, on the floor of Lake Hackensack. A narrow, glacially overdeepened trough extending to more than 300 feet (90 m) below sea level runs along the west edge of the Meadowlands and is filled with lacustrine fan deposits overlain by lake clay.
115.6	0.6	Cross Berrys Creek, move right for exit to NJ Turnpike. Giants Stadium on left, home to the NY [sic] Giants and NY [sic] Jets, and site of the 2014 Super Bowl.
116.1	0.5	Exit right to NJ Turnpike.
116.8	0.7	Bear right onto NJ Turnpike south after tolls.
118.4	1.6	View of Snake Hill, a diabase intrusion into siltstone, to left, and landfill mountains to right, resting on salt marsh atop Lake Hackensack clays.
121.6	3.2	Cross Passaic River. Newark to right, Pulaski Skyway to left.
122.2	0.6	Cross under Pulaski Skyway, completed in 1932. Depth to shale bedrock ranges from 30 to 130 feet along the 4-mile length of the viaduct. Lake clay overlies the bedrock.
129.5	7.3	Exit right at exit 13 for I-278 east and Goethals Bridge.
130.0	0.5	Bear right after NJ Turnpike tolls to I-278 east and Goethals Bridge.
132.2	2.2	Bear left onto I-278 east after Goethals Bridge tolls.
133.2	1.0	Move right for exit 8 to Victory Blvd.
133.8	0.6	Exit right to Victory Blvd. (exit 8).
134.1	0.3	Turn left at light at end of ramp onto Victory Blvd. and then move right.
134.2	0.1	Turn right into entrance to College of Staten Island.

END OF TRIP.

The Timing, Layering, Comagmatic Basalt Flows, Granophyres, Trondhjemites, and Magma Source of the Palisades Intrusive System

by

John H. Puffer[1], Jeffrey C. Steiner[2], Karin A. Block[2], and Alan Benimoff[3]

[1] Department of Earth & Environmental Sciences, Rutgers University, Newark, New Jersey, 07102

[2] Department of Earth and Atmospheric Sciences, City College of New York, 138th and Convent Avenue, New York City, New York 10031.

[3] Department of Engineering Science and Physics, College of Staten Island, 2800 Victory Boulevard, Staten Island, NY 10314

Abstract

Our understanding of the Palisades is currently undergoing considerable revision. New radiometric data is being accumulated showing that the duration of Palisades intrusive activity is much greater than previously estimated. Active research issues include: In-situ fractionation vs. multiple intrusion of pre-fractionated magma pulses; correlation of Palisades layers with Watchung basalt flows; the origin of the olivine layer; and the role of contamination vs. fractionation in the development of the Sandwich Horizon, granophyres, trondhjemites, and syenite bodies within the sill. We will examine evidence pertaining to each of these issues at field trip stops between Staten Island and Nyack Beach State Park.

Preliminary unpublished evidence gathered at these sites and elsewhere indicates that the Palisades was intruded over a prolonged time span by multiple pulses of magma that were variously pre-fractionated at depth. Each pulse underwent limited additional fractionation after emplacement. According to recently published work and unpublished mineralogical evidence, the lower Palisades consists of several layers that correlate with individual Orange Mountain basalt flows while each of several upper Palisades layers correlate with Preakness basalt flows. Recently published geochemical evidence also indicates that trondhjemite dikes and syenite veins near the base of the sill are products of country-rock fusion. In contrast, preliminary unpublished evidence suggests that the siliceous Sandwich Horizon and granophyres layers at central and upper sill positions are the product of complex processes including fractionation. Finally, most geologic evidence supports a Palisades magma source related to decompression -melting triggered by Pangean crustal thinning and rifting along previous plate sutures located over subduction enriched mantle.

Introduction

The Palisades sill has been the subject of considerable geologic literature. Up until the last few years it has generally been thought of as a composite intrusion consisting of an initial thin intrusion followed by a much thicker intrusion of similar composition that underwent considerable in-situ fractionation. Recent and ongoing research has modified this simplistic fractionation model for the Palisades that is presented in many petrology text-books and is making progress toward several additional new or unresolved issues including:

1) Geochronology - the unresolved timing of the intrusion and extrusion of Palisades magma based on the latest geocronological techniques;

2) In-situ fractionation vs. multiple intrusion of pre-fractionated magma pulses - development of a new concept that the Palisades was intruded as multiple pulses of magma (at least eight) that transmitted huge quantities of basalt to the surface from deep sources where fractionation and contamination occurred

before injection into the Palisades. This concept replaces the idea that most chemical variation within the Palisades is due to in-situ fractionation;

3) Correlation of Palisades layers with Watchung basalt flows - the unresolved correlation of individual layers of Palisades rock with overlying basalt flows based on a) geochemical evidence and b) mineralogical evidence;

4) The Olivine Zone - the unresolved origin of the olivine layer. Is it an olivine cumulate (F. Walker (1940), a facies produced as a contact effect during the intrusion of the second magma (Walker, 1969), an intrusion of a separate OLN magma batch (Husch 1989), or the emplacement of olivine and CPX enriched crystal mush from an unspecified source (Steiner et al. 1992).

5) The Sandwich Horizon and Granophyres – what was the role of contamination vs. fractionation in their development?

6) Trondhjemites and syenite – what was the source of trondhjemite melt that has been injected into the Palisades as several thin cross-cutting veins and syenite bearing migmatite found at the base of the sill?

7) The source of the Palisades (CAMP) magma – probably the most controversial aspect of Palisades research.

Each of these issues will be discussed in this guidebook. The field-trip stops that we have selected will display evidence bearing on these issues and will hopefully stimulate lively discussions. But first a brief review of the Geologic setting of the Palisades intrusive system and its position within the ENA and CAMP (Central-Atlantic-Magmatic-Province) classification system.

What's so special about the Palisades?

As a reminder to experienced geologists and as a reason for new students to pay attention is a short list of reasons why the Palisades is quite special. 1) The Palisades is part of one of the largest (if not the largest) magmatic events since the Precambrian. It is part of a LIP (large igneous province) that covered large parts of Europe, Africa, North America, and South America with a thick layer of volcanic rock. 2) The Palisades intruded into a rift system associated with the early stages of the opening of the Atlantic Ocean. The separation of Africa from North and South America began with Palisades magmatism and continues to this day at the mid-Atlantic-ridge. 3) The igneous rock that flowed through the Palisades and related intrusions (at or near the Triassic-Jurassic boundary) is responsible for one of the most important mass extinctions on earth; rivaled only by the effects of the meteor impact at the Cretaceous – Tertiary boundary and Siberian volcanism at the Triassic – Permian boundary (and possibly some early Paleozoic extinctions). About 23% of all families and 48% of all genera went extinct. Most large land animals were eliminated, leaving dinosaurs with little serious competition.

Geologic Setting

The Palisades Intrusive System consists of a composite sill together with a network of thinner co-magmatic sills and dikes exposed within the Newark Basin of New Jersey and New York. The Palisades sill portion of the system (Fig. 1) is mapped by Drake et al. (1996) as medium- to coarse-grained subophitic diabase to coarse-grained quartz-rich to albite-rich granophyre. The diabase is composed mainly of plagioclase (An50-70), clinopyroxene (mostly augite), orthopyroxene, magnetite, and ilmenite with accessory apatite, quartz, alkali feldspar, hornblende, titanite, zircon, and olivine. The Palisades sill is as much as 360 to 400 m thick (Drake et al. 1996) but is typically just over 300 m thick.

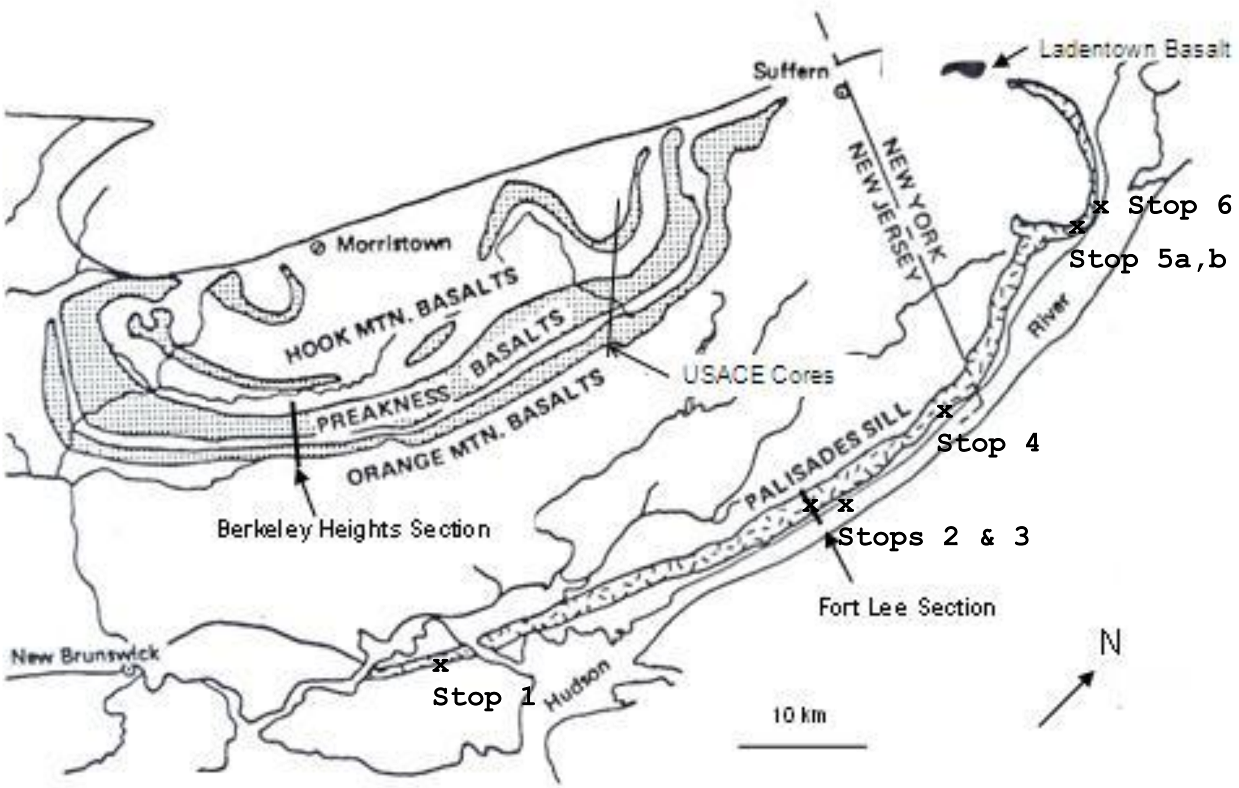


Figure 1. Distribution of early Jurassic igneous rocks throughout the northern Newark basin indicating the locations of data sources used by Puffer et al (2009) including the Berkeley Heights section, the US Army Corps of Engineers (USACE) drill-cores, the Fort Lee section, and Ladentown basalt exposures. Stop 1 is a partially fused xenolith exposure at Graniteville, Stop 2 is the olivine zone, Fort Lee, Stop 3 is a migmatite and trondhjemite exposure at the base of the sill near Ross Dock, Stop 4 is an exposure of cm-scale rhythmic banding at Alpine, Stop 5 is the Sandwich Horizon and some granophyres, and Stop 6 are exposures of the upper Palisades contact and some unusual rock types.

The Palisades system includes 16 additional early Jurassic diabase intrusive sills or sheets in the Newark Basin exposed west of the Palisades sill that have been identified by Gottfried et al. (1991a,b) as co-magmatic on the basis of similar chilled margin compositions. Husch (1992a) concluded that these 17 sheets constitute a single Palisades – Rocky Hill – Lambertville "megasheet" extending about 150 km from southern New York to Pennsylvania. He has shown that although the chill zone of each individual exposed portion of the megasheet is compositionally identical, the interior portions are highly variable. Steiner et al. (1992) use a cumulus-transport-deposition model to show that the Palisades responded to varying degrees of crystal settling, in situ crystallization, flow differentiation, and magma recharge resulting in distinct along-strike variations.

The intrusion of the first pulse of Palisades magma occurred at or very close to the Jurassic/Triassic boundary although the exact timing of the intrusion sequence is the subject of current research.

ENA and CAMP Classification of Palisades Magma

Weigand and Ragland (1970) were the first to recognize that eastern North American Mesozoic dolerite dikes plot as four distinct populations on TiO_2 – MgO – FeO_t variation diagrams. He named these the HTQ-type (high titanium quartz normative), the LTQ (low-titanium quartz normative), the HFQ (high-iron quartz normative) and OLN (olivine normative) magma types. Puffer et al (1981) applied this classification to the Watchung basalt flows of New Jersey and determined that the lower flows (Orange Mountain Basalt) are HTQ-type, most of the middle flows (Preakness Basalt) are LTQ-type and that the upper flows (Hook Mountain Basalt) are HFQ-type. Despite FeO contents below the range typical of HFQ, Tollo and Gottfried (1992) determined that the lower two of five Preakness flows qualify as HFQ-type on the strength of REE correlations with type sections. Gottfried et al (1991a,b) also found that chill-zone samples from each of the sills and dikes that make up the Palisades Intrusive System are HTQ-type. A series of papers by Puffer and others (1980-98) extended the HTQ-HFQ-LTQ classification to other ENA basalt formations including the Hartford, Culpeper, and Fundy Basins and to the Argana basin of Morocco and the Algarve basin of Portugal. Finally Marzoli and others (1999) expanded the classification to include Brazil, French Guiana, Surinam, Guyana, and parts of north-west Africa.

As the breadth of Ragland's applications increased Salters and others (2003) proposed these revisions: 1) LTi (low Ti) to now include the original OLN and some LTQ types and most CAMP dikes; 2) ITi (intermediate Ti, 1-1.5% TiO_2) to encompass HTQ-type and most CAMP basalt flows; and 3) HTi (high Ti) that links highly evolved basaltic rocks with 4-5% TiO_2, not common in North America, with similar varieties in Africa and South America. Note that some of the original LTQ and HFQ groups are included with the ITi group. However for purposes of distinguishing among Palisades and Watchung magmas the original Ragland nomenclature remains valid.

1. Geochronology

Dunning and Hodych 1990 on the basis of U/Pb dating of accessory minerals in the Palisades sill suggest intrusion occurred at ~201 Ma. This date agrees with $^{40}Ar/^{39}Ar$ dates (Hames et al., 2000) of plagioclase separated from the Orange Mt basalt (~201 Ma) but is 2 million years older than plagioclase from the Hook Mountain basalt (~199 Ma). However the analytical error is on the order of +/- 1.5 Ma. Much more precise techniques are available but depend on the occurrence of zircon. Blackburn et al (2009) have found zircon in the Preakness basalt and will shortly be publishing dates with analytical errors of +/- 0.2-0.25 Ma. In addition, a group led by Marzoli (in prep.) have new precise radiometric dates on Palisades and Watchung samples that will soon be available. These geochronological data have important implications for defining theTriassic-Jurassic boundary and the related mass extinction event caused by CAMP. In addition they bear directly on the length of time required for sills the size of the Palisades to fully evolve through complex multiple intrusion and crystallization steps.

If the Palisades sill was the conduit through which both Orange Mountain and Preakness basalts were extruded as proposed by Puffer et al (2009) the Palisades must have taken at least evolved for at least 260 Ky. Olsen et al. (2003) have shown on the basis of carefully measured Milankovitch cyclicity that the Feltville Formation separating the Orange Mountain basalt from the Preakness basalt represents about 260 Ky years of deposition. Puffer et al (2009) propose that intermittent magmatism (including Ladentown basalt extrusion) and hydrothermal activity kept a pathway unconsolidated and open to subsequent magma intrusions during the 260 Ky of Feltville deposition. Continuous pulse influx would also erode, assimilate, and eliminate any previous chill zones or any rock altered by previous vapor vents.

2. Correlation with Watchung Basalts

Part 1 Geochemical evidence

Puffer et al (2009) interpreted the Palisades sill as a progressively-inflated conduit for outpouring huge volumes of flood-basalt. The geochemical data are consistent with a Palisades structure fed by three compositionally distinct intrusion events. The first magma flowed through the sill and erupted near the northern terminus as three Orange Mountain basalt flows. Each of the three extrusive pulses is linked to discrete facies in the lower 150 m of the sill based on distinct geochemical reversals in vertical compositional trends. The end stage of each pulse is characterized by pyroxene phenocryst accumulations.

Magma from a second source inflated the sill by an additional 170 m after approximately 260 Ky of minor intermittent extrusive and intrusive igneous activity and sediment deposition (the Feltville Formation) onto the Orange Mountain basalt}. The second magma extruded as the highly fractionated 150 m thick Preakness basalt and comprises the central layer of Palisades diabase of similar composition. Subsequent extrusions of relatively thin Preakness flows (Magma 3) correlate with upper layers of the Palisades sill (Fig. 2).

Puffer et al (2009) provide geochemical evidence, including Cr distribution (Fig. 2), to document facies boundaries. Each of the Orange Mountain Basalt pulses is correlated to facies in the lower half of the Palisades sill (Fig 2). In addition, each Preakness basalt flow pulse correlates with representative diabase layers within the upper half of the sill (Fig. 2).

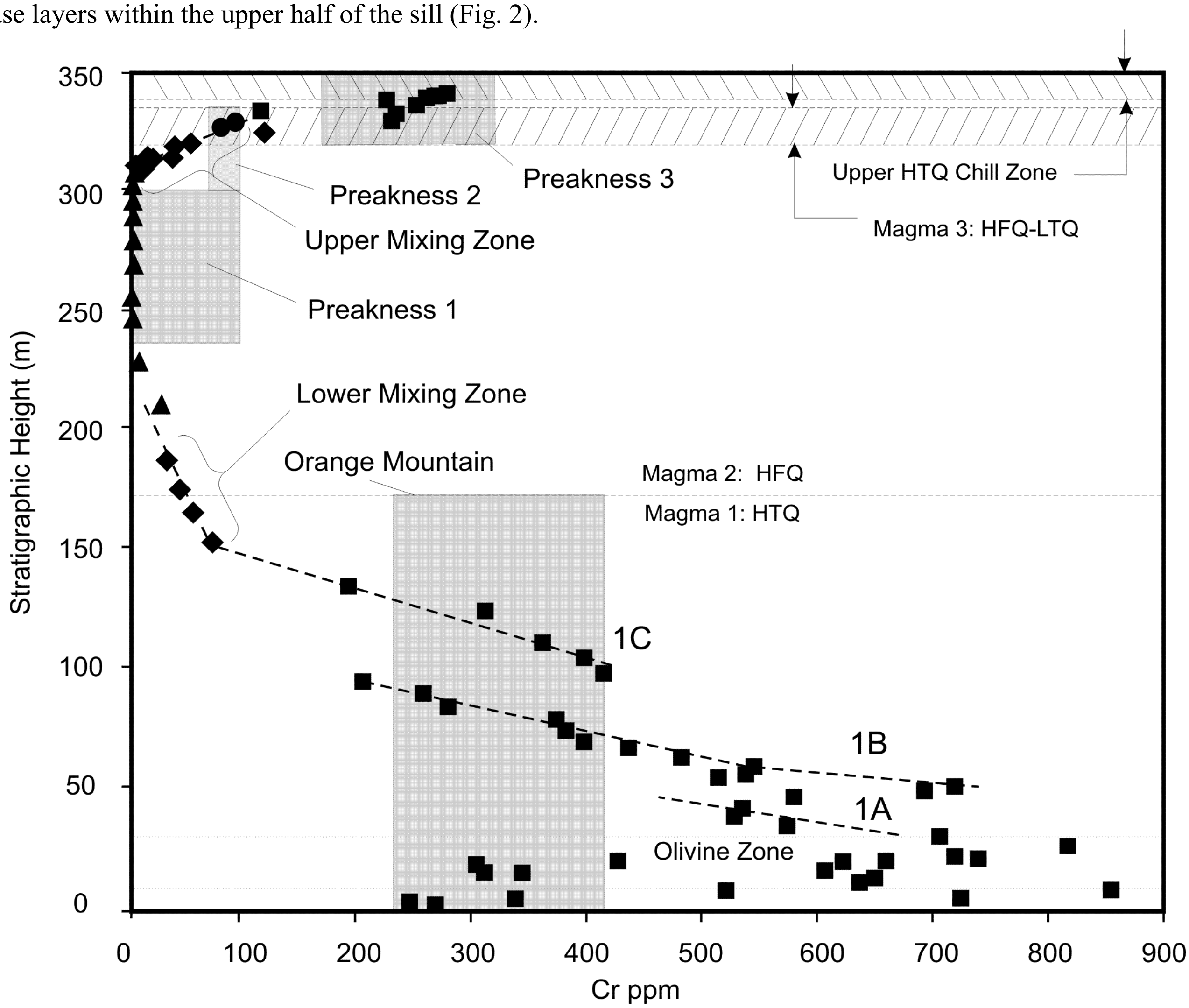


Figure 2. Cr content of the Palisades sill as a function of stratigraphic height (data points) together with Orange Mountain and Preakness basalt flow thickness and Cr ranges plotted as gray fields. Olivine zone (Hysalosiderite; Walker, 1969); 1A Pulse 1A of Magma 1; 1B Pulse 1B of Magma 1; 1C Pulse 1C of Magma 1; OMB Orange Mountain Basalt coeval with Magma 1; Preakness flow 1 coeval with Magma 2; Preakness flow 2 and Preakness flow 3 coeval with Magma 3. HTQ High Titanium Quartz Tholeiite; HFQ High Iron Quartz Tholeiite; LTQ Low Titanium Quartz tholeiite.

Part 2 Mineralogical evidence

Steiner, et al. (1993) have proposed that the 'olivine' layer is a restructured member of a disarticulated magma chamber formed at depth within the Palisades system. Although extensive compositional correlations have been outlined, there is very little comprehensive evaluation of either the series of petrographic facies within the diabase sheet or members of the correlative Watchung flow members. Steiner et al (2009) have established correlations with facies along the basal 50 meters of the Palisades with a comparable section of the Orange Mountain Basalt. For example, facies identified by Walker (1969) as chilled dolerite and early dolerite facies near the base of the sill contain~~s~~ glomeroporphyritic aggregates of augite, and orthopyroxene surrounded by reaction rims of altered olivine. These glomeroporphyritic aggregates are virtually identical in appearance to those that characterize the Orange Mountain Basalt.

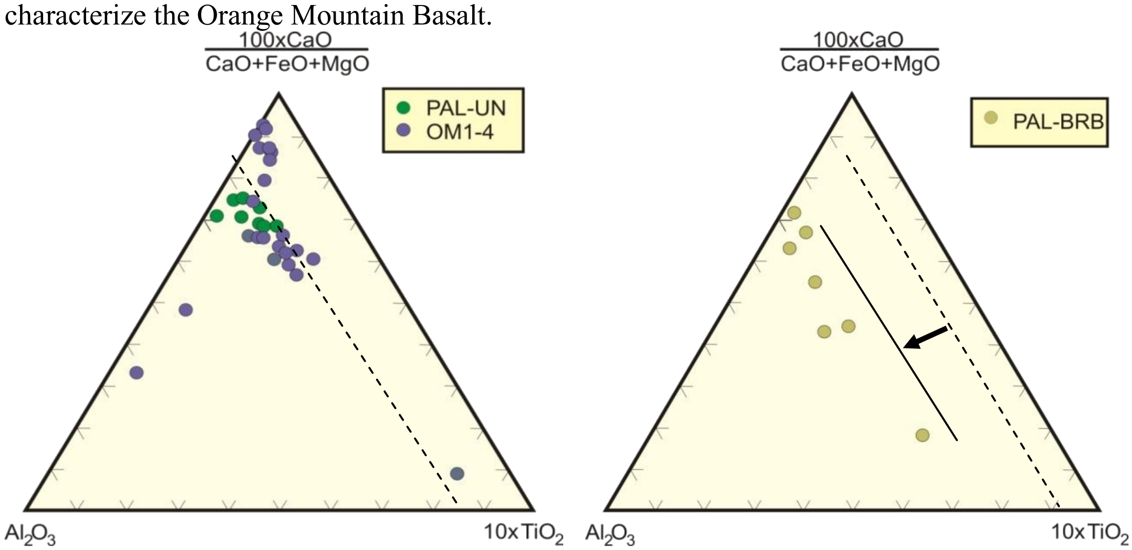


Figure 3. Diagram comparing (A) the augite composition from the lower 40 M of the Palisades with augite from the first of three Orange Mt basalt flows and with (B) augite from 45 m above the base of the Palisades. Note the distinct increase in alumina content.

Augite Chemistry - Steiner et al (2009) have shown that the Al content of the augite fraction in the Palisades sill shows a break at 50 m which is consistent with the bulk chemistry and x-ray diffraction data (Puffer et al, 2009). The variation in Ca ratio-Fe-TiO_2 in the lowermost OMB is consistent with the lowermost 40+ m in the Palisades and shows a stepwise increase from 15 relative wt. % Al at to 17 wt% at 45 m and then again conforms to the Palisades at 50+ m (Fig. 3). Correspondence in major element chemistry, petrographic fabric, the noted x-ray discontinuity and differences in augite chemistry support the conclusion that the Palisades acted as a feeder to the Orange Mountain Basalt and that the lowermost 50 m represents Pulse 1 of Magma 1.

3. In-situ Fractionation vs. Multiple Intrusion Pulses

Puffer et al (2009) interpret the distinct layering of the Palisades sill as injections of magmas that were largely pre-fractionated at deeper levels and then modified to varying degrees by in-situ processes. Most previously proposed Palisades models rely on transporting large quantities of pyroxene and plagioclase crystals long distances by settling or convection from one internal crystallization front across a high temperature interior to the opposite crystallization front. Injection of thinner largely pre-fractionated pulses removes these advective changes as an issue in the evolution of the Palisades (simplify these problems). In addition, Steiner and others (1992) have demonstrated though mass considerations that the cumulus composition of the lower Palisades is insufficient to produce a mass balance with the upper Palisades. Intrusion of magmas from independent sources as proposed by Puffer et al (2009) solves this paradox.

4. The Olivine Zone

A group including Steiner, Brock, and Puffer are currently working on a project pertaining to the olivine zone. The Steiner group has found enclaves of coarse-grained rock in an exposure of Orange Mountain Basalt at Garret Mountain, Paterson. The coarse grained rock is significantly more mafic than typical Orange Mountain basalt. This mafic rock may have been carried as xenoliths (Orange MT) and as a partially solidified magma mush (the Olivine Zone) from a deep source. In both cases it is unlikely that the mafic rocks represent shallow or in-situ olivine or pyroxene accumulation during fractionation.

5. Granophyres and the Sandwich Horizon

A group led by Karin Block is currently revising models for the development of granophyres within the upper Palisades sill and the 'sandwich horizon' of Shirley (1987). Walker (1969) describes silicic facies for the Englewood Cliffs section that occur approximately five kilometers to the north of Shirley's (1987) George Washington Bridge Section. These include fayalite granophyre, ferrodolerite, pegmatite dolerite, and granophyric dolerite, all residing at various stratigraphic levels. These facies are often juxtaposed to less silica-enriched rocks, consistent with a pulse model involving a somewhat chaotic mixing of magmas of possibly differing petrologic heredities. Of present interest are granophyres of Shirley's (1987) sandwich horizon which include Walker's (1969) granophyric dolerites.

The Palisades sandwich granophyres are texturally distinctive, late-stage silica and iron-enriched differentiation products (56-63 wt. % SiO_2; 14 to 16% FeO total), sharing features with the well-studied plagiogranites of mid-ocean ridge systems (Koepke et al. 2007). It may be argued that at least some granophyres in the Palisades, particularly those close to the upper and lower sill contacts, derive from felsic country rock assimilated into the magma during intrusion. This is supported by evidence for the partial melting of country rock at the lower contact with the Palisades (Benimoff and Puffer, 2000; Benimoff, Puffer and Sclar, 2000; Benimoff and Puffer, 2001, Benimoff and Puffer, 2007). We will examine this evidence at field trip Stop 3.

However, isotope evidence generated by the Block group indicates that the Palisades granophyres of the sandwich horizon are not primarily derived through assimilation of country rock. Neodymium isotope systematics unequivocally demonstrate that if country rock was eroded and incorporated into the Palisades magma, the amount of assimilated material was insufficient to influence magma composition to any significant extent. The new isotopic results demonstrate that granophyres from the 'sandwich horizon' of the Palisades Sill are products of magmatic differentiation and provide no evidence of a mixing relationship between sedimentary country rocks and ordinary diabase. Notably, the Nd-isotope composition of a diverse suite of Palisades rocks is nearly constant. Therefore, a multiple-pulse model for the petrogenesis of the Palisades sill requires a homogeneous magma source in terms of Nd-isotope composition. Nd and trace elements show that the Palisades are generally enriched and support Puffer's (2003) hypothesis that CAMP magmas may be products of a reactivated back-arc source.

The absence of a positive Eu anomaly in the Palisades rocks analyzed in this study contradicts compaction and in situ differentiation models proposed by Shirley (1989) and Philpotts et al. (2000). Nonetheless, a significant change in the trace element stratigraphic profile of the Fort Lee section that culminates in a maximum concentration in incompatibles at the sandwich horizon granophyres, is consistent with the evolution of a pre-differentiated second magma (Puffer et al. 2009) and provides adequate conditions for the sill to remain fluid over extensive time periods. This supports a recently proposed model of the Palisades as a composite intrusion that served as a feeder to the extrusive Watchung flood basalts west of the sill.

6. Trondhjemite and Syenite Intrusions

A series of papers by Benimoff, Puffer, and the late Charles Sclar have concluded that a network of trondhjemite intrusion into the Palisades sill are the result of fusion of Lockatong argillite and subsequent penetration into late joint systems. Occurrences of both Lockatong xenoliths and Lockatong beds in contact with the lower contact of the Palisades sill have been described by the Benimoff-Puffer-Sclar group and will be examined at field-trip stops 1 and 3.

In addition to trondhjemite, some syenite melt was generated at the lower contact of sill and dike members of the Palisades Intrusive System. Benimoff and Puffer (2005) conclude that the composition of the melt product depends largely on 1) the composition of the sediment that was melted and 2) on metametasomatic reactions that occurred before fusion. Figure 4.

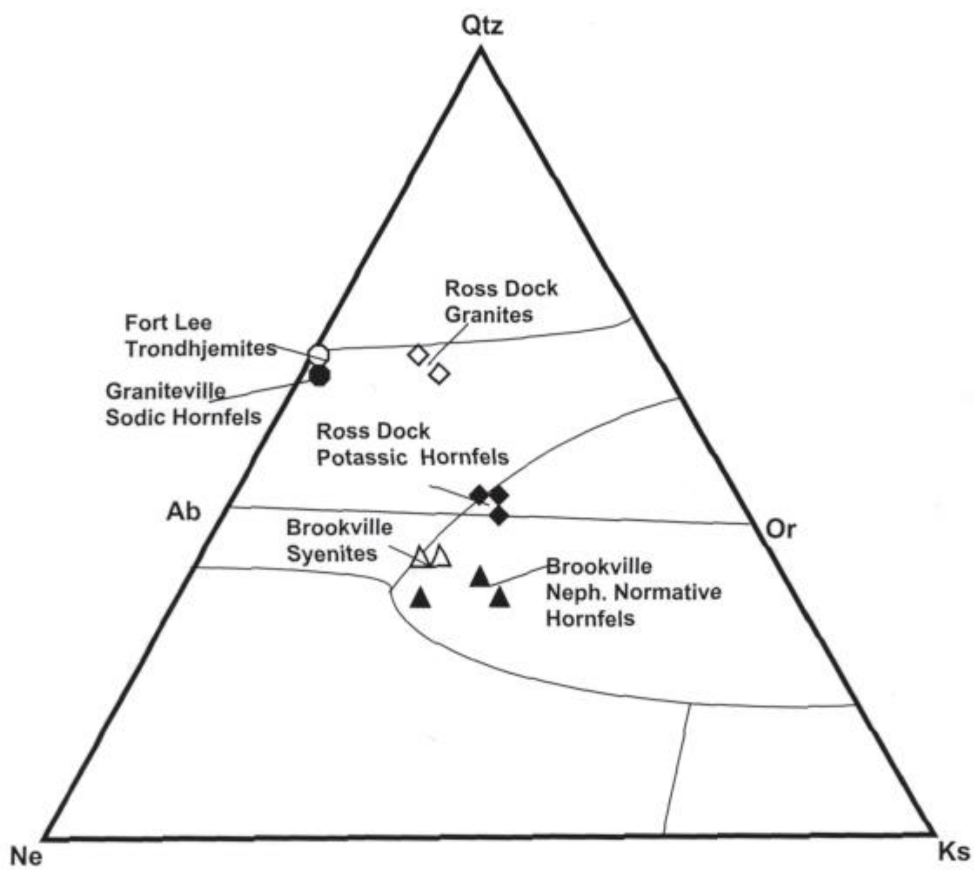


Figure 4 Quartz-Nepheline-Kalsilite phase diagram at 1 kilobar with trondhjemite, granite, syenite compositions together with their respective hornfels sources. Trondhjemites from Fort Lee (open circles) plot close to the eutectic point along the Ab-Qtz boundary. Granite from the George Washington Bridge site plot close to the thermal valley of the Quartz-Albite- Orthoclase system and syenites from Brookville plot close to the Albite-Orthoclase-Nepheline-Kalsilite minimum.

7. Palisades Magma Source

Most of the controversy pertaining to the source of Palisades (and by association CAMP magma) deals with the likelihood of a plume (hotspot) source vs. a fissure source related to the rifting or early brakeup of Pangea. Puffer (1992, 2001, and 2003) argues that decompression melting related to Pangean crustal thinning and rifting close to previous plate sutures is more consistent with the geologic evidence than plume or superplume magmatic activity.

Tectonic evidence for a non-plume source includes the fact that there is no hotspot tract. Instead, early CAMP magmatism, (HTQ or ITi-type) including the Palisades, extruded and intruded, along plate boundaries, and in most cases the same plate sutures associated with the Paleozoic assembly of Pangea. The underlying mantle source was, therefore, enriched during previous subduction.

Geochemical evidence includes the fact that initial Palisades magma resembles calc-alkaline or ARC-type magma in most respects including distinct negative Nb, and Ti anomalies on spider diagrams and a negative slope parallel to standard ARC magma Puffer (2001). In contrast Plume-type or Ocean Island Basalt is relatively enriched in REE's and high field strength elements. Additional geochemical evidence is provided by Rb, Sr, Nd and Sm isotopic data (Puffer, 1992; Pegram, 1983) that indicates early CAMP (Palisades) magma derivation from a source distinct from MORB but consistent with an enriched subcontinental mantle Strong support for a subduction enriched mantle source for HTQ magma is provided by Dorais and Tubrett (2008). They analyzed the cores of strongly zoned clinopyroxenes and found up to 1 wt% Cr_2O_3 and other data indicating that they are early crystallizing phenocrysts, not xenocrysts. Extended rare earth element diagrams for Cr-rich liquids in equilibrium with these cores show enrichment in incompatible elements similar to arc basalts. Their data indicate that the mantle source experienced subduction zone fluid metasomatism that was later modified by crustal contamination.

Puffer (2003) has also argued that an ARC source is consistent with a reactivation of the same magma source responsible for the Paleozoic andesites that extruded along advancing plates during the assembly of Pangea. This ARC source remained dormant until decompression associated with Pangean rifting. Finally, most CAMP magma extruded suddenly and ended quickly with relatively minor late magmatism unlike the prolonged several million year duration of typical plume and plume-tail magmatism.

Some of the current research that Steiner, Block, and Puffer are engaged in, particularly the micro-probe analyses of pyroxene and plagioclase phenocrysts in Orange Mountain and Palisades samples will add additional insight into magma source theory.

Field Trip Stops:

Stop 1. Xenolith fusion (trondhjemite) at Graniteville, Staten Island

(Forest Ave. between Van Name and Simonson Avenues south side), (Latitude 40.624689° ; Longitude -74.153697)

At this stop we will examine a xenolith of argillaceous Lockatong Formation enclosed in Palisades diabase that has undergone partial fusion to yield a trondhjemite. The sodium-rich slab-like xenolith of argillite is 30 m long,, 0.5 m thick and strikes N 30° E with a vertical dip. The xenolith was derived from Lockatong located below the sill. Completely surrounding the xenolith is a coarse-grained trondhjemite. Chemical analyses (Benimoff and Sclar, 1984) reveal that there was diffusion of ions across the liquid/liquid interface between diabase and trondhjemite melts. Fe, Mg, and Ca diffused into the trondhjemite, whereas Na, Rb, K, Ba and diffused into the diabase.

The diabase is composed dominantly of plagioclase ($An_{61}Ab_{38.8}Or_{0.2}$) and augite ($En_{34-44}Fs_{17-31}Wo_{35-42}$). The augite contains exsolution lamellae of pigeonite on (001), and typically exhibits simple contact twinning on (100). A granophyric intergrowth of quartz and K-feldspar is present in minor amounts. Grains of titanomagnetite with oxidation lamellae of ilmenite and discrete grains of ilmenite are common.

The trondhjemite is composed dominantly of quartz-albite granophyre containing large discrete crystals of albite and Ca-rich pyroxene. Minor constituents include interstitial calcite, titanite, ilmenite, optically homogeneous titanomagnetite, nickelian and cobaltian pyrrhotites, apatite, and sphalerite. The modal mineral percentages are clinopyroxene 38, albite 38, quartz 18, titanite 2.7, calcite 1.3, and opaques 2.0.

Petrographic examination shows that the xenolith is now a hornfels and exhibits a hornfelsic texture. The hornfels is composed dominantly of albite and quartz and subordinantly of calcite, titanite, apatite, ilmenite, and actinolite. The modal mineral percentages are albite 66, quartz 30, titanite 2.3, calcite 0.9, apatite 0.5, and actinolite 0.3. The bulk composition of the xenolith is variable, as shown in Table 1 of Benimoff and Sclar (1984) which is not unexpected for a meta-sedimentary rock. Normative albite ranges from 56.4 to 80.2 wt.%, whereas normative quartz ranges from 7.0 to 35.4 wt.%.

Stop 2. The Olivine Zone (Palisades Park, Fort Lee)

From River Road turn into Henry Hudson Drive; The olivine zone is on the left side of the road near the first left curve. (Latitude 40.846969°; Longitude -74.153697°

The olivine zone (hyalosiderite) occurs 7 to 20 meters above the base of the southern portion of the Palisades sill but is less well defined within the northern portion because it is replaced by a more hypersthenes-enriched facies with less olivine (altered to iddingsite) but with similar whole rock chemistry, particularly Cr/Mg ratios (Steiner et al, 1993). The northern end of the olivine layer is exposed at Nyack Beach State Park (Stop 6 of this field guide).

Stop 3. Trondhjemite and Syenite fusion (Ross Dock)

Continue on Henry Hudson drive and park at Ross Dock. Walk up to traffic circle and outcrop is just west of the circle at Latitude 40.857369° and longitude -73.959189°.

The chill zone at the base of the sill is clearly exposed along the park road above Ross Dock. The diabase here is aphyric except for glomeroporphyritic aggregates of pyroxene and plagioclase and contains very few xenoliths; characteristics shared by the Orange Mt basalt. The lower contact is largely parallel to the bedding planes of the underlying metasediments but is locally discordant. Flow of the first pulse of diabase through the Lockatong argillite has excavated a few channel-like cuts several meters across that truncate underlying Lockatong bedding plains at about 30°. In addition, anticlinal dome-like structures consisting of Lockatong Formation that rise a few meters into overlying diabase are exposed here (Fig. 5). It is at these dome-like structures where fusion has occurred.

Figure 5 (on page 94). Lower contact of the Palisades Sill along the road to Ross Dock. Migmatites are present to the right of the meter stick in an anticlinal dome structure.

Four meters of the section directly below the Palisades contact where considerable fusion has occurred includes 0.3 m of "buff arkose" in contact with the diabase, underlain by 0.2 m of "bedded siltstone", underlain by 2.9 m of "buff arkose" underlain by 1 m of "platy and bedded siltstone" and 0.5 m of "laminated siltstone". These lithologies have been described by Olsen (1980) as meta detrital cycles and he has correlated them continuously for 12 km along the base of the sill. The uplift of the black "laminated siltstone" has brought it to close proximity to the sill contact. The domed structures may have involved movement of volatiles derived from brackish groundwater. The salt content of these volatiles may have helped flux the melting process. Each of the lithologies exposed at the stop 3 dome exposure have been chemically analyzed by Benimoff and Puffer (2000). Both the syenite and trondhjemite fusion products are sodic typically containing 4 and 7.5 % Na_2O respectively, but K_2O and Rb is highly partitioned into the syenite, typically containing 5% and 125 ppm vs. only 0.5% and 25 ppm for the trondhjemite. Both the syenite and trondhjemite have similar REE contents comparable to the Lockatong Argillite (Benimoff and Puffer, 2000). The syenite, trondhjemite, and Lockatong laminated siltstone each display negative Nb anomalies in contrast to the Palisades diabase.

Stop 4. Alpine
(near Exit 2 Palisades Interstate Parkway; close to Palisades Interstate Park Headquaters)

The central portion of the Palisades sill is exposed along an extensive section extending from about 60 m above the base to approximately the 150 m level. Portions of this section are characterized by cm-scale rhythmic banding (Fig. 6). Thin sections (Block, et al, 2004) reveal associations of plagioclase and CPX tend to be separated by finer-scale plagioclase chains consistent with a convective flow model. Thin sections also reveal that the rhythmic banding is complicated by a cryptic mm-scale banding.

Figure 6. Centimeter scale rhythmic banding in Alpine Section of the Palisades Sill.

Stop 5A. Coarse Augite Dolerite (Valley Cottage)

As described by Steiner (1989) this is one of the few localities not in someone's yard which shows coarse augite dolerite (granophyre). However, the exposure here is weathered. The iron-enrichment of the Palisades at this stop is due to abundant titanomagnetite in the groundmass that will attract a magnet. The titanomagnetite is often dendritic and is considered to occur as both a primary and a quench feature. Chemically and to a certain extent petrographically this coarse dolerite appears to be equivalent to the late stage ferrodolerite, fayalite granophyres and other facies of Walker (1969), except that the iron enrichment is not accompanied by iron rich ferromagnesian minerals such as ferroaugite or fayalite. Ferroaugite rims may occur, but the bulk clinopyroxene is clearly augite.

REE patterns (Steiner, 1989) indicate extreme enrichment and are consistent with the derivation of the coarse augite dolerite from pigeonite facies through crystal liquid fractionation.

Stop 5B. The Upper Contact (Route 303 and Lake Road in Valley Cottage)

Parking can be accommodated at the local deli, across the street at the Bank, or in the large Valley Cottage Parking lot one block west on Lake Road; the latter will accommodate buses.

One of a few rare outcrops of the upper contact of the Palisades is exposed near the intersection of Route 303 and Lake Road in Valley Cottage and has been described by Steiner (1989). Another upper contact exposed about 3.2 km to the south is also described by Steiner (1989). The Valley Cottage location marks a sharp boundary between baked lake sediments and the upper quench zone contact with the Palisades. NYSGA members on the trip are challenged to find the contact. Note that the bench or small ridge that runs roughly east west toward Rockland Lake (from the intersection toward the Chase bank looking across the structure) represents the contact metamorphic zone. This location is on the partially exposed 'ring-dike' structure at the northernmost termination of the Palisades that extends northward along the Hudson toward Haverstraw than arcs westward, circling back to this location comprising in map view a large circular structure.

The texture of the contact diabase is microporphyritic with augite microphenocrysts. Within 2 meters of the contact the texture is ophitic, and beginning about 5 meters south patches of interstitial mesostasis appear. Farther south the diabase appears to plunge beneath the sedimentary cover. About 1 km north of Lake Road the contact reappears on the upper portions of the exposed slope which is inset 60 meters east of route 303. The dolerite dips roughly west at about 45 degrees under a veneer of shales. The intrusive contact is, therefore, complex, winding in a curvilinear fashion along its margin the general form of poorly exposed coalesced domes or localized arches.

Stop 6. Nyack Beach State Park

Enter Nyack Beach State Park, curve around the toll booth and follow paved incline up the hill to the left. Turn left at the top of the hill approximately 100 yards to the circular turn around and park. Latitude 40.95305°; Longitude -73.920490°)

Steiner (1989) described in detail the geology of Nyack Beach State Park. The escarpment at the eastern edge of the park represents the western wall of an infilled stone quarry. On this rock face, at approximately 13 meters above the presently obscured basal contact, a subhorizontal "rotten zone" can be observed. This zone presumably represents the last vestige of the olivine layer referred to by Walker (1969). It appears to fade northward of the present location.

A close up view of the olivine zone reveals that the basal part is undulose and uneven. It is marked by nodules of basalt enclosed in a weathered granular matrix. An analysis of one of the olivine bronzite nodules yields 10 percent MgO. This percentage is high for most of the Palisades, but less than the 19 percent reported by Walker (1969) for the hyalosiderite facies.

Here, as elsewhere, it appears that cooling cracks are reasonably continuous through the olivine zone indicating that it belongs to the same cooling unit as the overlying dolerite.

REFERENCES CITED

Benimoff, A. and Puffer, J. H., 2000, Syenitic and trondhjemitic fusion at Mesozoic diabase intrusion contacts with sedimentary rocks of northern New Jersey and Staten Island: Field-trip Guidebook, 35th Annual Meeting of the Northeastern Section, Geological Society of America, New Brunswick, New Jersey, 46 pp.

Benimoff, A. I., Puffer, J. H., and Sclar, C. B., (2000) Fusion of Lockatong Argillite at the base of the Palisades Sill *Geological Society of America Abstracts with Programs* V32, *No. 1* A-5

Benimoff, A., and Puffer, J. H., 2005, Tri-modal fusion of meta-argillites along margins of early Jurassic diabase intrusions, New Jersey and New York. Northeastern Geology & Environmental Sciences, v. 27, p. 265-275.

Benimoff, A. I., and Puffer, J. H. 2007. Effects of metasomatism and host rock fusion on the chemistry of early Jurassic Palisades diabase in the Newark Basin. Geol. Soc. of America, Northeastern Section, Abstracts with Program. 39, No 1, p. 94.

Benimoff, A. I. and Sclar, C. B. (1984) "Coexisting Silicic and Mafic Melts Resulting From Marginal Fusion of a Xenolith of Lockatong Argillite the Palisades Sill, Graniteville, Staten Island, New York" *American Mineralogist*, 69, 1005-1014.

Block, K.A., 2006. A refined multiple injections model for the Palisades Sill, NY and NJ. Ph.D. Thesis, City University of New York.

Block, K.A., Steiner, N.C., Steiner, J.C., and Puffer, J.H., 2007. Iron enrichment in basaltic rocks of the Newark Basin: Crystallization in an open system. Geol. Soc. Am. Abstr. Program 39, No 1, p. 93.

Block, K.A., Steiner, J.C., and Rice, A., 2004. Toward a comprehensive model for the Palisades: tracing new internal contacts. *In* Puffer J.H., and Volkert, R.A., eds. Neoproterozoic, Paleozoic, and Mesozoic Intrusive Rocks of Northern New Jersey and Southeastern New York, Field Guide and Proceedings, Geol. Ass. New Jersey. 21: 93-99.

Cartwright, J., and Hansen D. M., 2006, Magma transport through the crust via interconnected sill complexes. Geology 11: 929-932.

Dorais M.J. and Tubrett M., 2008, Identification of a subduction zone component in the Higganum dike, Central Atlantic Magmatic Province: A LA-ICPMS study of clinopyroxene with implications for flood basalt petrogenesis: Geochem. Geophys. Geosyst (G^3) v. 9, Q10005,doi:10.1029/2008GC002079.

Drake, A.A., Jr., Volkert, R. A., Monteverde, D.H., Herman, G. C., Houghton, H. F., Parker, R. A., and Dalton R. F., 1996. Bedrock Geologic Map of Northern New Jersey: U.S. Geological Survey Miscellaneous Series Map 1-2540-A, scale 1:100,000.

Durand, S.R., and Sen, G., 2004. Preeruption history of the Grande Ronde Formation lavas, Columbia River Basalt Group, American Northwest: evidence from phenocrysts. Geology. 32:293-296.

Forsha, C. J., and Zieg, M. J., 2006, Interpretation of system-wide injection dynamics in mafic sills; the Nipigon diabase, Ontario, Canada. Geol. Soc. Am. Abstr. Program 38, No. 4, p. 65.

Froelich, A. J., and Gottfried, D., 1988, An overview of early Mesozoic intrusive rocks in the Culpeper basin, Virginia and Maryland. *In* Froelich, A. J., and Robinson, G. R., Jr., eds. Studies of the Early Mesozoic Basins of the Eastern United States, U.S. Geol. Survey Bull. 1776:151-165.

Gibb F.G.F., and Henderson, C.M.B., 1992. Convection and crystal settling in sills. Cont. Mineral. Pet. 109:538-545.

Gibb F.G.F., and Henderson, C.M.B., 2006. Chemistry of the Shiant Isles Main Sill, NW Scotland, and wider implications for the petrogenesis of mafic sills. J. Pet. 47:191-230.

Gorring, M.L., and Naslund, H. R., 1995. Geochemical reversals with the lower 100 m of the Palisades sill, New Jersey. Cont. Mineral. Pet. 119:263-276.

Gottfried, D. Froelich, A. J., and Grossman, J. N., 1991a. Geochemical data for Jurassic diabase associated with early Mesozoic basins in the Eastern United States: Western Newark Basin, Pennsylvania and New Jersey. U. S. Geol. Surv. Open-File Report 91-322-D, 26 pp.

Gottfried, D., Froelich, A. J., and Grossman, J. N., 1991b. Geochemical data for Jurassic diabase associated with early Mesozoic basins in the Eastern United States: Eastern Newark Basin, New York and New Jersey. U. S. Geol. Surv. Open-File Report 91-322-C, 6 pp.

Husch, J. M., 1992a. Geochemistry and petrogenesis of the Early Jurassic diabase from the central Newark basin of New Jersey and Pennsylvania. *In* Puffer, J. H., and Ragland, P. C., eds. Eastern North American Mesozoic Magmatism. Geol. Soc. Am., Special Paper. 268:169-192.

Husch, J. M., 1990. Palisades Sill: Origin of the olivine zone by separate magmatic injection rather than gravity settling. Geology 18:699-702.

Johnson, D. M, Hooper, P. R., and Conrey, R. M., 1999. XRF Analysis of rocks and minerals for major and trace elements on a single low dilution Li-tetraboarate fused bead, JCPDS- International Centre for Diffraction Data, 843-867

Kent, D.V., and Olsen, P.E., 2007. Early Jurassic magnetostratigraphy and paleolatitudes from the Hartford continental rift basin (eastern North America): Testing for polarity bias and abrupt polar wander in association with the Central Atlantic Magmatic Province. J. Geop. Research, in review.

Kodama, K. P. 1983. Magnetic and gravity evidence for a subsurface connection between the Palisades sill and the Ladentown basalts. Geol. Soc. Am. Bull. 94:151-158.

Lehnert, K., Su, Y., Langmuir, C., Sarbas, B., and Nohl. U., 2000. A global geochemical database structure for rocks. Geochem. Geophy. Geosystems. 1, doi:10.1029/1999GC00026

Maes, S. M., Ferre, E. C., Tikoff, B., Brown, P.E., and Marsh, J. S., 2007, Rock magnetic stratigraphy of a mafic layered sill: A key to the Karoo volcanic plumbing system. J. Volcanology and Geothermal Research.

Marsh, B.D., 1996. Solidification fronts and magmatic evolution. Min. Mag. 60:5-40.

Marsh, B.D., 2004. A magmatic mush column Rosetta Stone: The McMurdo Dry Valleys of Antarctica. EOS, Trans. Amer. Geophysical Union 85:497-502

Marzoli, A., Renne, P., Piccirillo, E., Ernesto, M., Bellieni, G., and De Min, A., 1999. Extensive 200-million-year-old continental flood basalts of the Central Atlantic Magmatic Province. Science 284:616-618.

Merguerian, C., and Sanders, J. E., 1995. NE not SE directed paleoflow of the Palisades magma north of Staten Island: new evidence from xenolith and contact relationships. *In* Hanson, G. N., ed. Geology of Long Island and Metropolitan New York, State Univ. of New York at Stony Brook, NJ, Long Island Geologists Program with Abstracts, p. 64-77.

McHone, J. G., 2003. Volatile emissions from Central Atlantic Magmatic Province basalts: Mass assumptions and environmental consequences. *In* Hames, W. E., McHone, J. G.,Renne, P. R., and Ruppel, C., eds. The Central Atlantic Magmatic Province: Insights from Fragments of Pangea: American Geophysical Union Geophysical Monograph 136:241-254.

Olsen, P.E., 1980. The latest Triassic and early Jurassic formations of the Newark basin (eastern North America, Newark supergroup) : Stratigraphy, structure, and correlation. New Jersey Academy of Science Bull. 10 :40-77.

Olsen, P.E., Kent, D. V., Et-Touhami M., and Puffer, J. H., 2003. Cyclo, magneto-, and bio-stratigraphic constraints on the duration of the CAMP event and its relationship to the Triassic-Jurassic boundary. *In* Hames, W.E., McHone, J.G., Renne, P.R., and Ruppel, C., eds. The Central Atlantic Magmatic Province Insights From Fragments of Pangea, Geophysical Monograph 136, American Geophysical Union, Washington DC, p. 7-32.

Pearce, T. H., 1970. Chemical variations in the Palisades Sill. J. Pet. 11 :15-32.

Pegram, W. J., 1983, Isotopic characteristics of the Mesozoic Appalachian tholeiites: Geological Society of America, Abstracts with program, v. 15, p. 660.

Puffer, J. H., 2003, A reactivated back-arc source for CAMP Magma: *in* Hames, W. E., McHone, J. G., Renne, P. R., and Ruppel, C. (editors) The Central Atlantic Magmatic Province: Insights From Fragments of Pangea: American Geophysical Union Geophysical Monograph 136, p. 151-162.

Puffer, J.H., 2001, Contrasting HFSE contents of plume sourced and reactivated arc-sourced continental flood basalts: Geology, v. 29, p. 675-678.

Puffer, J. H., l992. Eastern North American flood basalts in the context of the incipient breakup of Pangea. *In* Puffer, J. H., and Ragland, P. C., eds. Eastern North American Mesozoic Magmatism: Geol. Soc. Am., Special Paper. 268:95-119.

Puffer, J. H., and Lechler, P. l980, Geochemical cross-sections through the Watchung Basalts of New Jersey: Geological Society of America Bulletin, v. 91, part 1, p. 7-10, part 11, p. 156-191.

Puffer, J. H., Hurtubise, D. O., Geiger, F. J., and Lechler, P. l981, Chemical composition and stratigraphic correlation of Mesozoic basalt units of the Newark Basin, New Jersey and the Hartford Basin, Connecticut: Geological Society of America Bulletin, v. 92, part 1, p. 155-159, part 11, p. 515-553.

Puffer, J. H., and Philpotts, A. R., l988. Eastern North American quartz tholeiites: geochemistry and petrology. *In* Manspeizer, W., ed. Triassic-Jurassic Rifting, Continental Breakup and the Origin of the Atlantic Ocean, Developments in Geotectonics. 22, Elsevier Scientific Publishers, Amsterdam. p. 579-606.

Puffer, J. H., and Student, J. J., l992. The volcanic structure and eruptive style of the Watchung Basalts, New Jersey. *In* Puffer, J. H. and Ragland, P. C., eds. Eastern North American Mesozoic Magmatism. Geol. Soc. Am. Spec. Paper 268, Boulder CO., p. 261-279.

Puffer, J. H., Geiger, F. J., and Camano, E. J., l982. The geochemistry of the Mesozoic igneous rocks of Rockland County, New York. Northeastern Geol. 4, 121-130.

Puffer, J. H., and Husch J., l996. Early Jurassic diabase sheets and basalt flows, Newark Basin, New Jersey: An updated geological summary and field guide. *In* Benimoff, A.I., and Ohan, A.A., eds. Field Trip Guide for the 68th Annual Meeting, New York State Geological Association, The College of Staten Island/CUNY, Staten Island, NY, p. 157-176.

Puffer, J. H., and Volkert, R. A., 2001. Pegmatoid and gabbroid layers in Jurassic Preakness and Hook Mountain Basalts, Newark Basin, New Jersey. J. Geol. 109:585-601.

Puffer, J. H., Block, K.A., and Steiner, J.C., 2009, Transmission of flood basalts through a shallow crustal sill and the correlation of sill layers with extrusive flows: The Palisades Intrusive System and the basalts of the Newark Basin, New Jersey, USA, The Journal of Geology, v. 17, p. 139-155.

Ragland, P. C., Cummins, L. E., and Arthur, J. D., 1992, Compositional patterns for early Mesozoic diabases from South Carolina to central Virginia. *In* Puffer, J. H., and Ragland, R. C., eds. Eastern North American Mesozoic Magmatism: Geol. Soc. Am. Special Paper. 268:309-332.

Ratcliffe, N.M., 1988. Reinterpretation of the relationship of the western extension of the Palisades sill to the lava flows at Ladentown, New York, based on new core data. *In* Froelich, A. J., and Robinson, G.R., Jr., eds. Studies of the Early Mesozoic Basins of the Eastern United States, U.S. Geol. Survey Bull. 1776:113-140.

Salters, V. J. M., Ragland, P.C., Hames, W.E., Milla, K., and Ruppel, C., 2003. Temporal chemical variations within lowermost Jurassic tholeiitic magmas of the Central Atlantic Magmatic Province. *In* Hames, W.E., McHone, J.G., Renne, P.R., and Ruppel, C., eds. The Central Atlantic Magmatic Province Insights From Fragments of Pangea, Geophysical Monograph 136, American Geophysical Union, Washington DC. p. 163-178.

Seifert, K. E., and Olmsted, J. F., 2004. Geochemistry of North Shore hypabyssal dikes and sills in the Midcontinent Rift of Minnesota: an example – the 47th Avenue sill. Can. J. Earth Sci. 41:829-842.

Shervais, J. W., Vetter, S. K., and Hanan, B. B., 2006. Layered mafic sill complex beneath Eastern Snake River Plain: Evidence from cyclic geochemical variations in basalt. Geology 34:365-368.

Shirley, D. N., 1987. Differentiation and compaction in the Palisades Sill, New Jersey. J. Pet. 28:835-865.

Simkin, T., 1967. Flow differentiation in the picritic sills of northern Skye. *In* Wyllie, P.J. ed. Ultramafic and Related Rocks. Wiley, New York, NY p. 64-69.

Smith, R.C., Rose, A.W., Lanning, R.M., 1975. Geology and geochemistry of Triassic diabase in Pennsylvania. Geol. Soc. Am. Bull. 86:943-955.

Steiner, J. C., 1989, The Nyack Section of the Palisades: Facies, Contacts and Lava Domes, *In* Weiss D., and O/Brien, L. eds. NYSGA Field-trip Guidebook 61st Annual Meeting, Middletown NY. p. 269-282.

Steiner, J. C., Walker, R.J., and Warner, R.D., 1989, Mineralogy and REE systematic in augite-dolerites of the Palisades Sill, Nyack New York: magma types and petrogenesis, Geol. Soc. America NE Section, Annual Meeting v. 21, p. 68.

Steiner, J. C., Walker, R. J., Warner, R. D., Olson, T. R., 1992. A cumulus-transport-deposition model for the differentiation of the Palisades Sill. *In* Puffer, J. H., Ragland, R. C., eds. Eastern North American Mesozoic Magmatism: Geol. Soc. Am. Special Paper. 268:193-218.

Sun, S.S., McDonough, W.F., l989. Chemical and isotopic systematics of oceanic basalts: implications for composition and processes. *In* Saunders A.D., Norry, M. J., eds. Magmatism in the Ocean Basins, Geol. Soc. Am. Spec. Pub. 42, Boulder, CO, p. 313-345.

Thomson, K. 2007. Determining magma flow in sills, dikes and laccoliths and their implications for sill emplacement mechanisms. Bull. Volcanology. 70:183-201.

Tollo, R. P., Gottfried, D., 1992. Petrochemistry of Jurassic basalt from eight cores, Newark Basin, New Jersey: Implications for the volcanic petrogenesis of the Newark Supergroup. *In* Puffer, J. H., Ragland, P. C., eds. Eastern North American Mesozoic Magmatism. Geol. Soc. Am. Spec. Pap. 268, Boulder CO, p. 233-259.

Vuori, S.K. and Luttinen, A. V., 2003. The Jurassic gabbroic intrusions of Utpostane and Muren: insights into Karoo-related plutonism in Dronning Maud Land, Antarctica. Antarctic Science. 15:283-301.

Walker, F., 1940. The differentiation of the Palisades diabase, New Jersey. Geol. Soc. Am. Bull. 51:1059-1106.

Walker, K. R., 1969. The Palisades sill, New Jersey: A reinvestigation. Geol. Soc. Am. Special Paper 111:178.

Weigand, P. W., Ragland, P. C., l970. Geochemistry of Mesozoic dolerite dikes from eastern North America. Contrib. Mineral. Pet. 29:195-214.

White, C. M., 2007. The Graveyard Point Intrusion: an example of extreme differentiation of Snake River Plain Basalt in a shallow crustal pluton. J. Pet. 48:303-325.

Fossil Great Lakes of the Newark Supergroup – 30 Years Later

Paul E. Olsen
Lamont-Doherty Earth Observatory
of Columbia University, Palisades, NY 10968

ABSTRACT

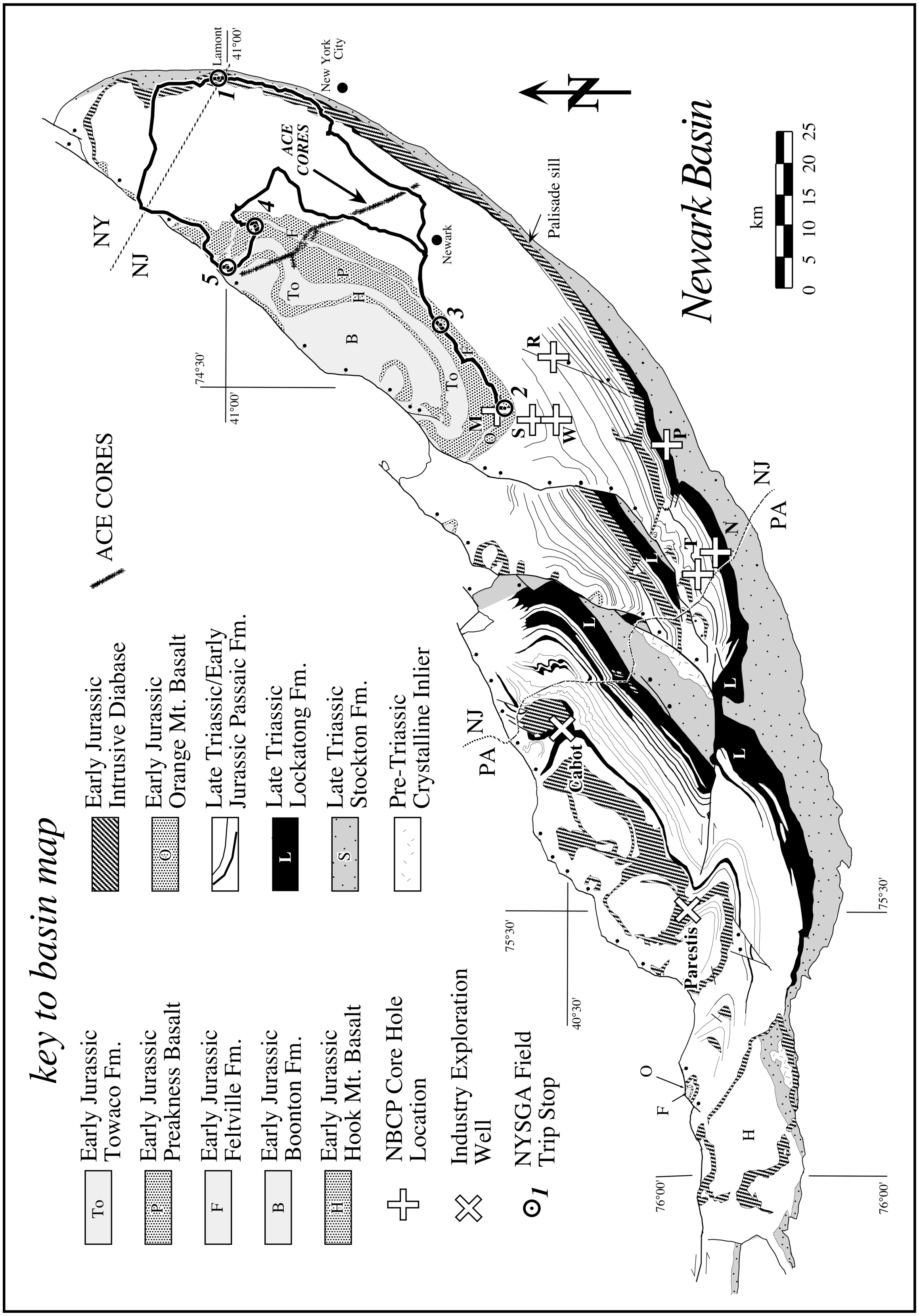
In the 1980 NYSGA meeting (Olsen, 1980b) I led a fieldtrip to look at strata of the Triassic and Jurassic great lakes of the Newark basin which are among the largest lakes known of any time. Our view of these giant lake systems has been much amplified since that time because of the collection of industry seismic lines, the drilling of two deep exploration holes, about 40,000 ft of core by the Army Corps of Engineers, 26,700 ft of core by the Newark Basin Coring Project (NBCP), and quite a few Ph.D. theses (including my own). The purpose of this fieldtrip is to examine outcrops and related cores in the Newark basin, that illustrate some of the key features of these giant lake strata and interbedded flood basalts and how they play key roles in current debates on climate change, mass extinctions, evolution, ascent of the dinosaurs, the rifting process, carbon sequestration, and chaos in the Solar System.

INTRODUCTION:
THE TRIASSIC-JURASSIC GREAT LAKES OF CENTRAL PANGEA

This guidebook focuses of the deposits, fossils, and context of huge lakes, related sedimentary deposits, and giant lava flows formed during latest Triassic and Early Jurassic in the Newark basin of New York, New Jersey, and Pennsylvania, one of a series of remnants of a rift valleys system preserved in Eastern North America (Figure 1). I define the term Great Lakes as meaning lakes whose dimensions during high stands were on the scale of over 100 km in longest dimension comparable to the scale of the American Great Lakes or the East African Great Lakes and perhaps approaching the scale of land-licked seas such as the Black or Caspian seas. This introduction discusses in general terms the major themes that guided the selection of the stops and which inform the impotance of observations we can make on the outcrop.

As should be expected ideas have changed considerably in 30 years and new observations, especially from cores, have molded new ones. As much as possible I will try to point out how ideas have changed and also how interpretations of outcrops have changed since 1980 (Olsen, 1980b).

This contribution is dedicated to Warren Manspeizer, editor of the 1980 NYGSA guidebook who encouraged a generation of scientists, including me, intellectually in general and urged me specifically to write two papers for that guidebook (Olsen, 1980a, 1980b).


key to basin map
To Early Jurassic Towaco Fm.
P Early Jurassic Preakness Basalt
F Early Jurassic Feltville Fm.
B Early Jurassic Boonton Fm.
H Early Jurassic Hook Mt. Basalt
NBCP Core Hole Location
Industry Exploration Well
1 NYSGA Field Trip Stop
Early Jurassic Intrusive Diabase
O Early Jurassic Orange Mt. Basalt
Late Triassic/Early Jurassic Passaic Fm.
L Late Triassic Lockatong Fm.
S Late Triassic Stockton Fm.
Pre-Triassic Crystalline Inlier
ACE CORES
Newark Basin
km
0 5 10 15 20 25
N
Palisade sill
Newark
New York City
Lamont
41°00'
74°30'
75°30'
40°30'
76°00'
NY
NJ
PA
Cabot
Parestis

The Rifting Process and Origin of the Atlantic Ocean

The rifting process that eventually resulted in the formation of the central part of the Atlantic Ocean was protracted over a period of at least 40 million years from the Late Permian to the Early Jurassic, with at least one spectacular and probably catastrophic punctuation (the CAMP eruptions). The rifting zone itself[1] is the largest we know of on our planet, arguably stretching roughly 8000 km northeast-southwest from Greenland and Europe to the Gulf of Mexico and the Pacific and north west-southeast about 1000 km (Figure 2). In this rifting zone there were many, perhaps hundreds, large, mostly normal, fault systems along which there was geologically significant hanging wall subsidence leading to the development of large rift valley systems, many of which were individually hundreds of kilometers long, and that filled with thousands of meters of sediments. Over much of this rifting region bisecting central Pangea, the rifts formed far from access to the sea and hence when subsidence exceeded the rate at which sediments could fill them, which apparently was most of the time, lakes, great lakes filled the deficit. Thus, in most cases the sediments filling the individual rift basins were completely of continental origin, and largely lacustrine. As far as we know, only adjacent to the Tethys and along the north-central axis of the rifting zone did marine brines or open marine waters make it into the rift basins, and most of that occurred late in the Triassic-Jurassic part of the rifting process.

As is the case with rift systems today, many of the individual rift basin systems within this huge rifting zone were almost certainly connected with one another not only by river systems, but also by veneers of alluvial strata shed from protruding footwall basement or inselbergs remaining from pre-rift topography. At times of extreme lake level highs many of the otherwise isolated great lakes may have become united into lakes much larger than the depocenters marking the main individual rift basins systems forming lakes the scale of the Caspian Sea as illustrated by the stops on this field trip.

The CAMP LIP

While the rifting process itself was protracted, it was marked by one brief but immense basaltic igneous event involving the emplacement of the Central Atlantic Magmatic Province (CAMP) (Marzoli et al., 1999) that may have lasted only about 600 ky (Olsen et al., 1993, 2003). The intrusive system of the CAMP is spread from France through Iberia, Eastern North America, West Africa, to western Brazil, covering and area

Figure 1: (previous page). Map of the Newark basin showing location of field stops, trip route, and wells and cores.

---

[1] It is useful to make the distinction between the broad rifting zone, which is the entire region of thinning crust with dimensions of thousands of kilometers cut by swarms of normal faults, and the many individual rift basins and rift basin systems with dimensions of hundreds of kilometers or smaller that formed along the largest of the normal fault systems. In context of our current understanding of these rifting zones, the concept of “failed rifts” or “aulacogen” as a “failed arm” of a triple junction makes no sense, as is also true of the concept of an “axial dike” in a rift basin.

of about 11 million square kilometers. The dikes form an apparent radial pattern (May, 1971) the locus of which is in southern Florida or the Bahamas (Figure 3). Plausibly this radial swarm reflects the stress pattern along a thermal welt caused by the plume head. However, the existence of the radial pattern itself has been questioned by a number of authors (McHone, 2000; Coltice et al., 2009), but it as close to a radial pattern as seen anywhere in our Solar System, notably Venus (Figure 4) (e.g., Ernst et al., 2001; Hansen & Olive, 2010). This radial pattern may be the consequence of the arrival of a plume head (White & McKenzie, 1989) as is supposed to be the case for most other large igneous provinces (LIPs), although this concept is also strongly questioned (Anderson, 2000)[2].

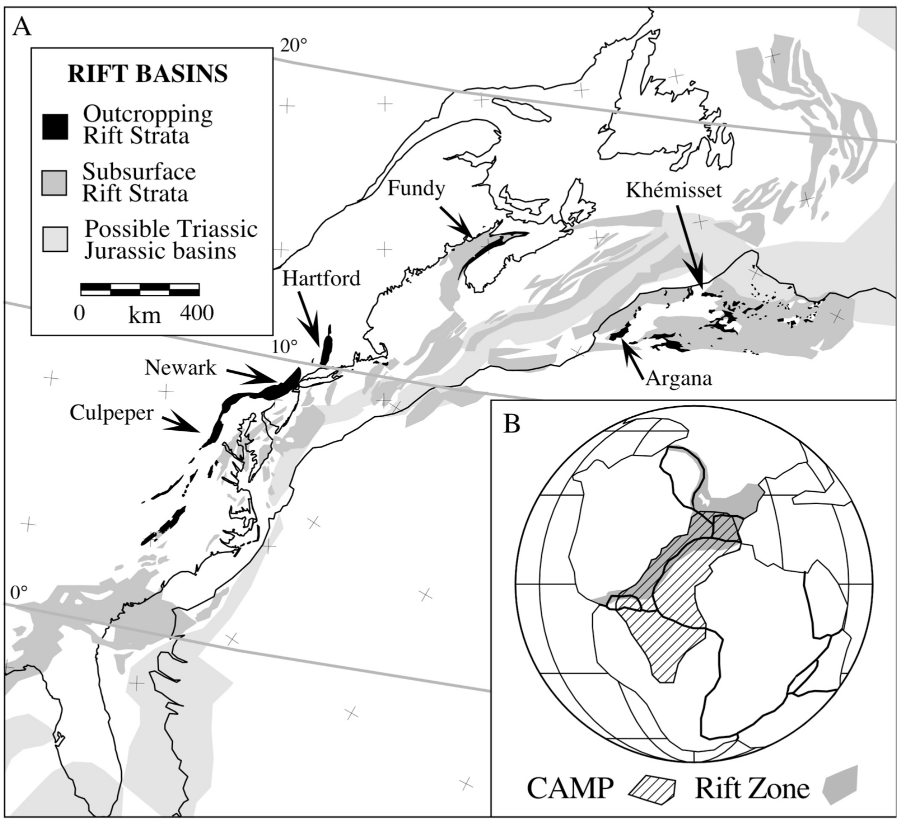


Figure 2: Distribution of rift basins of central Pangea and the distribution of the CAMP. A, Rift basins of the central Atlantic margins (for the Late Triassic); B, Pangea during the earliest Jurassic showing the distribution of the CAMP, overlapping much of the Triassic-Jurassic rift zone.

[2] One of the strongest arguments against the plume model for the CAMP has been the apparent absence of a "hot spot" track. It is clear that there is no such track on the North American plate. However, this is problem is illusionary because the North American Plate drifted almost straight northward after the CAMP and most of the hot spot track would be on the continental part of South American Plate, in northern South America, which did not move much during the Mesozoic, and where there are basaltic eruptives of appropriate age. Indeed modern plate reconstructions (e.g., Schettio et al., 2010) predict the current position of the CAMP hot spot to be at the Cape Verde Islands as suggested by (White and McKenzie, 1989; Oliveira et al., 1990; Ernst et al., 1995), which is of the right age, or to me less plausibly the Fernando de Noronha hot spot (Hill, 1991).

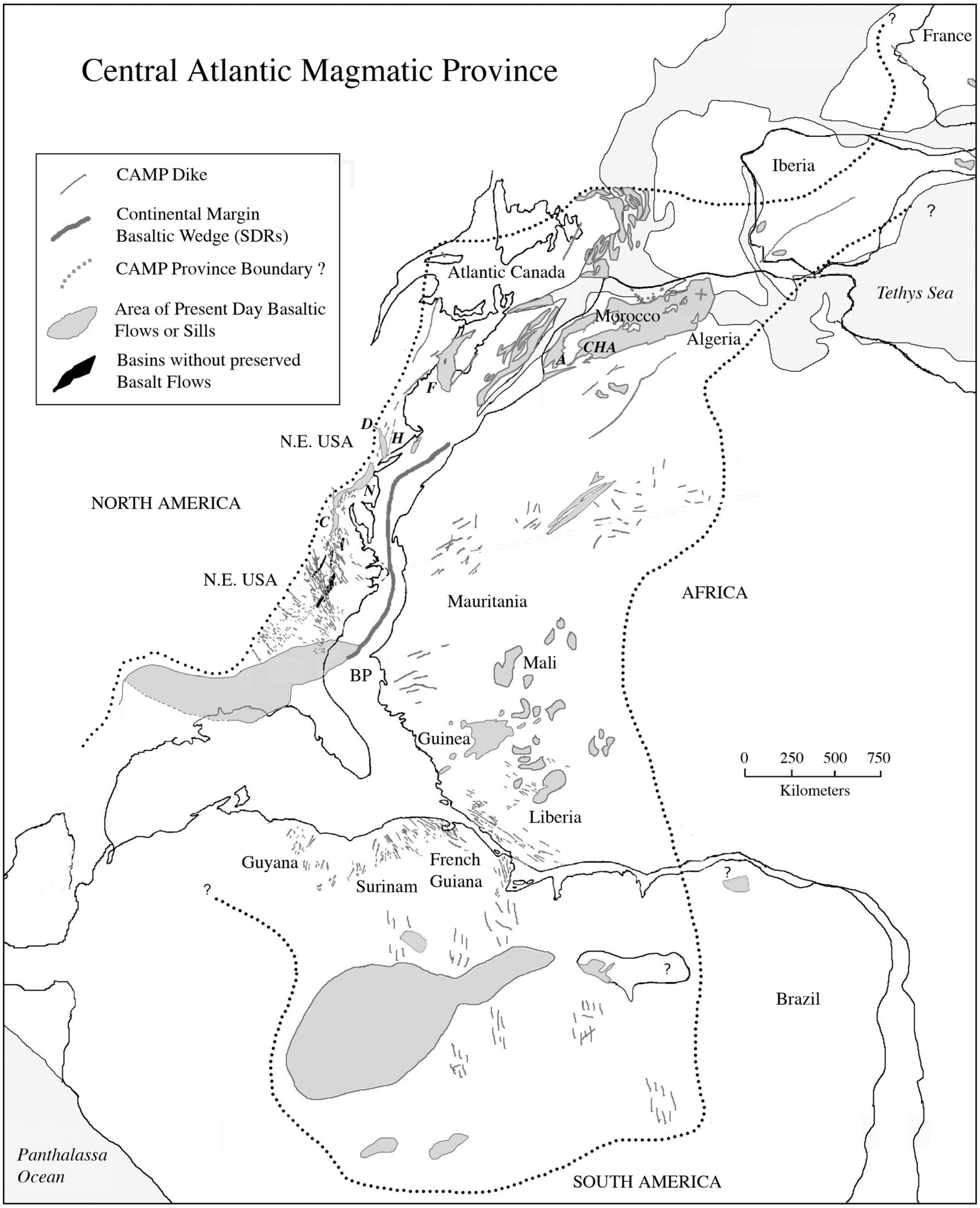


Figure 3: Distribution of the CAMP in central Pangea showing basins described in text. Modified from McHone (2000) and Whiteside et al. (2007).

The CAMP lava flows and other extrusives emanating from the intrusive system are known over a much smaller area than the intrusions, but this is certainly a result of extensive post-rift erosion. If the flows covered most of the area marked by the dike system the CAMP lavas would comprise the largest known continental flood basalt province on Earth, although whether they really covered such a large area remains to be tested (e.g., Olsen et al., 1999).

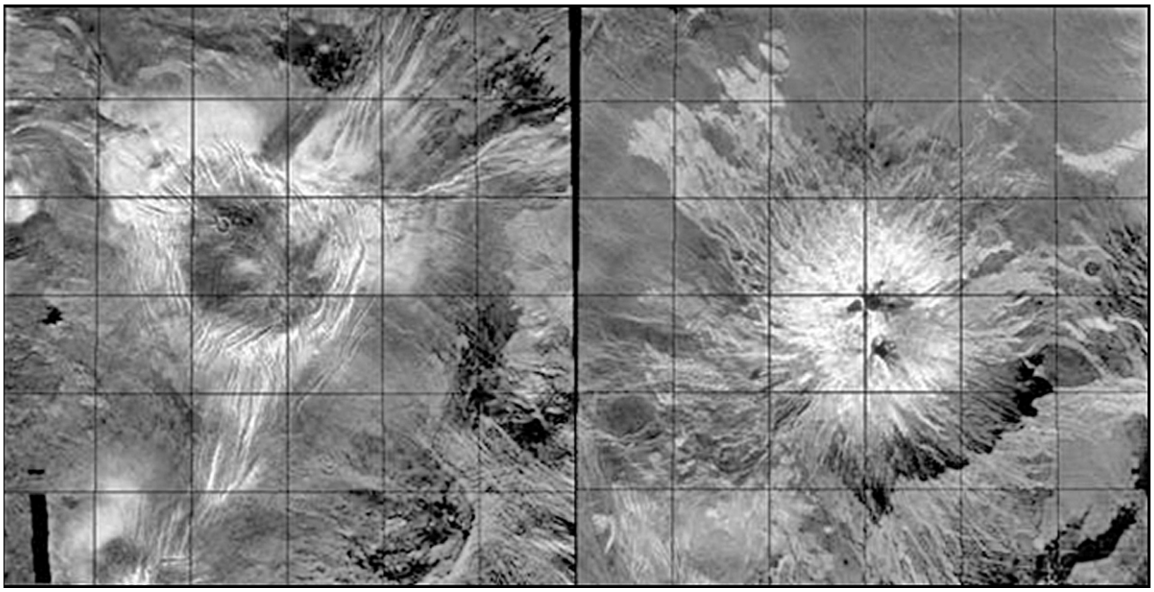


Figure 4: Radial dike swarms on Venus at Irnini Mons (left) and Sapas Mons (right). Note that these swarms are not purely radial, but rather have sets of parallel dikes and intersecting systems with different orientations. NASA Magellan Synthetic Aperture Radar (SAR) mosaic images from http://www.geology.pomona.edu/research/Faculty/ Grosfils/Venus/Volcano/large_ volcanoes.htm thus undermining the argument that such features in the CAMP dikes indicate the *lack* of a radial dike swarm and hence the lack of a plume.

CAMP igneous activity may also mark the initiation of Atlantic sea floor spreading (Schlische et al., 2003; Schlische, 2003; Olsen et al., 2003). What appear to be CAMP lavas (the Clubhouse Crossroads Basalt) are present in the subsurface in South Carolina and Georgia, and based on seismic reflection profiles, these seem to connect to the giant wedge shaped edifices of seaward dipping reflectors below the southeastern continental shelf (Austin et al., 1990; Holbrook et al., 1994; Kelemen & Holbrook, 1995; Oh et al., 1995; Talwani et al., 1995). The latter are taken to be subareally emplaced lavas formed at the onset of seafloor spreading as first identified in offshore Norway (Mutter et al., 1982). An accurate direct date of the Clubhouse Crossroads Basalt is still elusive, however, and the attractive hypothesis that the CAMP lavas mark the onset of seafloor spreading has yet to be critically and directly tested.

However, following the assumption that the seaward dipping reflectors are part of the CAMP to its logical conclusion, it seems plausible that the arrival of the CAMP plume head initiated the spread of magma, not only along the radial dike swarm, but perhaps also along crest of asthenosphere along the thinnest zone of crust approximating the axis of the rifting zone, setting off the extremely concentrated eruption of the seaward dipping reflectors, and initiating seafloor spreading. This axial asthenospheric crest would have been perpendicular to the minimum compressive stress direction on the northeastern side of the thermal welt. The thermal welt should also have produced maximum compressive stresses perpendicular to the normal fault system in the southeastern US, perhaps both terminating extension and causing small-scale tectonic inversion features consistent with the observed patterns seen in that region (e.g., Schlische et al., 2003; Maliconico, 2003), while at the same time accelerating continental extension further to the north along the rifting axis.

Critical to all arguments about the effects of the CAMP eruption are not just the huge volume of igneous rocks involved emplaced in a geologically relatively short period, but specifically the rate and magnitude of individual eruptive events. We will see some evidence at Stop 4 that individual lava flows spanned distances of minimally 800 km along the trend of the rifts and were in places over 200 m thick, making them among the largest if not the largest single lava flows known on Earth.

Cyclical Climate Change Reflected in the Deposits of Great Lakes.

Perhaps the single best-known feature of the Newark basin (and other Newark Supergroup basins) is the nearly pervasive sedimentary cyclicity in the abundant lacustrine deposits. This cyclicity was first identified implicitly by the mapping work of Dean B. McLaughlin of the University of Michigan in the 1930s through 1950s (e.g., McLaughlin, 1933, 1944, 1946a, 1946b, 1959) and explicitly by Franklyn Van Houten of Princeton University in the 1960s (Van Houten, 1962, 1964, 1969) and who's later works included a NYGSA guidebook paper (Van Houten, 1980) in the Manspeizer (1980) volume. Both workers, especially Van Houten, had a profound effect on my thinking and practice of stratigraphy (e.g., Olsen, 1980b).

Subsequent studies have confirmed and elaborated on the seminal works of McLaughlin and Van Houten, using quantitative methods of times series analysis that avoid the need to explicitly identify the cycles in advance (e.g., Olsen, 1996a, 1999). After application of an age model to the thickness periodicities, not only does do these methods recover the ~20 ky cycle, but also the ~100 and 405 ky cycles identified by Van Houten by counting cycles, as well as cycles with period of 1.8 and 3.5 m.y. In 1986 and 1996a, I named two the sedimentary cycles present in the Newark lacustrine strata after Van Houten and McLaughlin (Figure 5). Van Houten cycles are the meter-scale lake level cycle obvious at outcrop-scale paced by the ~20 ky climatic precession cycle and McLaughlin cycles are the map-scale cycles traced out by McLaughlin himself (Olsen, 1986; Olsen et al., 1996a) that were paced by the 405 ky eccentricity cycle. Van Houten cycles are lake level cycles controlled by precipitation and evaporation changes tracking the intensity of tropical insolation[3] (Olsen & Kent, 1996). The overall degree of

[3] The mechanism is thought to be that changes in tropical insolation (solar radiation energy received on a given surface area in a given time – watts/meter/unit time^2), controlled by celestial mechanical control of the Earth's orbital geometry, changes the intensity of upwelling warm air (tropical convergence) and hence precipitation. It rains, more or less, during the time and place of maximum insolation with the magnitude of precipitation being positively related to the magnitude of insolation.

development of wet and dry sedimentary facies (Figure 6) and facies-sensitive fossils in Van Houten cycles vary in their expression and in doing so comprise the four larger cycles. These are: the short modulating cycle, corresponding to ~100 ky eccentricity cycles; the McLaughlin cycle, corresponding to the 405 ky cycle, and the two additional eccentricity cycles of roughly 1.8 and 3.5 m.y. (Olsen & Kent, 1996, 1999).

.

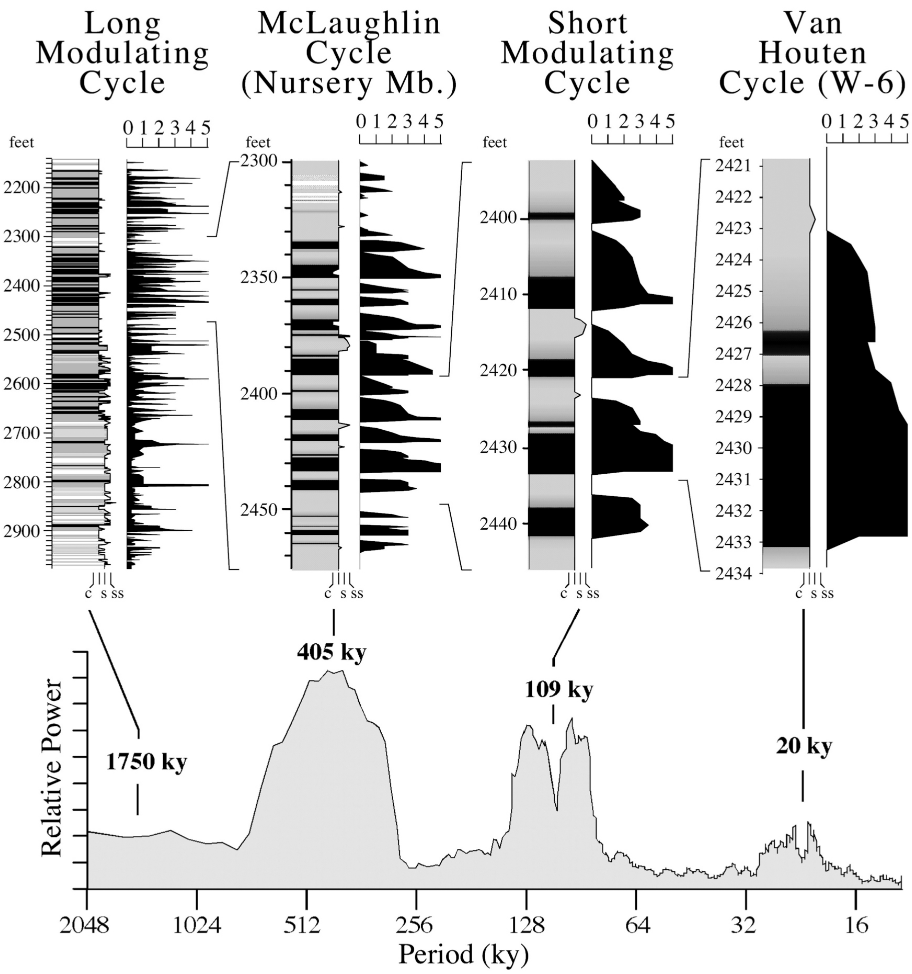


Figure 5: Lacustrine cycles of the Newark basin (above) from the Nursery no. 1 core and power spectrum tuned to the 405 ky cycle. White represents red , very light gray is purple, gray is gray, and black is black.

McLaughin never seemed to state an opinion in print of the origin of the cycles he mapped, although he and Van Houten were contemporaries and Van Houten correctly ascribed them to what we more precisely identify today as a 405 ky cycle caused by the gravitational interaction of Venus and Jupiter. This is somewhat ironic because McLaughlin was a professor of astronomy at the University of Michigan and in those circles is known for his work on volcanism on Mars, novae, and Be class stars.

Fossils found in Newark lacustrine strata occur in predicable facies within the context of the Van Houten cycles tracking the climatic precession cycle and the modulating longer term cycles paced by eccentricity (Figure 6). In particular, articulated fish tend to occur in the microlaminated facies in the deepest-water phases of the Van

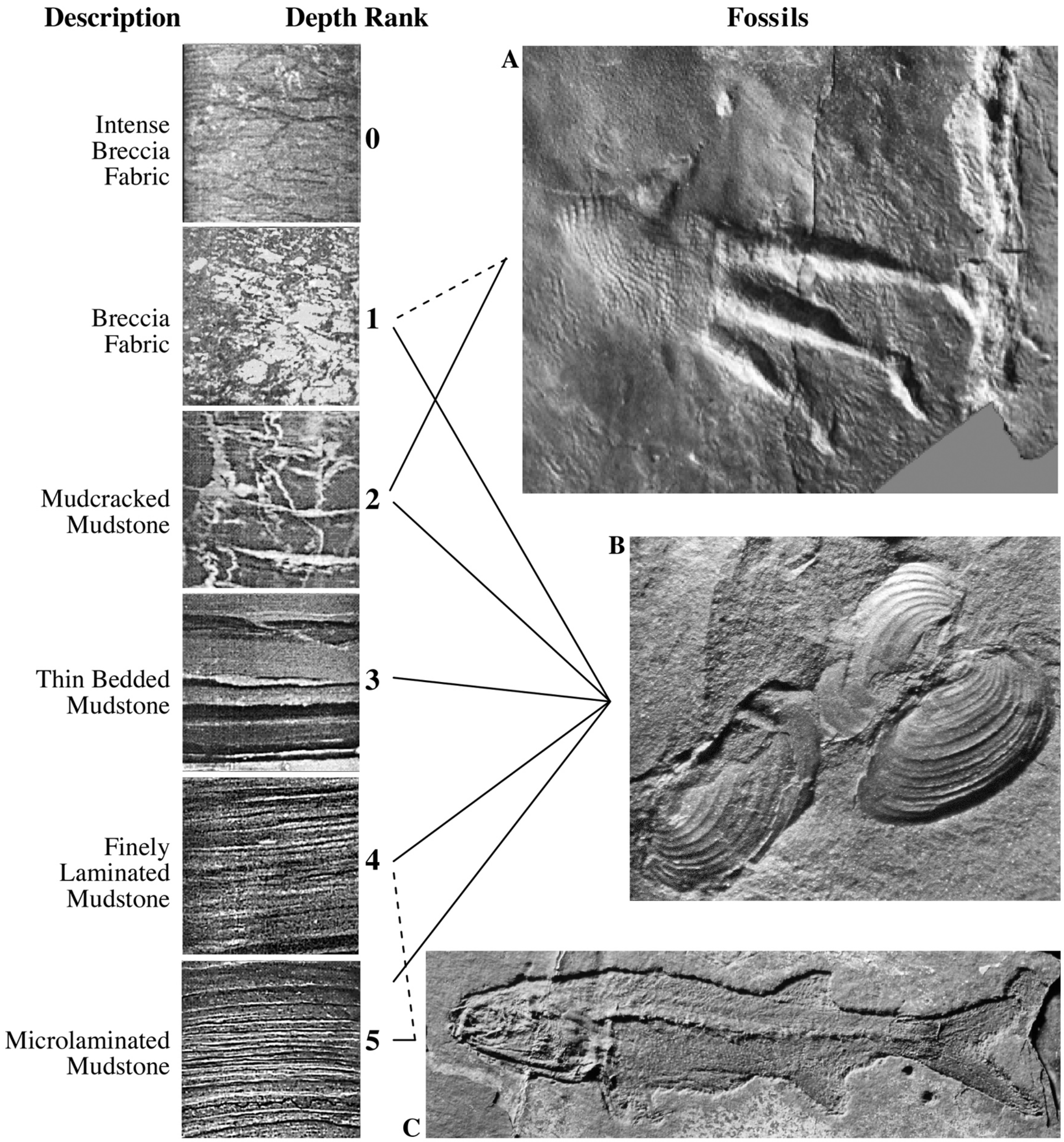


Figure 6: Sedimentary fabrics and fossils associated with various depth ranks (from Olsen & Kent, 1996). A, *Rhynchosauroides* cf. *hyperbates* from the Lockatong Formation; B, conchostracans from T-8 in the Argana basin; C, *Turseodus* sp. From the Lockatong Formation.

Houten cycles occurring during times of maximum precession variance. In the same cycles, footprints tend to occur in the shallowest water facies. In contrast, footprints tend to occur in the "deepest" water facies of those Van Houten cycles formed during times of lower precession variability, while few fossils at all are found times of low precessional variability in the overall 405 ky cycle.

The spectrum of Milankovitch cycles present in the Newark basin has allowed the development of a timescale for the Late Triassic (the Newark Basin Astronomically calibrated Geomagnetic Polarity Time Scale NBAGPTS: Figure 7). The historical development of this times scale is detailed in Olsen et al. (2010). The Milankovitch pattern and the NBAGPTS have withstood several independent quantitative tests

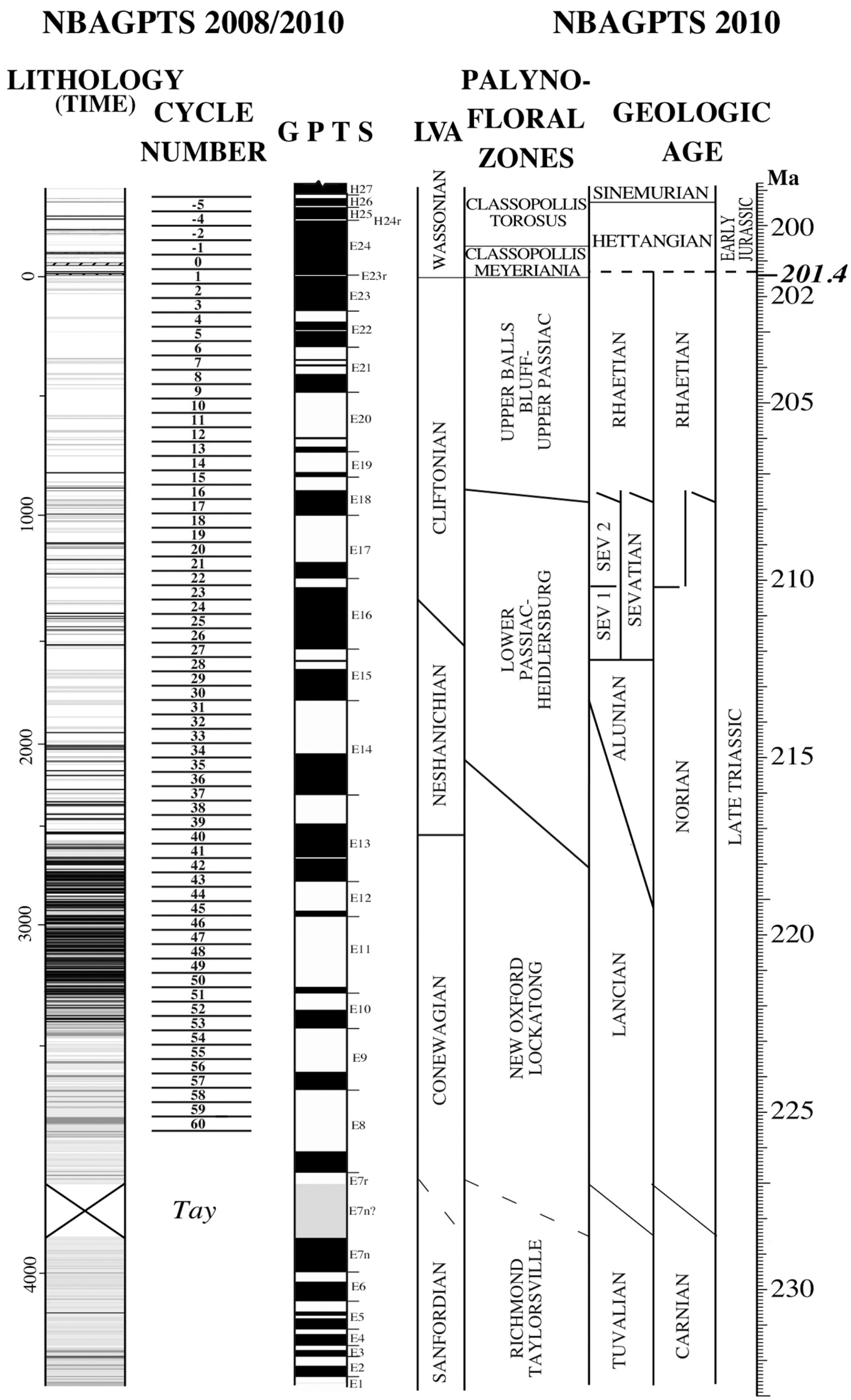


Figure 7: Newark basin astronomically calibrated geomagnetic polarity timescale (NBAGPTS). Modified from Olsen et al.. 2010.

(Kominz & Bond, 1990; Kominz et al., 1991; Baily & Smith, 2008) and independent radioisotopic ages registered to the NBAGPTS (Figure 8) via paleomagnetic polarity correlation (Furin et al., 2006; Olsen et al., 2010) as well as via ages from within the Newark itself (Rasbury et al. 2003; Blackburn et al., 2009). The NBAGPTS has also allowed determination of the duration of the stages of the Late Triassic (Muttoni et al., 2004, 2009; Channell et al., 2003) and earliest Jurassic (Kent & Olsen, 2009), as well as a guide to the duration of events during the Late Triassic and Early Jurassic such as the mass extinction near the close of the Triassic (Olsen et al., 2002a; Whiteside et al., 2007, 2010a), and duration of the CAMP episode (Olsen et al, 1996b, 2003).

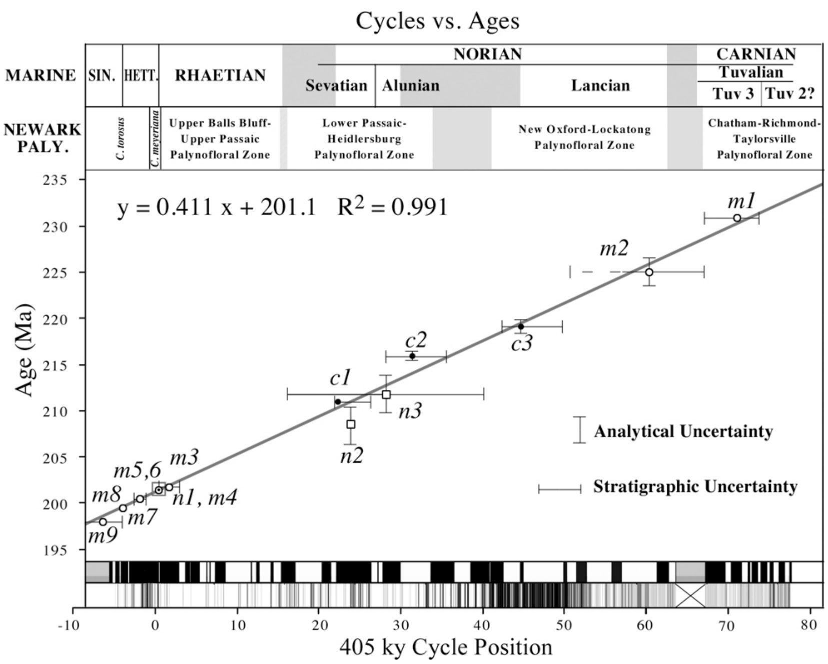


Figure 8: Correlation between number of putative 405 ky cycles in the Newark and Hartford basin sections (from Newark-APTS 2010) and 206Pb/238U ages with linear regression yielding a slope of 411±11 ky / cycle indistinguishable from the hypothesized 405 ky duration of the cycle. Ages are from marine sections (circles, m labels), the Newark and Fundy basins (squares, n labels), and the Chinle Formation (black dots, c labels). See Olsen et al., 2010 for list of ages used and their sources.

Size of the Great Lakes

Core and outcrop analysis of the lacustrine sequences in the Newark basin as well as other Newark Supergroup basins leaves little doubt that the high stands sequences of Van Houten cycles can be traced across the basin (Olsen et al., 1996). For the Newark basin, for Triassic strata this indicates the minimum area of these lakes was in the order of 7400 km^2 assuming the lake extended to the present edges of the basin exclusive of the "narrow neck". However, the Newark basin is deeply eroded (1-6 km: Malinconico, 2009), and its original size must have been considerably larger tan at present, but how much larger is a subject for debate.

The broad terrane hypothesis of Russell (1880) postulates that the Newark and Hartford basins, and other pairs of Newark Supergroup basins with boarder fault systems of opposing polarity were originally the sides of a full graben system that was postdepositionally arched along the graben axis and deeply eroded. In the original version of this hypothesis, the present geometry of the basins as half graben with strata tilting towards the border fault systems would be completely postdepositional. This concept was elaborated in a more modern form by Sanders (1963) who nonetheless retained the basic

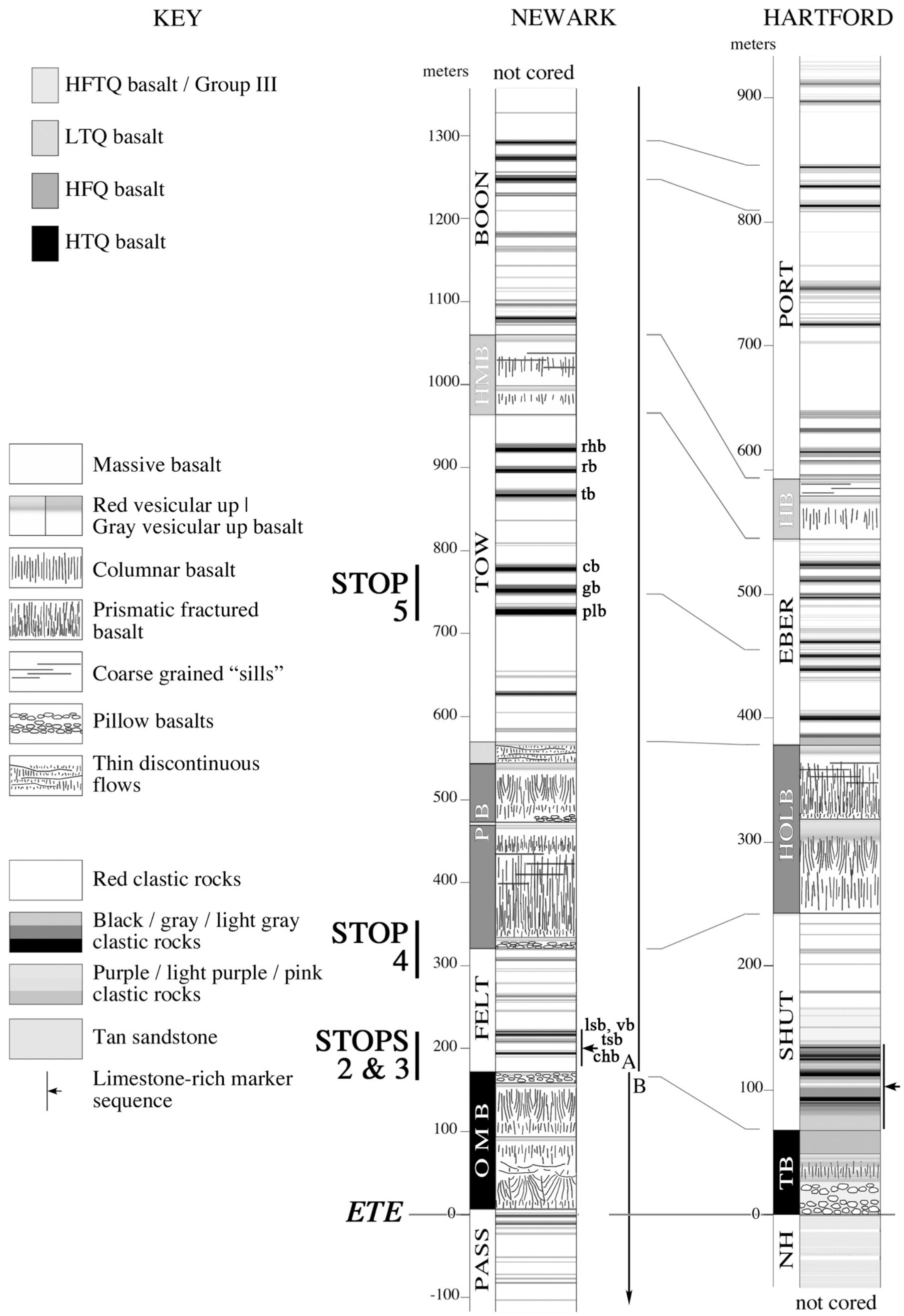


Figure 9: Measured Triassic-Jurassic sections from the Newark and Hartford basins showing the field trip stops and distribution of lacustrine strata of different colors. Darker colors generally conform to units exhibiting deeper water lacustrine facies, while more purple and red colors correspond to shallower and drier lacustrine facies, except in the cases of the New Haven (NH), Sugarloaf formations (SA), and Fall River (FB) beds, which are in part fluvial, and the white limestones that can also be partially fluvial. Abbreviations for lithological units are: BOON, Boonton Formation; EBER, East Berlin Formation; FELT, Feltville Formation; HB. Hampden Basalt; HMB, Hook Mountain Basalt; HOLB, Holyoke Basalt; NH, New Haven Formation; OMB, Orange Mountain Basalt; PASS, Passaic Formation; PB, Preakness Basalt; PORT, Portland Formation; SHUT, Shuttle Meadow Formation; TB, Talcott Formation; TOW, Towaco Formation; Abbreviations for measured sections are: A, Martinsville no. 1 core; B, ACE, Army Corps of Engineers cores (from Olsen et al., 1996b) Arrows with bar at apex are significant carbonate rich intervals. From Whiteside et al. (2007). Bed designations are: cb, Colfax Bed; chb, Cataract Hollow Bed; gb, Glen Bed; lsb, Lake Surprise Bed; plb, Pines Lake Bed; rb, Roseland Bed; rhb, Riker Hill Bed; tb, Totowa Bed; tsb; Trailside Bed; vb, Vosseller Bed.

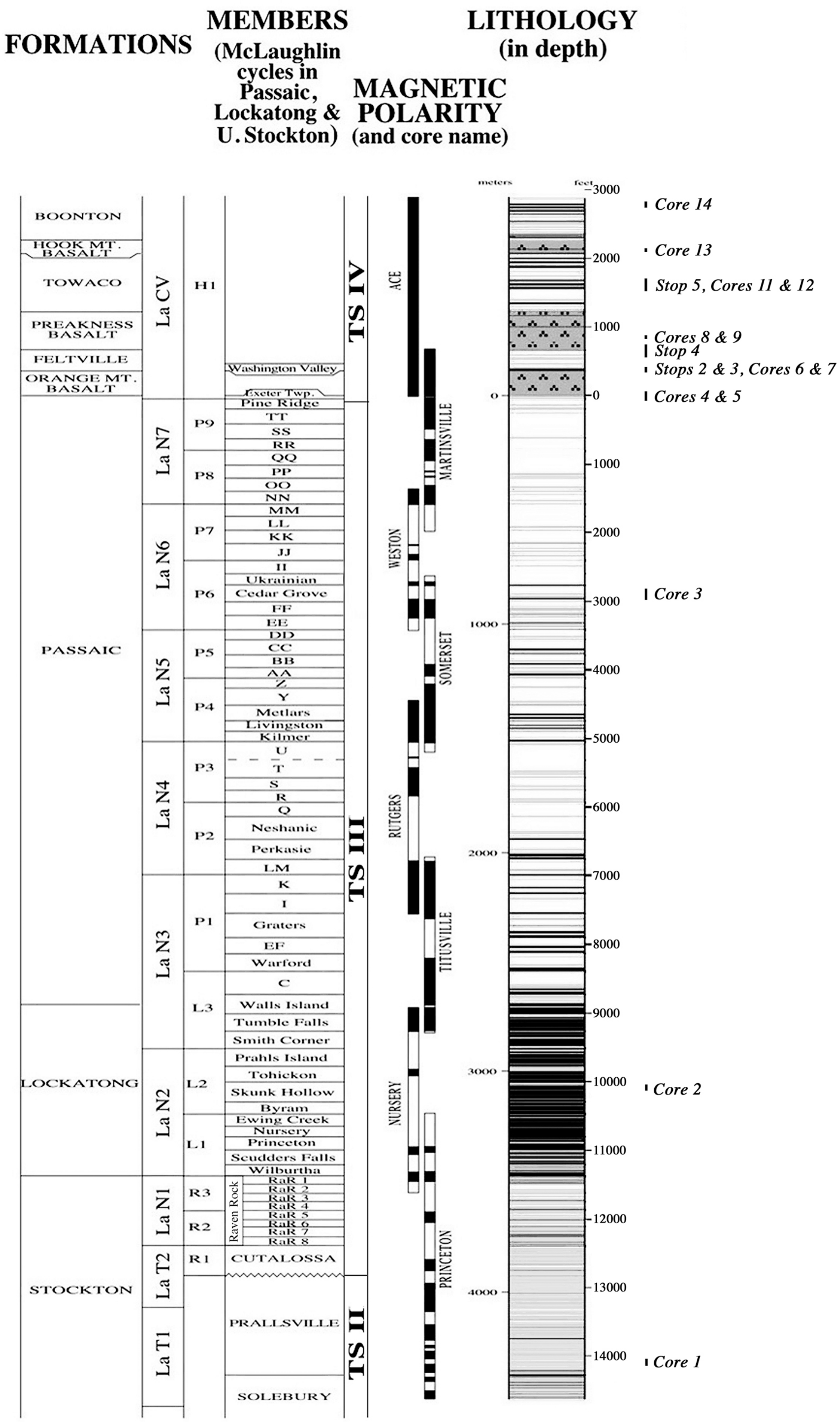


Figure 10: Newark basin sections showing position of stops and cores (Stop 1).

form of the idea. Ample data from industry reflection seismic profiles, correlation between cores, drill holes, and outcrops summarized by Olsen (1997), Schlische (2003), Schlische & Withjack (2005), and Withjack & Schlische (2005) shows that strata in the Newark and Hartford basin exhibit divergence towards the border fault systems and hanging wall onlap, demanding syndeposition assymetrical accommodation space growth (see Stops 1 – 3). In addition, the small Pomperaug basin in Connecticut has a very condensed section of all units relative to the Newark and Hartford basins and has clasts with local crystalline basement provenance (LeTourneau& Huber, 2006), which is inconsistent with its position relative to the original broad terrane model. As originally proposed the broad terrane model cannot be supported.

However, the degree to which the now isolated basin depositional centers were connected by continuous depositional systems during the Triassic and Jurassic remains largely unknown because of deep erosion. The same is true of the interconnectedness of the lakes themselves. McHone (1996) argued for a modified broad terranne hypothesis in which basalt flows extended over basement highs during the Early Jurassic extrusion of the CAMP. Condensed and abbreviated sections of sediment could well have connected the individual depocenters. In regions where post rift erosion has not proceeded as far as in eastern North America, such as Morocco, this is exactly what is seen (Medina et al., 2007). As many have pointed out the eastern North American basins may have looked topographically similar to modern rifting areas such as the Basin and Range or the East African rift system consisting of depocenters interconnected by broader expanses of condensed and abbreviated section.

In fact, as we have learned more about the details of the stratigraphy and geochemistry of the individual basins, some rather startling similarities present themselves, even in the face of clear data supporting the growth of local accommodation space and local provenance of sediment. In specific, there is a striking similarity in the syn-CAMP stratigraphic details in basalt geochemistry, paleomagnetics, and especially cyclostratigraphy (Puffer. 1992, 2003; Prévot & McWilliams, 1989; Olsen et al., 2003; Kent and Olsen, 2009; Whiteside et al., 2007) (see Figure 9). The recent discovery of a thin air fall ash (Pompton Tuff) in the precisely homotaxial cycle in the Hartford and Newark basins has shown that the lacustrine cyclicity of those two basins was linked at the decadal to seasonal level (Stops 1 and 5) and leaves no doubt at all of the synchronicity of the cycles. The remarkably tight environmental linkage this implies seems greater than we might expect of lakes with different watersheds and suggests either that the strength of regional climate and even weather variation overwhelmed local effects or that the lakes were actually the same water body during high stands.

Faill (2003) suggested that the Newark, Gettysburg, and Culpeper basins were part of a much larger more or less continuous basin and depositional system that he termed the Birdsboro basin. The present extent of the Birdsboro basin is nearly as large as Lake Tanganyika, Malawi, or Baikal in length and would certainly be as large as these basins in terms of what would be preserved in the geological record (Whiteside et al., 2010b). The stratigraphy of the Gettysburg and Culpeper basins is not nearly as well known as that of the Newark and Hartford basins, but what is known suggests the lakes of the Birdsboro basin could also have been connected at high stands. If there was one continuous lake during high stands of the syn-CAMP sedimentary units, this lake would be over 800 km in length and be larger than all but the largest of modern inland waters, the Black and Caspian Seas, and would be the largest rift lake known to exist on Earth.

End Triassic Mass Extinction

Whatever the origin of the CAMP, plume or otherwise, associated with the initiation of seafloor spreading or not, it is clear from recent high-precision U-Pb ($^{206}Pb/^{238}U$) dating of zircons from some of the oldest CAMP basalts and from marine strata bracketing the Triassic-Jurassic boundary (Pálfy & Mundil, 2006; Schoene et al., 2010) that eruption of the lavas was extremely and perhaps synchronous with the initiation of the mass extinction at the close of the Triassic, the end-Triassic extinction event (ETE)[4] and that the marine and terrestrial extinction were synchronous within the ±30 ky level of precision of the method (CA-ID-TIMS; Mattinson, 2005). The oldest basalt flows in the Newark Supergroup include the high titanium quartz normative (HTQ) basalts of Puffer and Lechler (1980) and Puffer (2003) notably the Orange Mountain Basalt of the Newark basin and the North Mountain Basalt of the Fundy basin, both of which lie above the initial part of the ETE (Fowell & Traverse, 1995; Olsen et al., 2002; Whiteside et al., 2010a; Cirrilli et al., 2009), with latter yielding a $^{206}Pb/^{238}U$ age of 201.38±0.02 Ma, an age indistinguishable from $^{206}Pb/^{238}U$ ages from ashes just above the last appearance of the Late Rhaetian guide ammonite *Choristoceras* (Schoene et al., 2010) and the latter are about 70 +220/–70 ky older than an ash above the first appearance of the GSSP marker ammonite *P. spelae*. This radioisotopic estimate is close to the more precise astrochrological estimate of 120 ky derived from the Newark and Hartford basins (Whiteside et al., 2010a) and 140-160 ky derived from the marine St. Audries Bay section in Somerset England (Ruhl et al, 2010). In any case, in terms of Newark basin stratigraphy, the position of the Triassic-Jurassic boundary should lie within the middle Feltville Formation (Figure 10). This movement of the boundary to within the extrusive zone certainly makes matters difficult for geo-cartographers but there is no apparent escape.

Proposed mechanisms for the ETE via the CAMP generally involve abrupt injections of gasses into the atmosphere causing catastrophic climate change (see review by Rampino, 2010), and oceanic acidification by CO_2 absorption (van de Schootbrugge et al., 2007), but direct poisoning by CAMP-related gases has also been proposed (Svensen, 2009). The climate-altering gasses could have included just CO_2, or methane-derived

[4] Formerly the ETE was considered synonymous with the Triassic-Jurassic boundary. However, the recent establishment of the GSSP (Global boundary Stratotype Section and Point) of the base Hettangian at the marine section at Kuhjoch, Austria (Northern Calcareous Alps) at the first occurrence of the ammonite *Psiloceras spelae* (Hillebrandt et al., 2007; Morton, 2008a, 2008b; Morton et al. 2008) and its ratification by the IUGS in 2010 (Morton, 2010), defines the ETE as a Late Triassic event. By onset of the ETE, or the initial ETE, I mean the first wave of extinctions, which in eastern North American paleotropics include the vessicate pollen taxon Patinasporites densus and its "relatives", and the footrint taxon *Brachychirotherium*, and in the marine realm includes the last appearance of the ammonite *Choristoceras*.

from methane clathrate dissociation[5] (Hessebo et al., 2002), that would oxidize to CO_2, or thermogenic methane derived from CAMP intrusions into organic-rich sediments[6] (Svensen, 2009), that would oxidize to CO_2, all of which would cause global warming – a super-greenhouse - that could have lasted thousands of years for *each* large eruption in the CAMP. Sulfur aerosols have also been proposed as a CAMP eruptive product that would have caused global cooling (Schoene et al., 2010), but with a much shorter recovery time on a per-eruption basis. Global warming, higher CO_2 and related phenomena may also have caused ocean anoxia (Hallam & Wignall, 1997). Toxic thermogenic chorine or fluorine could have been released by intrusions injected into evaporates (Svensen, 2009). Some scenarios for CAMP-related extinction have become very complicated indeed involving thermal expansion-driven transgressions (Kidder & Worsley, 2010). All these are possible, even, plausible causes, but what is needed is direct evidence of the timing of the CAMP relative to the extinctions and observable effects of specific proposed mechanisms.

There is in fact evidence for a sharp rise in CO_2 at the ETE, based on plant leaf stomatal data from Greenland and Sweden (McEwain et al., 1999) and Germany (Bonis et al., 2010). This rise appears to coincide (in Greenland for sure) with a dramatic drop in plant diversity (McEwain et al. 1999, 2009). In Greenland, the data indicate a fourfold increase in CO_2 from pre-ETE levels of 600 to 2100-2400 ppm during the ETE corresponding to a 3° to 4°C super-greenhouse warming (McEwain et al., 1999). Unfortunately, these data are from beds lacking any CAMP volcanic products and thus the relation to the timing of CAMP are highly inferential involving long distance correlation (e.g., Whiteside et al., 2010a).

It is thus important to be clear about what we actually know of the timing relationship between the CAMP and the onset of the ETE. Presently, in North America there are what have been taken to be Jurassic-aspect pollen and spore assemblages between the ETE and the oldest CAMP lavas in all of the basins producing pollen below the oldest basalt (Fowell, 1994; Fowell & Traverse, 1995; Olsen et al., 2002; Whiteside et al., 2007, 2010a). This has been taken to indicate that the ETE could not be caused by the CAMP eruptions because the extinction *preceded* the proposed cause (Fowell et al., 2007; Whiteside et al., 2007).

However, even the initial ETE could still have been caused by the CAMP despite the eruptions postdating the beginning of the extinction - if there were a lag between the intrusion of CAMP sills and the surface eruption of the lavas, with venting of thermogenic environmentally deleterious gases preceding the eruption of the flows - which is not an unreasonable scenario. Indeed we know that CAMP sills did metamorphose a large body of organic rich strata, notably the Triassic Lockatong Formation[7] in the Newark basin (Van Houten, 1969), organic rich Paleozoic shales and

---

5 Originally proposed for the Paleocene-Eocene thermal maximum (PETM) by Dickens et al. (1997).

6 Also proposed for the PETM by Svensen et al. (2004).

7 The entire volume of the metamorphic aureole of the CAMP Palisade sill complex, including its continuation into Pennsylvania (Quakertown and Coffman Hill intrusions etc.) and New York is an order of magnitude smaller than that modeled for the North Atlantic Vøring and Møri basins for the PETM (<u>very</u> roughly 3.6×10^{12} m^3 for the Palisade sill and 3.5×10^{13} m^3 for the PETM). Even if the Palisade complex was intruded entirely into the Lockatong, which it is

evaporites in the Amazonian basin (Milani & Zalán, 1999; Barata & Caputo, 2007) Svensen et al., 2009), and CAMP intrusions also invaded Triassic evaporites in Morocco (M. Et-Touhami, pers. com.). However, proxies for halide or halocarbon gasses are not known.

However, there is some evidence that some CAMP flows were in fact coeval with the initial ETE. Marzoli et al. (2004) described palynoflorules in the Argana and High Atlas basins in Morocco of Triassic aspect directly in contact with the oldest basalts, which if true would argue that those basalt were slightly older than the oldest basalts in Eastern North America and indeed could have been contemporaneous with the onset of the ETE and therefore could have caused it. More recent work by Whiteside et al (2010a) confirms two cases in Morocco (Khemisset and Central High Atlas basins) that conform to Marzoli's observations[8]. In addition Cirrilli et al. (2010) have discovered pollen and spore taxa immediately above the North Mountain Basalt in Nova Scotia typical of the Northern European expression of the latter part of the ETE and by definition of latest Triassic age (in as much as the first appearance of *P. spelae* is above in the same sections). This seems to indicate that extinctions *persisted* after the oldest basalts even in eastern North America. This is also indicated by the stable carbon isotope data of Whiteside et al. (2010a) that suggests that the extinctions of the conodonts in marine strata followed shortly after the extrusion of the earliest Newark basin basalts. Finally the exposed lavas of the CAMP are but a small portion the entire provinces and slightly older CAMP lavas could in fact be widespread.

In 2002 (Olsen et al., 2002) my colleagues and I showed that in eastern North America, tetrapod extinctions, based largely on the spectacular Newarkian track record, was synchronous with the floral extinctions (Figure 11) and a "fern spike". We related this extinction event to a modest iridium anomaly that we reasoned by analogy to the K-T iridium anomaly could have been produced by the impact of a giant bollide and caused the mass extinction (Figure 12). Subsequent work in the Fundy basin by Tanner and Kyte (2005) and Tanner et al. (2008) showed that there were multiple Ir anomalies associated with the initial ETE in that basin and resampling of a much larger stratigraphic section around the ETE in the same section in the Newark basin reported on in 2002 also showed there to be multiple Ir anomalies, each associated with a redox boundary (unpublished data by N. Nair and P. E. Olsen). This argues against a bolide impact for the origin of the ETE, and at the present time there is no convincing evidence of a role for impact at this

---

not, and even if the Lockatong Formation were as rich in organic carbon as the North Atlantic Cretaceous and Paleocene mudstones (TOC averaging ~1.4%), which it probably was not, the expected amount of thermogenic methane would still be an order of magnitude less than that postulated for the PETM (following all the same assumptions as Svenson et al. (2004) and thus unlikely to be an important cause for the extinction by itself. No other large volumes of organic-rich strata intruded by giant sills are known to be present in the Newark Supergroup.

[8] That said, Whiteside et al. (2007) presented data from the Argana basin that differed from that of Marzoli et al. (2004) by lacking vessicate pollen in the samples just below the basalt, while still having abundant other pollen, thus conforming to the eastern North American pattern. The same kind of observation was made by Deenen et al. (2010) in a different section in the Argana basin. This may be due to differential thermal degradation of the vessicate forms adjacent to the basalt as suggested by Deenen et al. (2010).

mass extinction event. In my view the role for the CAMP in the ETE is much more compelling, but requires further testing and exploration.

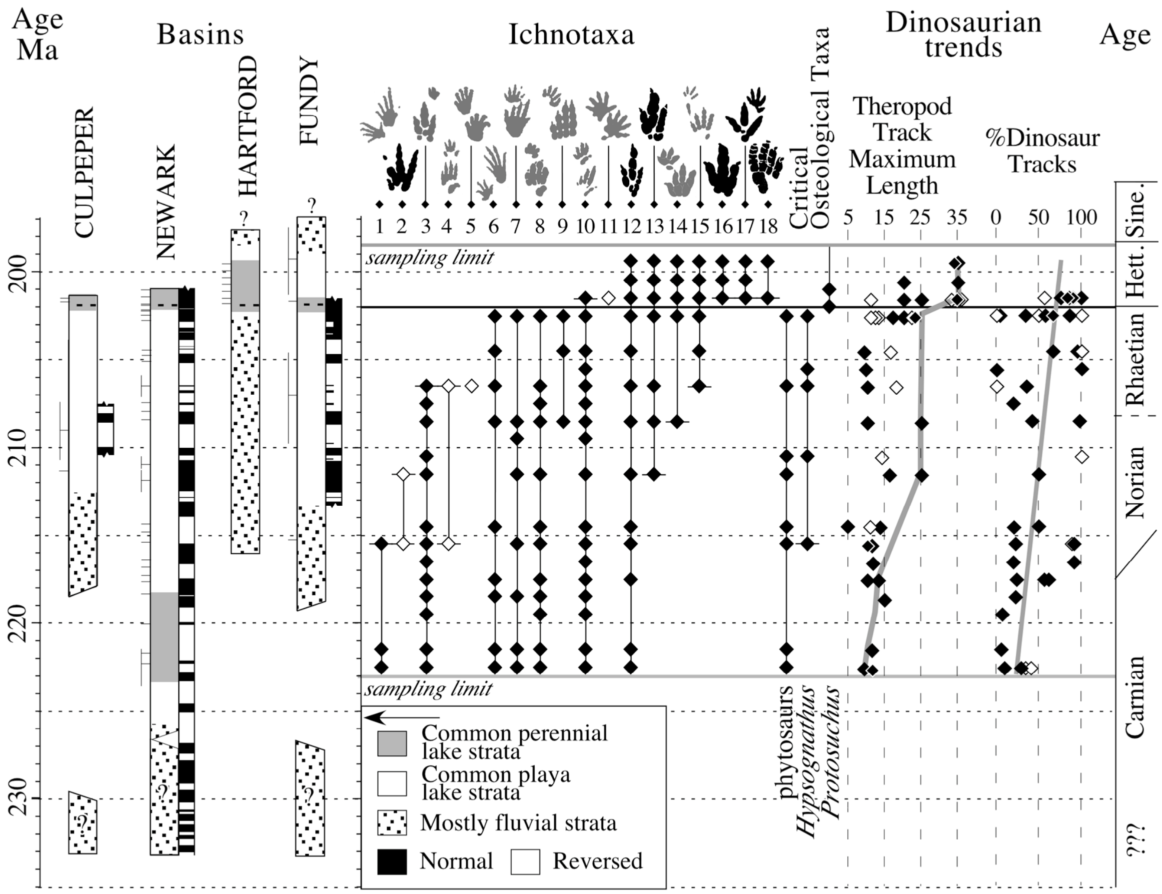


Figure 11: Correlation of four key basins of the Newark Supergroup showing the temporal ranges of footprint ichnogenera and key osteological taxa binned into 1-My intervals showing the change in maximum theropod dinosaur footprint length (line drawn through maximum) and percent at each 1-My level of dinosaur taxa. Short, horizontal lines adjacent to stratigraphic sections show the position of assemblages, and the attached vertical lines indicate the uncertainty in stratigraphic position. Solid diamonds indicate samples of footprints, and open diamonds indicate samples with <10 footprints. Horizontal, dashed gray lines indicate the limits of sampling; thick gray line indicates trend in maximum size of theropod tracks; ?, age uncertain. Ichnotaxa are as follows: 1, *Rhynchosauroides hyperbates*; 2, unnamed dinosaurian genus 1; 3, *Atreipus*; 4, *Chirotherium lulli*; 5, *Procolophonichnium*; 6, *Gwyneddichnium*; 7, *Apatopus*; 8, *Brachychirotherium parvum*; 9, new taxon B (8); 10, *Rhynchosauroides* spp.; 11, *Ameghinichnus*; 12, "*Grallator*"; 13, "*Anchisauripus*"; 14, *Batrachopus deweyii*; 15, "*Batrachopus*" *gracilis*; 16, *Eubrontes giganteus*; 17, *Anomoepus scambus*; and 18, *Otozoum moodii*. Ma, million years ago; Hett., Hettangian; Sine., Sinemurian. Modified from Olsen et al. (2002).

Within the Newark basin are several features of the sedimentary record that appear to be tied to the climatic and tectonic events related to the ETE and CAMP episode and these are illustrated at our field stops and cores (Figure 10). The first feature is readily seen in the cores at Stop 1 and outcrops as Stops 2, 3, and 5. Based on the pervasive cyclicity, the accumulation rate of the interflow units increases from the upper

Passaic Formation by a factor of more than 4 (5 m per cycle to 20 m per cycle). This is despite the fact that accommodation space was being filled very fast by CAMP lava flows. Second, the deeper water units within sedimentary cycles in the Feltville Formation are very carbonate-rich (Stops 1, 2 and 3), in contrast to the succeeding cycles in the overlying Towaco and Boonton Formation or the underlying Passaic or Lockatong formations. A very similar pattern is seen in the Hartford basin of Connecticut and Massacussetts, the Culpeper basin of Virginia, the Fundy basin of Nova Scotia and New Brunswick Canada, and even in the Moroccan basins (Whiteside et al., 2007). It seems plausible that these carbonate beds are related to elevated weathering rates of early CAMP basalts under intense super-greenhouse conditions with resulting heavy loading of bicarbonate and calcium from the basalt in the Feltville and coeval lakes. Also peculiar to the Feltville Formation, but similar to its correlates in other basins is the abundance of the large leathery-leaved dipteridaceous fern *Clathropteris meniscoides* (Figure 13) and elevated levels of fern spores. Both of these features are also characteristic of the strata underlying the Orange Mountain Basalt but above the ETE. Again this is plausibly related to elevated CO_2 and perhaps also to increased frequency of fire which in turn increased the frequency of early successional plant groups. Cornet (1975) showed that *Clathropteris* was associated with cherolepidaceous conifers (*Brachyphyllum* spp.) that had unusually short and small leaves with thickened papillate cuticle. This is consistent with adaptations to aridity, but the associated sediments show no signs of unusual aridity such as evaporites, suggesting that perhaps the adaptations were not to aridity *per se*, but rather to high heat stress and high canopy evaporation levels, inline with the thermal damage hypothesis of McElwain et al. (1999) for the ETE super-greenhouse.

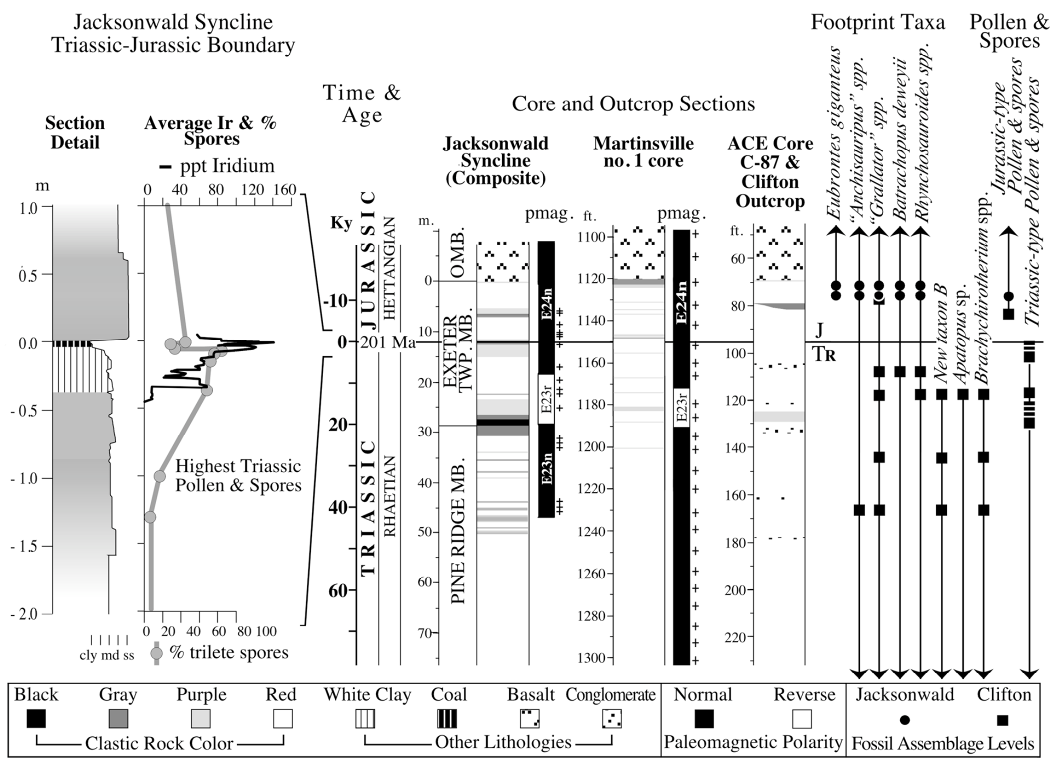


Figure 12: Details of the ETE in the Newark basin (from Olsen et al., 2002).

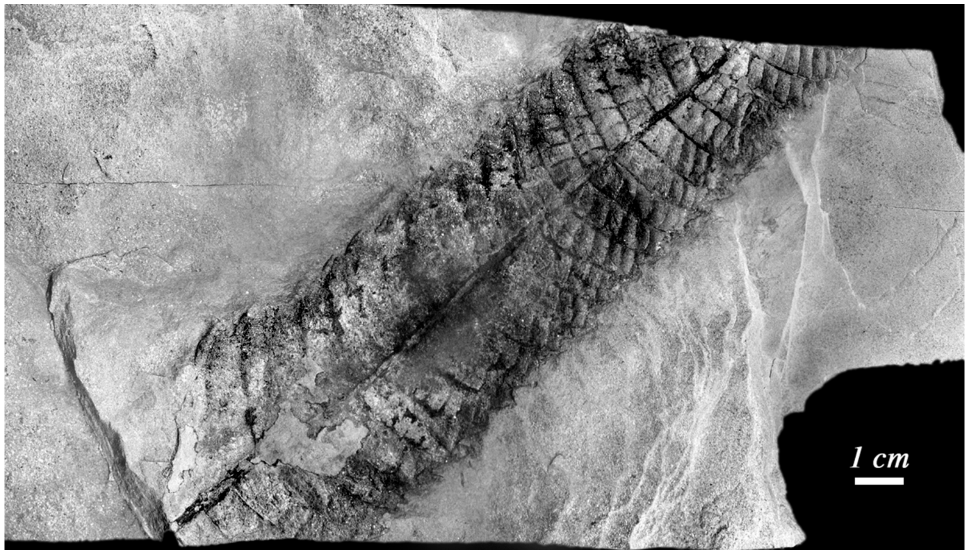


Figure 13: Example of the fern *Clathropteris* from the cycle bearing the Cataract Hollow Bed, lower Feltville Formation, Stop 3, Station 3.

Ascent of the Dinosaurs

Ironically, the ETE decimated the diversity of tetrapods at the end of the Triassic, notably hitting the curotarsians[9] (crocodile relatives) especially hard (Brusatte et al., 2008a, 2008b), but afterward dinosaurs and crocodiliomorph crurotarians became ecologically dominant, and stayed that way until the close of the Cretaceous. The ecological ascent of the dinosaurs is particularly well-displayed, albeit somewhat indirectly, in the strata of the Newark basin, where the transition in tetrapods is seen in particular detail thanks to the extremely abundant footprints around the ETE. Olsen et al. (2002) examined the transition using the NBAGPTS for time calibration (Figure 11). The last appearances of the larger curotarsian tracks such as *Brachychirotherium* and *Apatopus* were within 30 ky of the floral extinction marking the ETE. In the oldest assemblages above the floral transition marking the initiation of the ETE, and below the oldest Newark basalt (Orange Mountain Basalt) the size of the theropod dinosaur tracks is 20% larger than anytime before, corresponding to the oldest *Eubrontes giganteus*. Olsen et al. (2002) argued that this could be explained either by character release

[9] Crurotarsians are members of the Crurotarsi (often called crocodile-line archosaurs), closely related to the concept of the Pseudosuchia, which include modern crocodilians, and are the sister group to the Ornithodira (or bird-line archosaurs), which include the non-avian dinosaurs and their near relatives as well as birds. The Crurotarsi are defined (Sereno, 1991; emended by Sereno, 2005) as all forms closer to *Crocodylus niloticus* (Nile crocodile) than to Passer domesticus (the House Sparrow) and thus includes all forms evolutionarily closer to crocodiles than to birds.

following ecological release[10] after the extinction of dinosaurian competitors, or by emigration of the larger theropods from elsewhere, but favored the former hypothesis. This transition from a cruotarsian-dominated ecosystem to a dinosaur dominated one marks the true ecological ascent of the dinosaurs that had evolved at least 30 million years earlier.

Lucas et al. (2006) and Lucas & Tanner (2007a,b) have argued that "… tracks of large theropod dinosaurs assigned to *Eubrontes* (or its synonym *Kayentapus*) are known from the Triassic of Australia, Africa (Lesotho), Europe (Great Britain, France, Germany, Poland–Slovakia, Scania) and eastern Greenland, invalidating the "ecological release" hypothesis…" (Lucas & Tanner, 2007a,b). However, the concept of *Eubrontes* used by these authors is broad and encompasses basically all of the larger brontozooid (or grallatorid) kinds of footprints, while the ecological release hypothesis is framed using only the largest of these, the ichnospecies *Eubrontes giganteus*, which is the type species (Olsen et al., 1998), the other species within the genus being explicitly excluded. Of the occurrences listed by Lucas of *Eubrontes*, none can be assigned to *E. giganteus* except the Scania (Sweden) occurrence. The latter is part of an assemblage of brontozoid tracks from the roof of the Gustaf Adolf coal mine in Höganäs and was originally described by Bölau (1952). The sandstone natural casts in the roof of the mine are from the Rhaetian Bjuv Member ("lower" or "B" of coal seam) of the Höganäs Formation. This interval in other mines produces the upper occurrences of the seedfern *Lepidopteris*. This taxon has been used as the guide fossil for the upper Rhaetian, and hence it has been assumed that the tracks are Triassic in age. I see no reason to doubt the association of *Lepidopteris* and *Eubrontes* giganteus at this locality. If this occurrence were indeed earlier than the ETE it would in my view indeed falsify the ecological release hypothesis. However, as McElwain et al. (2009) has shown, the ETE begins before the last appearance of *Lepidopteris*, which in fact has its last occurrence at the end of the ETE in Greenland, and therefore this occurrence of *Eubrontes* is probably approximately contemporaneous with the first occurrence in Eastern North America. There is no evidence of *Eubrontes giganteus* below the ETE and the ecological release hypothesis still stands as viable[11].

Interestingly, herbivore tracks, unlike later Jurassic assemblages are very rare, in fact absent from the earliest *Eubrontes giganteus* localities in the Passaic and lower Feltville formations. Skeletal evidence also suggests herbivores were relatively rare. These, just-post-ETE theropods, were probably opportunistic carnivores, and may have been part of a largely aquatic-based ecosystem. There is some skeletal evidence for this in the form of apparent fish-eating adaptations among the post-ETE theropods (Milner &

---

[10] Ecological release is a term coined by Wilson (1961) for the release of competitive (and selective) pressure that occurs when a species arrives on an island devoid of its competitors from whence it came. While couched originally in terms of geography and island biogeography (MacArthur & Wilson, 1967) it can easily be viewed in terms of an extinction of competitors.

[11] Gigantism is an often-cited consequence of ecological release and a particular good example of which is the recently extinct giant eagle (*Harpagornis moorei*) of New Zealand derived from a much smaller Asian/Australian form 1.8 million years ago (Scofield & Ashwell, 2009). This example provides a potential way to test the ecological release hypothesis in theropod dinosaur skeletons, because Scofield, & Ashwell (2009) show brains size lags behind body size and thus we would predict the encephalization quotient of the early very large theropods would be below expected for all theropods.

Kirkland, 2007). In any case, the post-ETE track assemblages are the oldest in which dinosaurs, particularly large theropod dinosaurs, were unquestionably ecologically dominant.

Evolution of Species Flocks

While the dinosaur tracks of the Passaic Formation and the shallow-water portions of Van Houten cycles in succeeding formations interbedded with the lavas of the CAMP document the one set of the effects of ecological release on the ascent of the dinosaurs, the deep-water portions of the Van Houten cycles formed during times of high precessional variability show the effects of ecological release on fish assemblages, in this resulting in the proliferation of closely related species in the same area in this case a lake, called a species swarm or flock (Mayr, 1963). The genus containing these species flocks is *Semionotus*, a holostean fish related to living gars (Figure 14). The Feltville, Towaco, and Boonton formations all have species flocks of *Semionotus*. McCune (1990, 1996, 2004), and we will visit 3 exemplar localities in the former two formations (Stops 2, 3, 5).

McCune (1996, 2004) has described the species flocks of semionotids, mostly in Towaco and Boonton forms, and to a lesser extent in the Feltville. Morphological variation in *Semionotus* species in a single Van Houten cycle is largely in body shape and details of dorsal ridge scale pattern. Similar body shapes of *Semionotus* occur within different clades in successive cycles, strongly suggesting iterative evolution.

Other taxa than semionotids are less common but include rare *Ptycholepis* cf. *marshii* and *Redfieldius* sp. in the Feltville Formation and one specimen of *Ptycholepis* new sp. and relatively common *Redfieldius* spp. and rare *Diplurus longicaudatus* in the Boonton Formation (Figure 15). In this assemblage, the semionotids probably were in a large variety of niches involving visual selection. *Redfieldius* has a mouth and jaw suspension apparatus suggestive of modern planktivorous taxa. *Ptycholepis* was probably a small-prey piscivore. *Diplurus* was a top predator and its distinctive phosphatic coprolites composed of digested fish bones are often present in large numbers.

It is difficult to assess diversity in the non-semionotid fishes because they are so uncommon, with the exception being *Redfieldius* in the Boonton Formation that does seem to be represented by several species. One of the points made by my 1980 NYSGA paper was that it is very difficult to assess true diversity because even in the face of a very large numbers of specimens, such as the 3000 of so specimens collected by McCume most taxa are rare, so rare they are not sampled at all.

The diversification of *Semionotus* species flocks in the latest Triassic and Early Jurassic Newark lakes reflects another example of ecological release followed by character release and adaptive radiation, perhaps analogous to the macroevolutionary pattern seem among tetrapods, particularly dinosaurs after the ETE. Except in the case of the semionotids it happened again and again as lakes waxed and waned tracking the orbital cycles, all without any perceptible directional change, closely analogous to the cichlid fishes in the East African Great Lakes, which also fluctuate dramatically to orbitally driven climate changes (Johnson et al., 1996; Cohen et al., 2007).

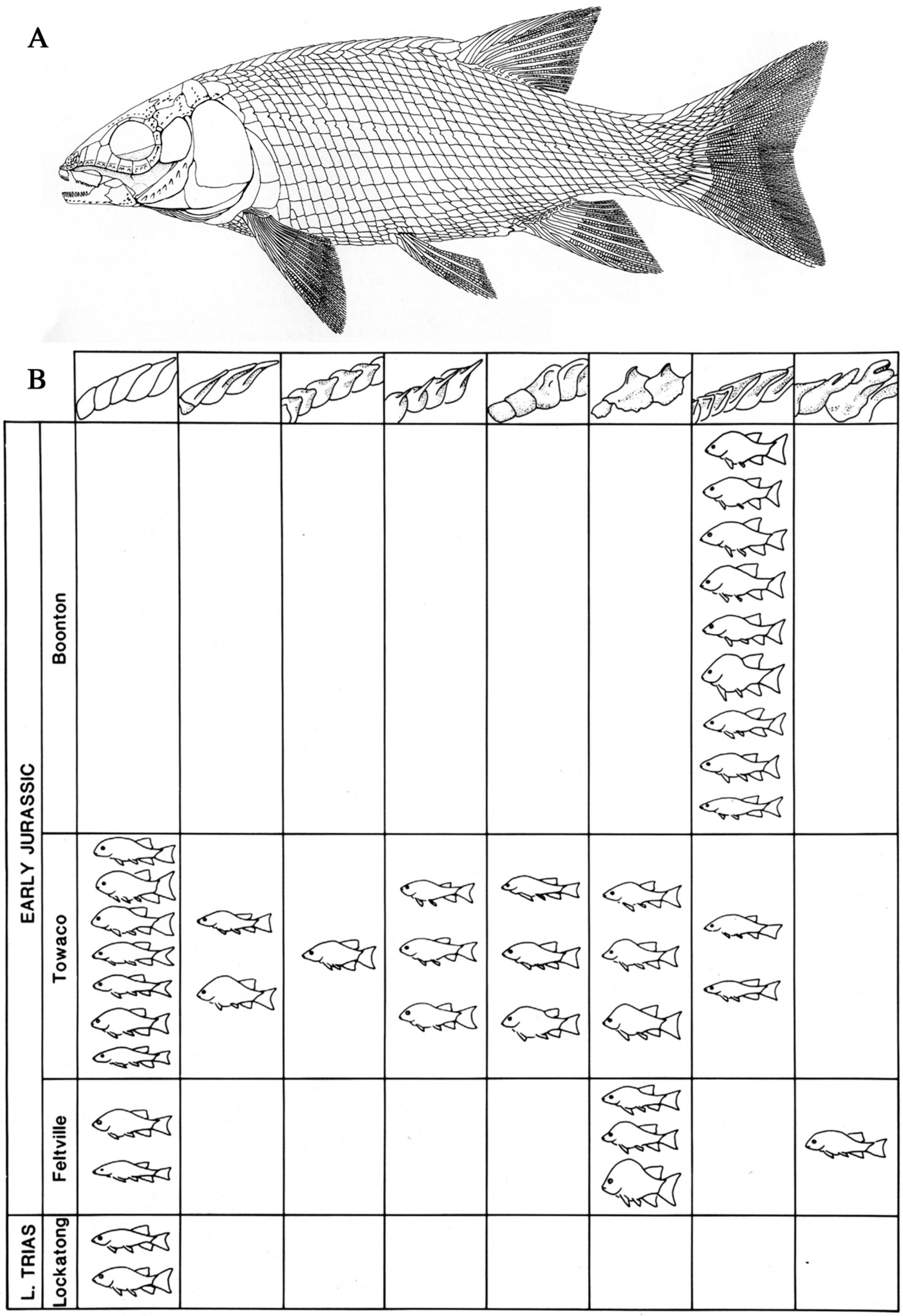


Figure 14: *Semionotus* (A) and species flocks of *Semionotus* (B) of the Newark basin, from Olsen & McCune (1991) and McCune et al. (1984).

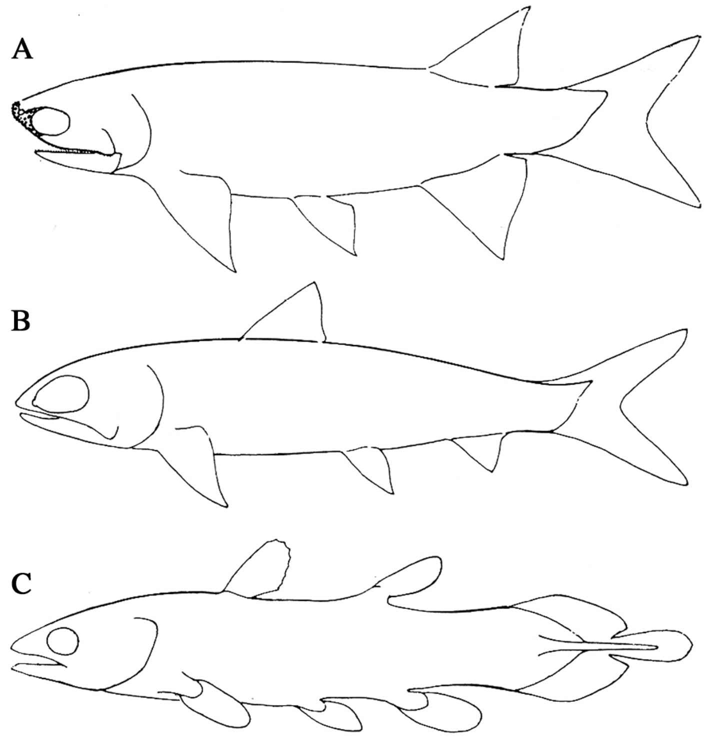


Figure 15: Other non-semionotid fish genera from the syn- and post-CAMP interval. A, *Redfieldius*; B, *Pycholepis*; C, *Diplurus longicaudatus*. From Olsen et al. (1982).

Chaotic Evolution of the Solar System

Milankovitch cyclicity is caused by variations in the orientation of the Earth's spin axis and shape and orientation of the Earth's orbit caused by the gravitational interactions of bodies in the Solar System following the laws of celestial mechanics. These were first formulated in their simplest form by Kepler and then explained within the context of calculus and the law of universal gravitation by Newton and a plethora of subsequent workers. With analytical solutions of the one and two body problem solved in the 18th century it seemed that the motions of the celestial bodies could be projected forward and back infinitely in a clock-work-like solar system with just a little more mathematical elaboration. However the 3-body and n-body problems proved intractable[12] and in the 19th century Poincaré showed that they are fundamentally not soluble using differential equations and in the process laid the foundations for the modern theory of deterministic chaos[13]. The Solar System has proven not in fact to be clock-work-like, but rather is chaotic on timescales of 10s of millions of years (Laskar, 1999). Thus, numerical solutions of the gravitational problem for the Solar System diverge completely after 10s of millions of years with extremely minor input variations which means it is impossible to construct accurate curves of insolation for the Earth for the Triassic or Jurassic or even to predict the periods of the climate cycles of eccentricity and obliquity modulating

[12] The n=3 and the n > 3 problems were solved in specific cases by Sundman (1912) and Wang (1991), respectively.
[13] A very readable book reviewing the interwoven history of celestial mechanics and chaos theory is by Diacu & Holmes (1996).

cycles greater than 400 ky, particularly the cycles caused by the interaction of Earth and Mars. In addition a significant percentage of the solutions (~1%) exhibit catastrophic singularities for the inner planets after 100s of millions of years including planetary close encounters, collisions, and ejections (Batygin & Laughlin, 2008; Laskar & Gastineau, 2009).

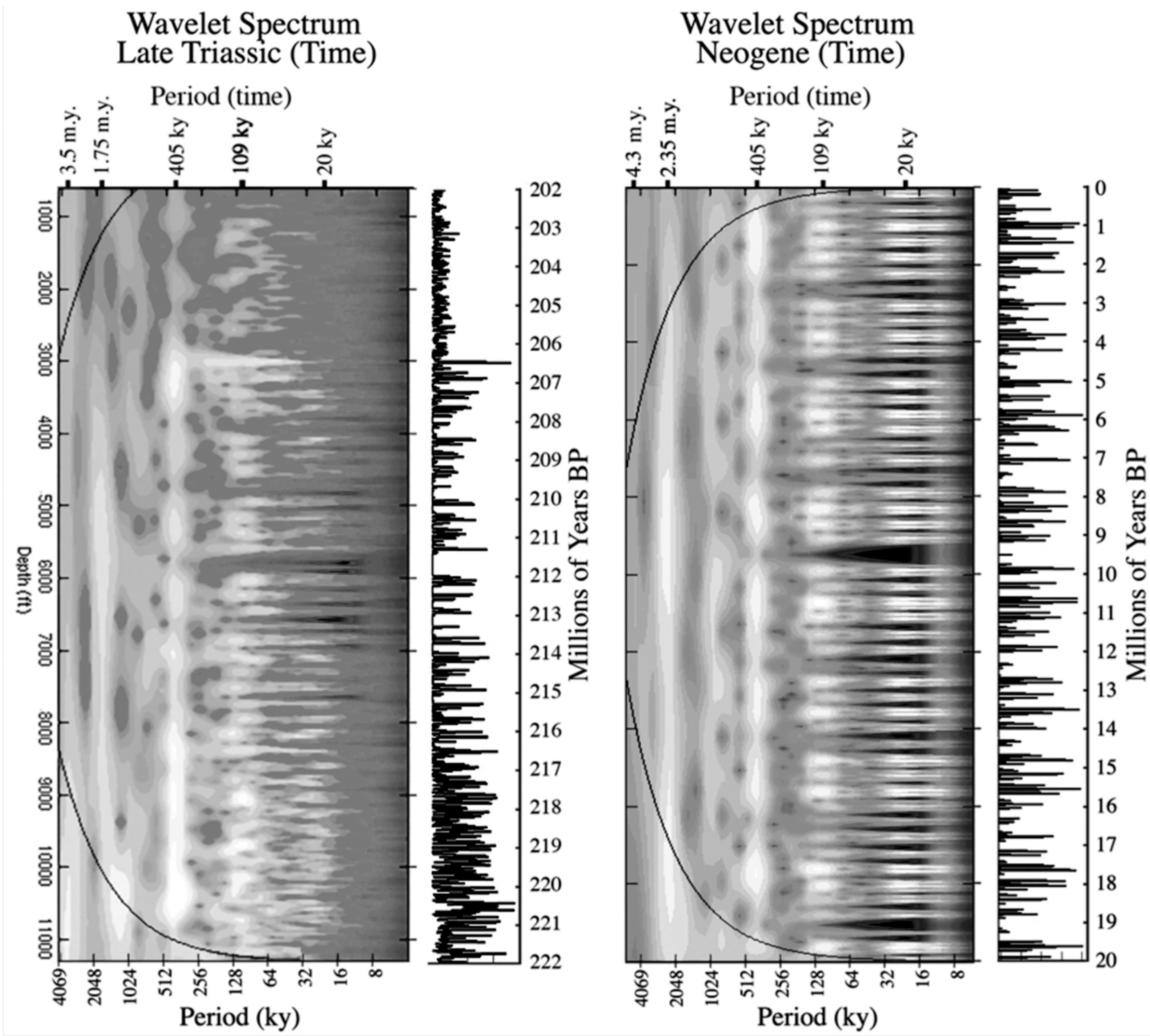


Figure 16: Comparisionn of wavelet spectra of depth rank data from the Passaic and Lockatong formations of the Newark basin and a clipped insolation curve of the last 20 m.y. (based on Laskar et al., 2004). White denotes highest spectral power.

While chaotic diffusion prevents the recognition of accurate solutions of planetary behavior and hence accurate predictions of long-period climatic cycles, those planetary interactions did occur in the past and the geological record of climate change related to those cycles provides a way to limit the range of possible solutions. Presently radioisotopic dates do not provide a direct mechanism of calibrated geological records at the precision necessary to select among of potential solutions. However one celestial mechanical cycle is close to metronomic accuracy on scales of 100s of millions of years, and that is the eccentricvity cycle caused by the interaction of Jupiter and Venus which has a very constant period averaging 405 ky with a maximum error of less than 500 ky projected back to 250 Ma (Laskar et al., 2004). Of course this cycle is the same as what was used to tune the NBAGPTS and which corresponds in lithology to the newarkian McLaughlin cycle.

Tuning the Newark basin record to the 405 ky cycle reveals longer term cycles of 1.8 and 3.5 m.y. (Olsen & Kent, 1999). A comparison between a wavelet spectrum of the tuned Newark lacustrine relative water depth record and 20 m.y. of the Laskar et al. (2004) reveals that the homologous modern cycles have periods of 2.4 and 4.5 m.y. (Figure 16) and are caused largely by the interaction of Earth and Mars. The differences between the current and Triassic values are well within the range of chaotic diffusion. That the Triassic values for these cycles are indeed reasonable and not due to periodic episodes of non-deposition as suggested in a review of my 1999 paper by F. Hilgen (Olsen & Kent, 1999) is shown by the correspondence between the number of McLaughlin cycles present and U-Pb dates from the Newark or correlated to the Newark by paleomagnetic polarity stratigraphy (Figure 8). Thus, the gravitational solutions must conform to this data, while current published versions do not.

In map view, the 1.8 m.y. cycle shows up as clusters of the named members of the Passaic Formation and the ones with the majority of the fossils. These same intervals, which derive from times of maximum precession variability generally support topographic highs. This cycle provides a natural way of segmenting the stratigraphy at a large scale.

Interestingly, one of the times of maximum precessional variability in the 1.8 m.y. cycle begins at the ETE. This is not to infer that the 1.8 m.y. cycle caused the ETE, but it is quite possible that the high CO_2, directly or indirectly, caused by the CAMP enhanced the strength of the hydrological cycle during the time of maximum precession variability. This in turn enhanced both seasonal and precessional contrasts, thus further destabilizing ecosystems. Long period orbital cycles do seem to have paced Neogene mammalian evolutionary turnover (van Dam et al., 2006). The mean lifespan of Neogene mammals is about 2.5 m.y. which appears to be regulated by the present 2.4 m.y. eccentricity cycle. The same was probably true for Late Triassic and Early Jurassic tetrapods, but the Newark record does not provide enough vertebrates to see this level of taxonomic change. Correlative vertebrate records from western North America might have high enough sampling density to see these sorts of patterns once the time scale is worked out in sufficient detail via coring and correlation to the Newark (e.g., Olsen et al., 2008; Geissman et al., 2010).

Carbon Sequestration

Burning of fossil fuels is adding CO_2 to the Earth's atmosphere in ever increasing amounts. The increase in CO_2 is analogous ti what seems to have happened around the ETE because of the emplacement of the CAMP. It is therefore profoundly ironic that one mechanism for sequestering anthropogenic CO_2 is to pump CO_2 back into CAMP basalts and diabase where the carbonation reaction results in the production of carbonate minerals such as magnesite[14]. Goldberg et al. (2010) describes possible CAMP reservoirs

[14] One carbonation reaction for basalt is olivine (Mg_2SiO_4) + $2CO_2$ = magnesite ($2MgCO_3$) + quartz (SiO_2). Other reactions occur with feldspars.

for CO_2 sequestration in CAMP basalts that can have zones with elevated porosity and permeability in the form of vesicular intervals at the top of flows (Stops 1 & 4) or fracture zones such as in the lower Preakness Basalt (stop 4).Measurements on samples from the Martinsville core of the Orange Mountain Basalt (Figure 17) have porosities of ~10% (Goldberg et al. 2010). In these cases rather than a threat for instability, time is an ally in extending the reaction time with the basalt that brings longer stability. Other possible targets for CO_2 sequestration in the Newark basin are sandstones, and that possibility is also being investigated (e.g., http://www.tricarb.org/tricarb/default.aspx).

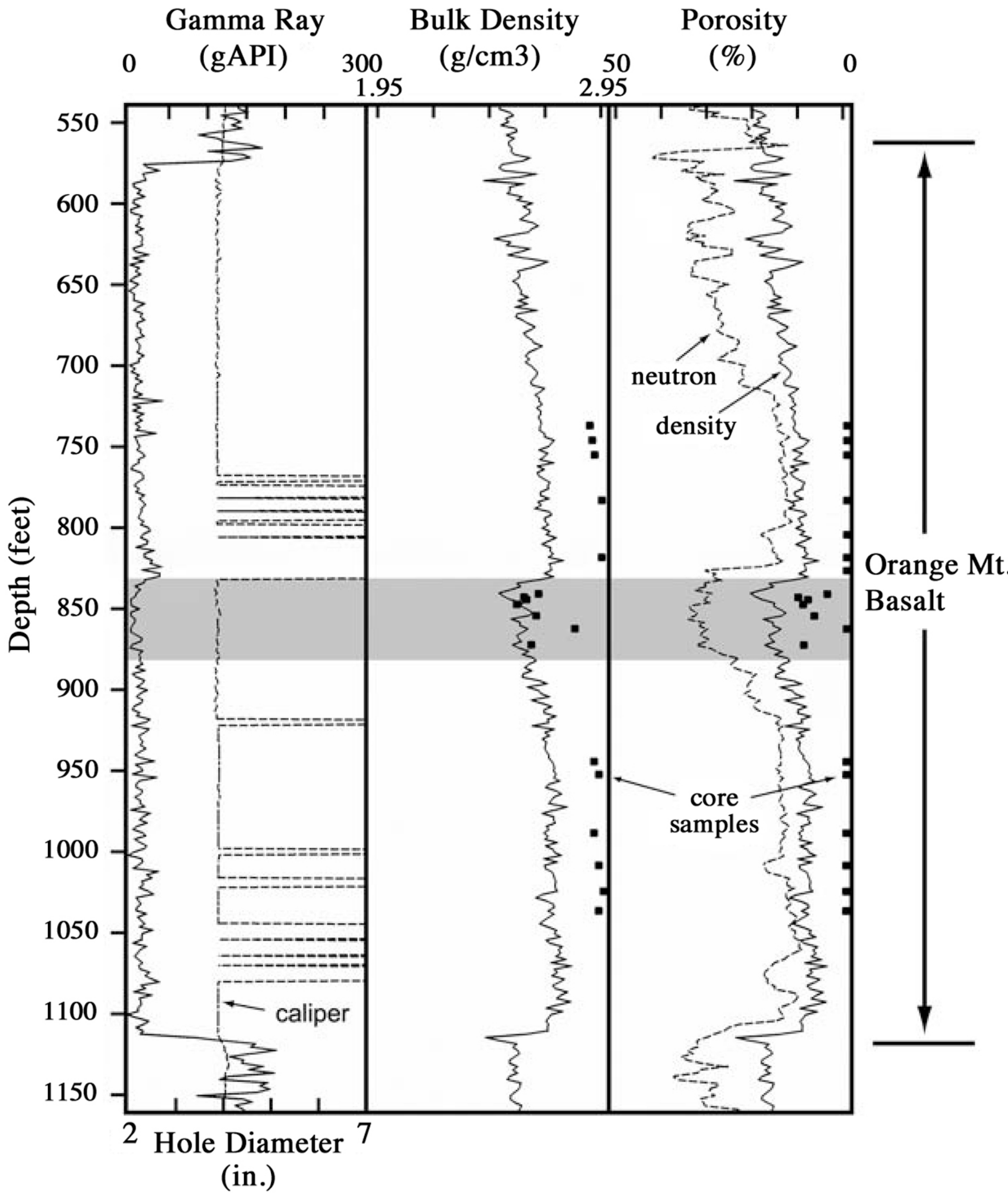


Figure 17: Geophysical log profiles through Orange Mountain basalt. Modified from Goldberg et al. (2009).

FIELD TRIP AND ROAD LOG

The field trip begins and ends at Lamont-Doherty Earth Observatory (LDEO), Palisades, NY in the parking lot closest to 9W. (http://www.ldeo.columbia.edu/about-ldeo/maps-contact). The route (Figure 1) takes us through the Palisades sill, Locakatong and Passaic formations and then into the CAMP lavas and finally back through the Passaic Formation and Palisade sill.

0.0 mi Walk from parking area to the south side of the Machine Shop of LDEO.

Stop 1: Lamont-Doherty Earth Observatory (Machine Shop Area): Newark Basin Cores, Palisade Sill Outcrop, and Borehole Through Sill.

Latitude and Longitude: 41° 0.239'N, 73° 54.760'W.
Stratigraphic Units: Stockton – Boonton Formations and Palisade Sill
Age: ?Late Carnian (early Late Triassic) to Early Jurassic (~230-200 Ma)
Main Points:

1. Representative cores of the Newark basin
2. Lacustrine cyclicity
3. Correlation between Newark and Hartford basins based on core and outcrop
4. Pompton Tuff and the size of the great lakes
5. Carbon sequestration

Cores: Cores will be on display of all of the formations of the Newark basin (Figure 10) as well as a core of the Shuttle Meadow Formation of the Hartford basin (Figure 18) for comparison with the Feltville Formation. Detailed descriptions and interpretations of the sedimentology of these cores and related outcrops are given by Smoot (2010).

Core 1: Stockton Formation, Prallsville Member (Figures 7, 10).

This segment of core is from the Princeton no. 1 core drilled on Princeton University property near the Forrestal Campus (Kent & Olsen, 1995) (Figure 1). It is a fluvial sandstone with some of the highest directly measured porosity and permeability in the basin (3174.50 ft: 131-118 millidarcys permeability, 14.2-14.0% porosity). The Prallsville Member is below the postulated tectonosequence boundary in the Stockton (Figure 7) and is thus of Late Carnian age at about 231 Ma on the NBAGPTS.

Core 2: Lockatong Formation, Tohickon Member (Figures 7, 10).

This segment, from the Nursery no. 1 core (Figure 1) taken off Nursery Road in Ewing Township, is representative of the middle Lockatong Formation and shows nearly one complete Van Houten Cycle. The age of this unit is Early Norian about 219.5 Ma on

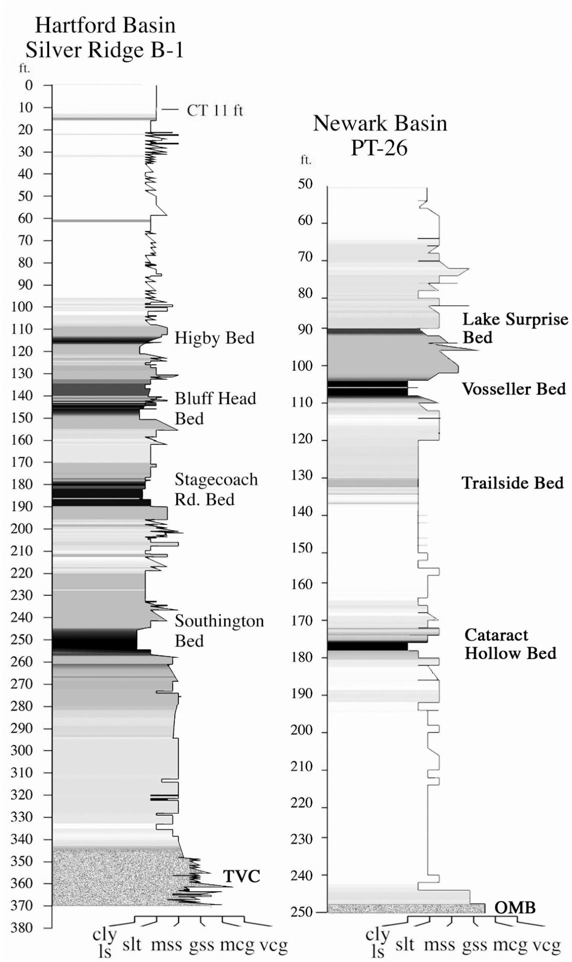


Figure 18: Comparison between the lower Shuttle Meadow Formation of the Hartford basin and the lower Feltville Formation of the Newrak Basin (from Olen et al., 2005).

the NBAGPTS. This represents one of the cycles deposited during times of high precessional variability, but does not represent the deepest of the Lockatong lakes and has not produced any fossil fish in outcrop. This core has well-developed pinch-and-swell lamination and some bioturbation in the deeper water intervals, polygonal desiccation cracks, and brecciated massive mudstone. When the brecciated massive mudstone is very well-developed it is difficult to see anything at all in the core. According to Smoot (2010) the brecciated massive mudstone formed by repeated wetting and drying of mud initially deposited in standing water in a playa with cracks forming over previously filled cracks.

Core 3: Passaic Formation, Cedar Grove Member (Figure 7).

This long segment is from the Weston Canal no. 1 core near Zarephath, NJ. shows much of a 100 ky cycle at around 207.4 Ma. The Cedar Grove Member is one of the most easily mapped units being identified from Newark New Jersey along Route 287 to near Birdsboro, PA. This is a nice example of penecontemporaneous gypsum including abundant nodules and anhedral crystals. According to Smoot (2010), the crystal styles and upward-coarsening successions resemble modern gypsum soil occurrences (Smoot and Lowenstein, 1991) where gypsum dust is blown onto the surface, and then redistributed downward by rainfall rather than gypsum crystallizing out of brines in playas. There are also veins of various orientations filled with fibrous gypsum that formed much later. The muds themselves however formed in very shallow lakes, with the gypsum dust possibly being blown in from very far away (as suggested by Smoot, 2010) along with a considerable amount of lithic dust. The lower part of the member has a Van Houten cycle with one of the best-laminated deep-water intervals in the Passaic Formation. The very well-developed nature of this cycle contrasts with the poorly developed succeeding cycles. In outcrop this well-developed cycle has produced palynomorphs and reptile footprints of characteristic Triassic aspect including the silesaurid dinosauromorph track *Atreipus milfordensis*, the crurotarsian *Brachychirotherium* cf. *parvum*, and the lepidosauromorph *Rhynchosauroides* sp.

Cores 4 & 5: Passaic Formation, Exeter Member and Orange Mountain Basalt (Figure 7, 10).

Two cores are shown here from the Passaic Formation – Orange Mountain Basalt contact. Core 4 is a segment of the Martinsville no. 1 core of the NBCP from Martinsville, NJ from a more central basin location and Core 5 is part of the ACE series, PT-38, spanning the same interval but closer to the northern edge of the basin. Outcrops of this interval in the Jacksonwald syncline have darker shales and much more obvious cyclicity.

In addition to the facies change seen across these three sites, from better cyclicity with dark gray shales in the deeper water units in the southwest, to obscure cyclcity in red beds in the northeast, there is also a parallel change from more or less consistent vertical patterns of cycles to a vertical pattern with fully fluvial conglomeritic facies below to a more marginal lacustrine facies above at the site of PT-38. The Jacksonwald area has been a major source of evidence on the initial ETE, including abundant Triassic-aspect crurotarsian tracks such as *Brachychirotherium* and *Apatopus*, while outcrops adjacent to the location of PT-38 have produced a post-initial-ETE assemblage with abundant theropod dinosaur tracks including *Eubrontes giganeteus* as well as the crocodylomorph track *Batrachopus* and a new species of *Rhynchosauroides*. Also present is a *Brachyphyllum*-*Clathropterus* macrofossil assemblage with a *Classopollis*-dominated sporomorph assemblage.

Cores 6, 7, 8: Feltville Formation, Washington Valley Member and Orange Mountain Basalt (Figure 10) and Shuttle Meadow Formation (Figure 18).

Van Houten cycles in the lower Feltville Formation (Washington Valley Member of Olsen 1980a, 1980c) are unusual in the Newark basin in having deeper water intervals that are very carbonate rich, often limestones. These cores display the lower Feltville Formation in the Martinsville no. 1 core of the NBCP (Core 6) and PT-14 of the ACE cores (Core 7) , and the correlative part of the Silver Ridge B-1 core of the Shuttle Meadow Formation of the Hartford basin (Core 8). These cores were essential to understanding the cyclostratigraphy of the Feltville and Shuttle Meadow Formation and hence to the chronology of the continental Triassic-Jurassic boundary. In fact, it is doubtful that the basic stratigraphy could have been worked out in outcrop without the model and testable hypotheses provided by both the ACE and Silver Ridge cores.

Correlation between cores and outcrops of the lower Feltville Formation is now relatively straightforward. But that was not the case in 1980 when there were only outcrops to work with. Having not seen more than one lake level cycle in superposition I had a stratigraphic model in mind in which there was only one lake level cycle in the Feltville Formation. I interpreted various outcrops along strike accordingly and assumed the dramatic differences between outcrops was due to lateral change (Olsen et al. 1980). I had the same model in mind for the lower Shuttle Meadow Formation. However, when the ACE cores of the Feltville were examined in 1985 it was obvious that there were several Van Houten cycles in the lower Feltville (Olsen et al., 1988; Fedosh & Smoot, 1988). Instead of one carbonate-rich lake level cycle there were two. I revisited the type section of the Feltville (our Stop 3) and was able to test the outcrop generality of this pattern by digging that there were indeed 2 carbonate bearing cycles and then in late August, 2010, I did the same at our Stop 2. This has allowed unambiguous correlation of the cycles in the lower Feltville laterally. One pattern emerging from this correlation (Figure 18, 19) is a progressive thinning and onlap of the lower Feltville onto the Orange Mountain Basalt along strike from the northeast to the southwest, prima facie evidence of syndepositional folding in the Watchung syncline.

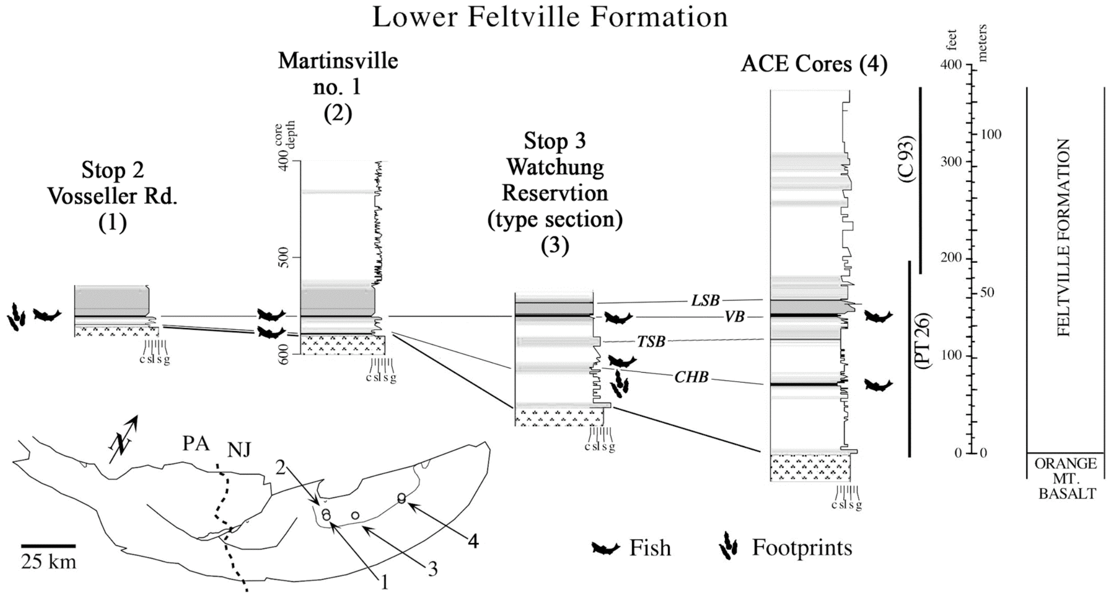


Figure 19: Lateral correlation of basal Feltville showing progressive onlap and thinning. Abbreviations are: *LSB*, Lake Surprise Bed; *VB*, Vosseller Bed; *TSB*, Trailside Bed; *CHB*, Cataract Hollow Bed.

Comparison of the stratigraphy of the lower Feltville and Shuttle Meadow Formation reveals another pattern, and that is that the cyclostratigraphy is more subtle and complicated than expected. Correlation between the Shuttle Meadow Formation and the Feltville is surprisingly obvious (Figure 18), but it makes clear that rather subtle lithological features seen in outcrop and core of the Feltville have important cyclostratigraphic implications. In order to facilitate discussions about Feltville cyclostratigraphy I have given names to the individual deeper water units within the Washington Valley Member (Figures 10, 18). Whiteside et al. (2010) shows that the cyclostratigraphy of the Shuttle Meadow Formation reflects not only the ~20 ky climatic precession cycle but also a ~10 ky hemiprecessional cycle. In specific the Cataract Hollow, Trailside, and Vosseller beds reflect mostly the ~20 ky cycle, while the Lake Surprise Bed reflects largely the ~10 ky cycle. The presence of this hemiprecessional cycle outside the equatorial tropics is not expected of local insolation forcing[15] and

[15] A 10 ky hemiprecessional cycle caused by direct forcing should be limited to ±5° of the equator, while the Shuttle Meadow Formation was deposited at about 21° N.

suggests an expansion or export of equatorial climate variability and an enhancement of the hydrological cycle perhaps due to the super-greenhouse conditions of the syn-CAMP world.

Cores 9 & 10: Preakness Basalt: Flow number 2.

We will see the stratigraphy of the lower Preakness Basalt at Stop 4, but what we will not see there are the gabbroid layers that characterize parts of flow 2. These are important because in some areas they contain zircons that crystallized from the melt and provide high-resolution dates for the flow (e.g. Blackburn et al., 2009). Several segments of ACE cores PT-20 (Core 9) and C-97 (Core 10) will be on display showing these gabbroid layers. These particular layers did produce zircons but these were all apparently assimilated from crustal rocks through which the feeders flowed (S. Bowing, pers. com.) as they have ages 100s of millions of years older than the crystallization age.

Cores 11 & 12: Towaco Formation and Pompton Tuff.

The Towaco Formation is especially well represented in the ACE cores with many parts of the formation being represented by multiple cores. At Stop 5 we will see the 3 prominent Van Houten cycles formed during times of maximum precessional variability in the middle Towaco Formation. The bed names for each of the prominent deep-water units for these cycles are given in Figure 9. Portions of cores C-128 and PT-14 (cores 10 & 11, respectively) that contain the Colfax Road Bed are on display. This bed contains the Pompton Tuff, an airfall ash we will see in outcrop at Stop 5, but here we see it in two cores. The Pompton Tuff also extends to the Hartford basin where it is known from 4 localities a slab from one of these will be on dispaly as well.

Philpotts (in Olsen et al. 2005 and paraphrased here) described this unit in the Hartford basins as thin bed that tends to erode proud of the surrounding deep-water lacustrine unit. The only detectable particles in the layer are all euhedral plagioclase laths with a high aspect ratio that show no signs of rounding. They must have originally been enclosed in glass that is now converted to a clay or to what appears to be chalcedony. There are small circular areas that contain a mottled gray birefringent material that looks very much like chert. These observations are entirely consistent with the small laths of plagioclase being part of a basaltic crystal tuff. There are also small rounded particles that seem to be bits of basalt. The tuff would seem to represent an explosive eruption of the CAMP that is separate from those that produced the stereotypical sequence of known CAMP flows.

What you can see in these cores, the thin sections on display, and slabbed portion of the Westfield bed of the East Berlin Formation of the Hartford basin is a very close match not only between the tuff, but also between the surrounding laminae. As discussed above the similarity is so strong that it seems plausible the lakes in the two basins were connected.

Core 13: Hook Mountain Basalt.

On display is an ACE core C-104 in the Hook Mountain Basalt which exhibits thin gabbroid layers. These too produced only zircons inherited from crustal sources.

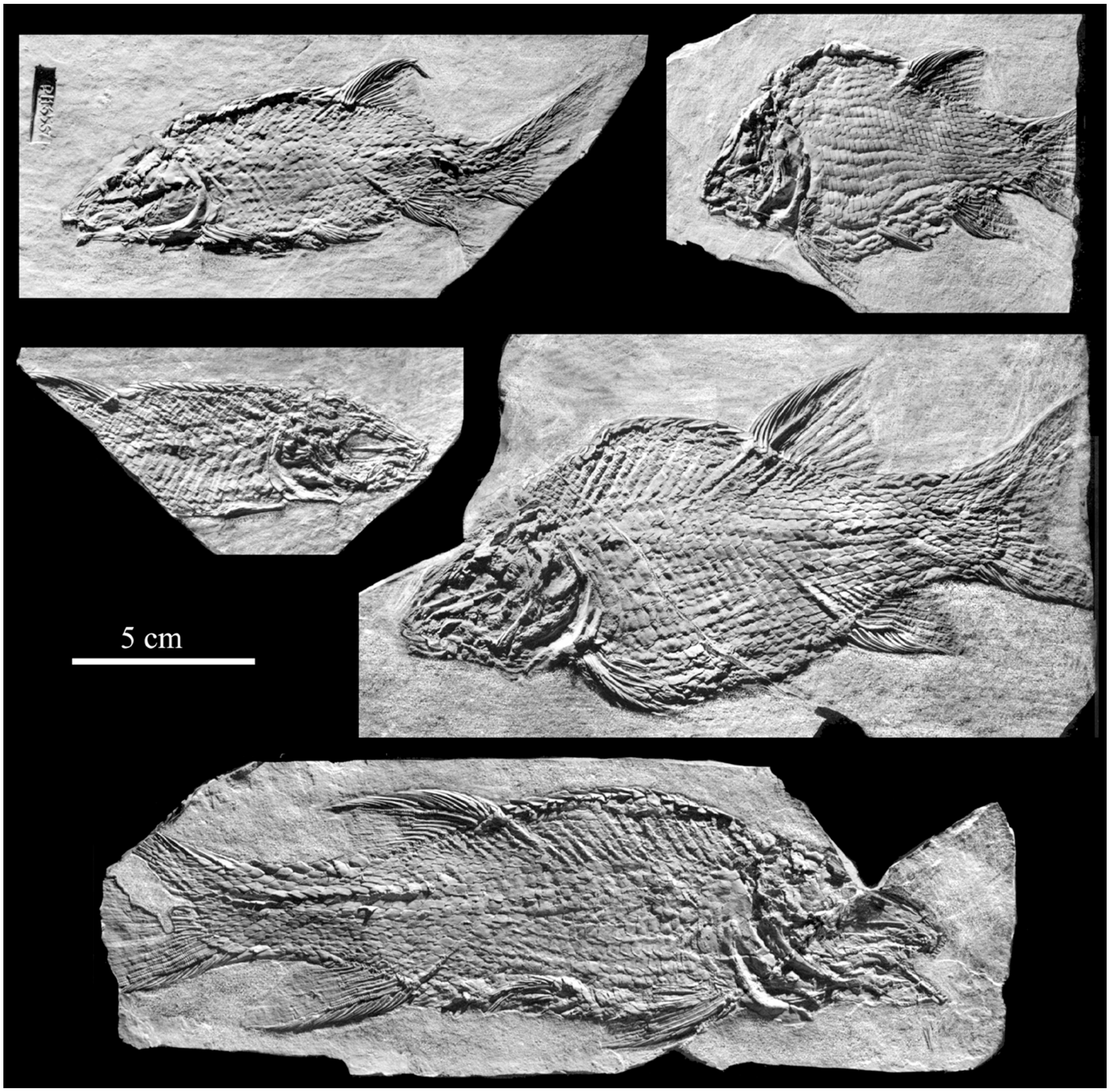


Figure 20: Representatives of the Boonton species flock of *Semionotus*.

Core 14: Boonton Formation, Rockaway River Bed (Boonton fish bed)

The Boonton Formation as sampled in the ACE cores closely resembles the Towaco Formation in its pattern of cyclicity and facies. ACE core PT-6 on display includes the famous bed that produced abundant fish described from the Rockaway River in Boonton, New Jersey, named here the Rockaway River Bed (Figure 10). The semionotids from this bed comprise a distinct species (20).

Lamont Carbonation Experiment Borehole

A borehole drilled at LDEO has been used for experiments in carbon sequestration carbonation using *in situ* carbonation in mafic igneous rocks, in this case diabase (Matter et al. 2007). Palisade diabase is visible behind (west) of the borehole site. The small small-scale CO_2 injection experiment at this site showed large decreases in Mg^{++} and Ca^{++} concentrations over short periods of time (200 hours) suggested both mixing between the injected solution and aquifer water and the release of cations from water–rock dissolution, that neutralized the introduced carbonic acid (Goldberg et al. (2010). This the same kind of reaction that would result in the weathering of CAMP basalt flows and the accumulation of carbonates in the early CAMP lakes of the lower Feltville Formation, which now has the potential to sequester the carbon from the burning of fossil fuels. As Goldberg et al. (2010) point out, if large CAMP reservoirs are found in proximity to major industrial carbon sources huge volumes of CO_2 could be sequestered permanently by the *in situ* carbonation reaction.

Return to vehicles and leave LDEO via the exit to 9W.

0.0 mi. Leave entrance to LDEO, turn left and proceed south on Rt. 9W along the strike of the Palisade sill. There fill be small exposures of the sill along Rt. 9W towards Fort Lee.

10.4 mi. Take the ramp on right onto US-9W S.

10.9 mi. Continue onto Fletcher Ave. Exposures on north side of ramp for NJ-4 of coarse gabbroid of Palisade sill. This gabbroid produces zircons yeilging a very precise age (Blackburn et al., 2009).

11.0 mi. Turn right to merge onto US-1 S/US-46 W/US-9 S toward I-95 S/I-80 W Continue to follow US-46 W. Driving down the dip slope of Palisade sill. Exposures to north along US-80/I-95 reveal contact of the sill and overlying Lockatong Formation.

13.9 mi. Take the I-95 S/I-80 W/N J Turnpike ramp.

14.0 mi. Continue toward I-95 S.

15.0 mi. Take exit 16W toward Rutherford.

15.6 mi. Merge onto I-95 S.

25.3 mi. Take exit 14 to merge onto I-78 W toward US-1/US-9/US-22/Newark Airport.

35.3 mi. Exposures of Orange Mountain Basalt.

37.0 mi Approximate contact between Feltville Formation and overlying Preakness Basalt. Notable is the high degree of distinctive splintery fracturing that characterizes much of the second Preakness flow, which we will see up close at Stop 4.

44.4 mi. Take exit 40 toward The Plainfields/Gillette/Watchung.

44.6 mi. Turn left at Hillcrest Rd.

45.2 mi. Passing into Feltville Formation.

45.7 mi. Turn right at County Route 527 S/Valley Rd. Continue to follow County Route 527 S.

45.8 mi. Slight left at County Route 527 S/Stirling Rd.

45.8 mi. Turn left to stay on County Route 527 S/Stirling Rd.

45.8 mi. Slight right at County Route 527 S/Mountain Blvd. Continue to follow County Route 527 S. Route follows strike of Feltville Formation, mostly on or near the contact with the Orange Mountain Basalt.

49.6 mi. Continue onto Washington Valley Rd. Route now follows along middle Feltville Formation.

51.6 mi. Turn left at Vosseller Ave. Headind down section in Feltville.

52.0 mi. On right is unpaved access to Eastfields Park of Somerset County Park Commission. Entrance is on upper Orange Mountain Basalt.

Park.

Stop 2: Lower Feltville Formation (Washington Valley Member).

Latitude and Longitude: 40.5932°N, 74.5452° W.
Stratigraphic Units: Lower Feltville Formation, Washington Valley Member, Orange Mountain Basalt
Age: Latest Triassic (Latest Rhaetian) to ?earliest Jurassic (201.3 Ma)
Main Points:

1. Condensed section of Washington Valley Member
2. Best exposure of Vosseller Bed
3. Very condensed section of Cataract Bed

4. Onlap onto Orange Mountain Basalt
5. First 100 ky after initiation of ETE
6. Limestones as products of super-greenhouse
7. Intense disruption of Vosseller Bed and "dead horses"
8. Species flocks of semionotid fishes
9. Dinosaur tracks

Walk about 250 ft north to outcrops along Blue Brook and walk downstream past the outcrops along the north bank of the stream about 800 ft west to the contact with the Orange Mountain Basalt.

About 12 m of Feltville section outcrops here. The complete Van Houten cycle containing the Cataract Hollow Bed outcrops along with nearly the complete cycle containing the Vosseller Bed, this being the best outcrop of the latter. The section is closely comparable to what is seen in the Martinsville no. 1 core, except it is even more condensed. The section begins with the vesicular upper surface of the Orange Mountain Basalt overlain by a decimeter of basalt gravel and cobbles in a bluish-gray mudstone. This is followed by a decimeter of gray/white, seemingly? oscillatory rippled sandy calcarenites passing up into two meters of red mudstone. This bed is clearly the Cataract Hollow Bed. In my section of this interval in 1980 I completely missed this bed., because I did not have an expectation of its existence. In August 2010, I looked for the Cataract Hollow Bed in the completely submerged interval near the Orange Mountain Basalt. This illustrates the need to have some kind of rigorous testable hypothesis of what should be present. Without that, why would I explore every hidden nook and underwater cranny.

The red beds above the Cataract Hollow bed is overlain by 20 cm of blue gray mudstone with fabrics like the underlying red mudstone. This is abruptly overlain by ripple crosslaminated gray fine sandstone with sole casts of dinosaur tracks. These tracks are so far all brontozooids (grallatorids).

Above the ripple crosslaminated sandstone is a grey mudstone with floating blocks to flakes of black laminated to microlaminated limestone and claystone. Presumably, these blocks once comprised continuous layers. They are in fact comparable to what I have termed "dead horses" (Olsen et al., 1989), although these are much larger clasts ("horses"). These blocks generally exhibit signs of bedding plane-parallel shear, including folds and imbricate thrust faults. The name derives from the term "horse" for a fault-bounded sliver of rock and their "dead" or dismembered condition and the name was coined to call attention to the phenomenon. The gray mudstone shows few or no depositional sedimentary structures. There is no well-defined upper surface of the bed. In this and other deposits such as the East Berlin Formation (Olsen et al., 2005), or Towaco Formation (Stop 1), the associated massive mudstone suggests that partial liquefaction accompanied the deformation that produced the dead horses, and I interpret them as a form of low-pressure structural mélange. These features are common in fine-grained organic rich rocks but they have usually been interpreted as depositional units, such as muddy turbidites (e.g., Dyni & Hawkins, 1981). Dead horses are still a feature of the Vosseller Bed at Stop 3, but seem absent in the ACE cores.

The limestone blocks themselves can be more than a meter in diameter and 30 cm thick, especially those with folds. Fractures in the limestone contain a tarry bitumen that is liquid, if highly viscous, at body temperature but generally solid in the water. However,

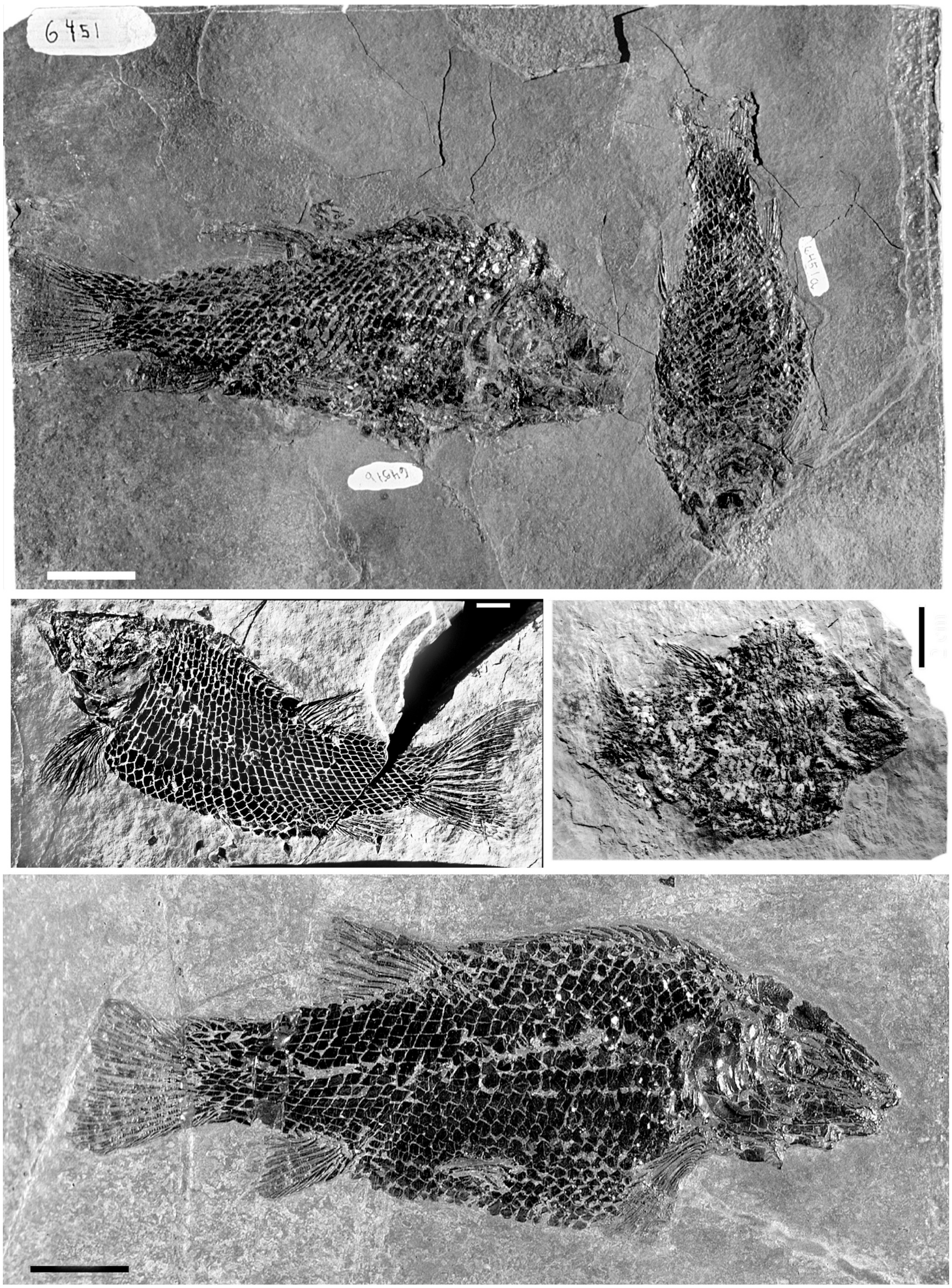


Figure 21: Examples of the Feltville, Vosseller Bed, semionotids from Stop 2.

the oil slicks that appear on the streams surface when blocks are pried loose from the matrix attest to the presence of more fluid hydrocarbons. The thermal maturity of these beds are within the oil window based on a vitrinite reflectance value of 1.18 R_0% from the Vosseller bed of the nearby Martinsville no. 1 core (Malinconico, 2009), and the hydrocarbons present were surely generated from the limestones.

These blocks of limestone themselves contain mostly articulated, and some disarticulated fish, including *Semionotus*, *Redfieldius*, and *Ptycholepis* (Figure 21) The presence of coprolites suggests the presence of coelacanths yet to be found. The semionotids comprise a species flock with most variation being in overall body shape and the morphology of the series of spined scales, called dorsal ridge scales along the dorsal midline (Figure 14). In fact the name *Semionotus* is derived from "signal back". Preservation is often exquisite and with chemical preparation[16] extremely fine details can be seen. McCune recorded six species in the Feltville (McCune et al., 1984; Olsen et al., 1989).

The origin of the carbonate so prevalent in the Vosseller, Cataract Hollow, and Feltville in general is worth commenting on because such carbonate rich beds are virtually absent from the rest of the Newark basin. Carbonate and limestones are not at all uncommon in lakes, and such famous units as the Green River oil "shales" are in fact carbonates. The carbonate was mostly derived from biologically mediated chemical precipitation in the upper part of the water column, rather than from the tests of organisms. As discussed above, the unusual amount of carbonate in the Feltville and in fact in all of the post-initial CAMP basalt flows in central Pangea may relate to weathering of the CAMP basalts themselves under super-greenhouse conditions in the first 100 ky after initiation of ETE.

Proceed back to vehicles.

52.0 mi. Turn left (northeast) onto Vosseller Avenue toward Perrine Rd.

52.5 mi. Turn right at Washington Valley Rd.

55.2 mi. Continue onto County Route 527 N/Mountain Blvd Ext. Continue to follow County Route 527 N.

58.3 mi. Turn left at County Route 527 N/Stirling Rd. Continue to follow County Route 527 N.

58.4 mi. Turn left at Hillcrest Rd.

59.2 mi. Merge onto I-78 E via the ramp towards Newark.

62.9 mi. Take exit 44 toward New Providence/Berkeley Heights.

---

[16] In chemical preparation, the side of the fish exposed upon splitting the rock is embedded in a resin such as epoxy or polyester and the maturix is subsequently removed by dissolution in acetic acid, leaving the bone adhering to the plastic.

63.1 mi. Turn left at County Route 527 N/Glenside Ave.

63.5 mi. Take the 1st right onto Cataract Hollow Rd.

63.6 mi. Turn right into parking area for Watchung Reservation, Union County Parks an Community Renewal.

Park

Stop 3: Lower Feltville Formation (Washington Valley Member) Type section.

Latitude and Longitude: 40° 41.009'N, 74° 23.292'W.
Stratigraphic Units: Lower Feltville Formation, Washington Valley Member, Orange Mountain Basalt
Age: Latest Triassic (Latest Rhaetian) to ?earliest Jurassic (201.3 Ma)
Main Points:

1. More expanded section of Washington Valley Member
2. Best exposure of Cataract Bed and superposition of Vosseller bed.
3. Fissures in Orange Mountain Basalt
4. Copper at contact with Orange Mountain Basalt
5. First 100 ky after initiation of ETE
6. Limestones as products of super-greenhouse
7. In situ *Clathropteris*
9. Dinosaur tracks

Walk southeast and then southwest to the beginning of the loop near the end of Cataract Hollow Road about 2090 ft from the parking area. At the trail head, follow trail to southeast 330 ft and then about 600 ft to the northeast along the raised path (old canal levy) to Station 1 from where we will proceed to 3 additional stations to cover the stratigraphy at the type section of the Feltville Formation (Figure 19). Feltville is the name of the abandoned village of Feltville along Cataract Hollow Road, now part of the Watchung Reservation of the Union County Department of Parks and Community Renewal.

<u>Station 1</u> at about 40˚ 40.747'N, 74˚ 22.718'W is an outcrop of pink, gray, and white limestone interbedded with red siltstones and fine sandstones. This is the best vertical outcrop of the Cataract Hollow Bed and its type section. Fragmentary to articulated fish and occasional coprolites occur in the lower laminated part of the limestone. The fish bones and scales are pink and dark brown. The upper carbonate beds at this station are calcarenites that grade into the overlying red beds. The laminated and laterally continuous nature of the base the Cataract Hollow Bed as well at the articulated fish shows that it was deposited in quiet water, below wave base, generally lacking oxygen in contrast to the bed at Stop 1.

To the southeast, on the opposite side of Blue Brook is a ravine that cuts down down the dip-slope of the Orange Mountain Basalt. Thin to thick-bedded red siltstones of the basal Feltville Formation fill what is clearly a fissure in the basalt. Such fissures are common in the Orange Mountain Basalt and can cut down through much of the Formation, as seen at various quarries. In Nov Scotia, such fissures contain tetrapod bones, and it quite likely they do here as well, although this site has not been sufficiently prospected.

Proceed 500 ft northeast to bluff on west side of southeast flowing tributary (out of Cataract Hollow) to east-facing bluff that is Station 2.

<u>Station 2</u> at approximately 40˚ 40.822'N, 74˚ 22.718'W is an outcrop of poorly exposed black laminated carbonate and mudstone of the Vosseller Bed and surrounding gray clastics. Careful examination of the geometric relationship between this outcrop and Station 1 shows that this bed in superposition to the Cataract Hollow Bed, but this can be seen unambiguously at Station 4. Proceed southeast along the trail 270 ft crossing the small south-flowing tributary at Cataract Hollow and proceed about 620 ft to trail heading east up the slope (noting limestone of the Cataract Hollow Bed in the path). Continue 250 ft east to Station 3.

<u>Station 3</u> is at about 40˚ 40.860'N, 74˚ 22.718'W and consists of an outcrop extending from this ridge to the stream below and to the south. The contact between the Orange Mountain Basalt and the overlying Feltville is nearly at stream level and the overlying beds consist a nodular limestone succeeded by the transgressive portion of a Van Houten cycle bearing the Cataract Hollow Bed, the Cataract Hollow Bed itself and the overlying regressive red beds of the cycle, which we saw better exposed at Station 1. The larger blocks of the nodular limestone are very fine-grained and contain darwinulid ostracodes. The gray sandstones and siltstones of the transgressive portions of the cycle conatin plants in groth position including the horsetail *Equisetites* and the fern *Clathropteris* as well as dinosaur footprints. The Cataract Hollow Bed is deeply weathered here and invaded by modern tree roots, and the overlying beds are highly weathered. Just upstream from this station is the old copper mine described by Manspeizer (1980) at this site. I do not think there is evidence of intrusion here and the thermal maturity of the organic material here is not elevated. Proceed north and down the slope about 150 ft to path and walk about 960 ft northeast to Station 4 on the bluff on the north side of the west flowing tributary.

<u>Station 4</u> is at about 40˚ 41.015'N, 74˚ 22.718'W and has good outcrops of the Lake Surprise Bed, Vosseller Bed, through the Trailside Bed and then poor outcrops down through the Cataract Hollow Bed. It was at this site that I was able to confirm the superposition of the Vosseller and Cataract Hollow beds by digging for the latter after seeing the ACE cores. The Trailside Bed consists of purple and gray siltstomnes containg abundant *Brachyphyllum*. The cyclostratigraphic equivalent Stagecoach Road bed of the Shuttle Meadow Formation is a black calcareous mudstone with whole fossil fish and conhcostracans. The Lake Surprise Bed is a finely laminated black shale bed here with partly articulated *Semionotus*. The Vosseler bed is unfortunately very weathered and crumbly, although I have found articulated fish in it here.

Return to vehicles.

63.6 mi. Turn left from parking are onto Cataract Hollow Rd.

63.7 mi. Turn left at County Route 527 S/Glenside Ave.

64.0 mi. Turn right to merge onto I-78 E.

72.1 mi. Take exit 52 to merge onto Garden State Parkway N.

91.1 mi. Take exit 160 toward Fair Lawn/Hackensack.

91.3 mi. Slight right at Garden State Plaza Parkway.

91.3 mi. Take the 1st left onto Paramus Rd/W Passaic St. Continue to follow Paramus Rd.

91.8 mi. Merge onto NJ-4 W via the ramp to NJ-208 N/Fair Lawn/Hawthorne.

92.1 mi. Continue onto NJ-208 N.

97.5 mi. Take the Grandview Ave exit toward Wyckoff.

97.6 mi. Turn right at Grandview Ave.

97.9 mi. Turn right at Goffle Hill Rd.

98.5 mi. Continue onto Sicomac Ave.

98.8 mi. Turn left at Mountain Ave.

99.7 mi. Continue onto Sicomac Rd.

100.2 mi. Turn left at High Mountain Rd.

100.3 mi. Slight right at Belmont Ave/Passaic County 675.

101.7 mi. Turn right at W Overlook Ave.

101.8 mi. Turn left into parking area for strip mall.

Park

Stop 4: Upper Feltville Formation and Preakness Basalt, William Paterson University.

Latitude and Longitude: 40° 57.083'N, 74° 11.524'W.
Stratigraphic Units: Upper Feltville Formation, Preakness Basalt (flows 1 and 2)
Age: Earliest Jurassic (201.1 Ma)
Main Points:

1. Upper sandy Feltville Formation
2. Contact between Feltville and Preakness Basalt
3. Basal pillowed flow of Preakness Basalt (P-1)
4. Flow P-2 of Preakness basalt with low paleomagnetic inclinations
5. Largest single flow in world?
6. CO_2 sequestering possibilities in basalt

Walk west from parking area about 290 ft to entrance to abandoned quarry on south side of West Overlook Ave head towards stream walking about 300 ft to open area of the old quarry.

Manspeizer (1980) described this locality in that year's NYGSA guidebook noting the three basic units present. The upper Feltville Formation, a lower pillowed flow of the Preakness Basalt and a second massive and highly fractured flow of the Preakness Basalt. The Feltville Formation exposed in the old quarry and adjacent stream consists of interbedded tan and red sandstones and red and gray and purple siltstones. The latter contain sporomorphs and conifer fragments. In terms of the cyclostratigraphy of the Feltville this sequence is the interval of maximum precessional variability in the 100 ky cycle exhibiting the lowest precessional variability of its 405 ky cycle. The contact with the Preakness has a cyclostratigraphic age of about 201.1 Ma.

The basal flow of the Preakness as exposed here consists of a complex of pillowed basalts and the more massive flow lobes that fed the pillows all of which show considerable vesicularity. This is Preakness flow P-1 of Tollo & Gottfried (1992) and is a distinct and mappable unit extending at least from the ACE core transect to the north to near the border fault. It seems to absent from at least West Orange (I-280) to near Somerville, NJ, but reappears along I-275 at Pluckemin, NJ. I (Olsen, 1980a, 1980d) described outcrops of the lower flow but did not different it from the overlying flow P-2 of Tollo & Gottfried (1992). At this locality there is little or no visible metamorphic effect on the underlying Feltville Formation, which to be expected because P1 was extruded into water, presumably as deep as the thickness of the pillowed flow itself.

Proceed from the quarry up the hill on the north to West Overlook Avenue and then to the west to exposures of basalt. The contact between P-1 and Tollo & Gottfried's (1992) flow P-2 can be seen here. P-2 is the very thick flow that is present all over the entire extent of the Preakness Basalt and is characterized by having an intense splintery or prismatic fracture (Faust, 1975) and gabbroid layers (Puffer & Volkert, 2001). Prévot & McWilliams (1989) noted that this flow has unusually low magnetic inclinations compared to the other Newark basin flows. They also found the same low inclinations in the second flow of the Hartford basin Holyoke Basalt and the Deerfield basin Deerfield Basalt. Hozik (1992) showed that the Sander Basalt shared the low inclinations unlike all the other basalts of the Culpeper basin. All of these low-inclination flows have the same chemistries (Puffer, 1992, 2003) and all have gabbroid segregations and all but the Deerfield Basalt have the characteristic splintery fracture. In the Hartford and Deerfield basins, a flow of similar chemistry but having a tendency to be pillowed is present below the flow with the low inclinations. As pointed out by Prévot & McWilliams (1989) these directions would seem to indicate correlation, and correlation within the time frame of secular variation, suggesting that these flows represent the same eruptive event of no more than 10s or 100s of years. The Sander flow with the low inclinations is over 200 m thick, and P-2 of the Preakness Basalt is more than 90 m thick, and if the flow extended from Massachusetts to Virginia, it would be on of the largest lava flows known on Earth. In terms of environmental effects, it is surely the rate as well as the magnitude that matters, and the eruption of this flow, if it was indeed one eruption, would have had significant environmental effects. The largest single flow of the Columbia River Basalt is on the order of 5000 km^3 (Tolan et al., 1989), but if the Preakness and equivalents averaged 100 m in thickness, spanned 800 km along strike, and were 100 km wide prior to erosion, it would have a volume of 8000 km^3. This does not include the area spanned by dikes of the same composition that extend well into Canada (McHone, 1996).

Flow P-1 is highly vesicular at this locality, and P-2 has very significant fracture porosity and permeability. As previously discussed, Goldberg et al. (2009) argue that carbonation reaction in porous zones of basalt could provide a significant locus for carbon sequestration. The large amount of porosity and permeability is obvious here, but whether that holds at depth has yet to be demonstrated.

Return to vehicles.

101.8 mi. Turn left (west) from parking are onto W Overlook Ave toward Lenox Ave.

102.1 mi. Turn right at Mills Dr.

102.4 mi. Continue onto College Rd.

103.2 mi. Turn right at County Rd 504 W/Hamburg Turnpike. Continue to follow Hamburg Turnpike.

108.1 mi. Turn right at Terhune Dr. (US-202).

108.5 mi. Turn right at Brook Terrace.

108.8mi. Take the 1st right onto Green Knolls Dr/Pines Lake Dr W.

108.8 mi. Green Knolls Dr/Pines Lake Dr W.

Park

Stop 5: Middle Towaco Formation, The Glen, Pompton, NJ.

Latitude and Longitude: 40° 59.467'N, 74° 16.083'W.
Stratigraphic Units: Middle Towaco Formation, Colfax, The Glen, and Pines Lake Beds
Age: Early Jurassic (200.8 Ma)
Main Points:

1. Classic 19th century site never exploited
2. Well developed cyclcity
3. Basin margin facies
4. Lower trophic levels well represented
5. Species flocks of semionotid
6. Tracks
7. Pompton Tuff
8. Size of lakes

The Glen is a community reserve managed by the Pines Lakes Association. Visits to this site should always be coordinated through them.

About 65 m of the middle Towaco Formation extensively outcrops along the creek and its tributaries joining Pines Lakes and Pompton Lake (Figure 22), including the top of one Van Houten cycle, two complete cycles, and the lower half of another. Three deep-water units of the upper three successive Van Houten cycles outcrop in their entirety, with the cycles averaging about 25 m thick. These three beds are termed in succession, the Pines Lake Bed, the Glen Bed, and the Colfax Bed, corresponding to the designations P-5, P-4, and P-3 in Olsen et al. (1989) and papers by McCune, respectively.

The Glen was made known as a fossil locality by W. C. Redfield (1842) who reported on the lower two shales in succession producing fossil fish and intermediate strata producing reptile tracks. It is worth mentioning that William C. Redfield, was the first president of the American Association for the Advancement of Science in 1848 and great grandfather of Alfred C. Redfield of Redfield Ratio fame.

From the parking area, walk 0.2 mi southeast across bridge and turn west into woods to the west at the small tributary to the main stream. Proceed down hill about 135 ft along tributary to Station 1.

<u>Station 1</u> is at approximately 40° 59.467'N, 74° 16.083'W. Exposed in the small run is the lowest deeper water unit, the Pines Lake Bed (P-5). This unit lies on gray sandstones and conglomerates with organically preserved roots, which in turn lie on red clastic rocks. The Pines Lakes Bed is a laminate, but is not microlaminated and has rare whole fish and fish fragments. The most distinctive aspect of the bed is the presence of stromatolites around trees (Figure 23). These have been termed arboreal stromatolites by Whiteside (2004), which were a part of the assemblage of lacustrine primary producers. The Pines Lake Bed is presumably the lower of the two beds described by Redfield (1842), although he gives the distance to the next fish-bearing unit is 60 m (200 ft) while it is certainly closer to 25 m. Proceed downstream about 150 ft south to near where the stream turns west and then northwest to Station 2.

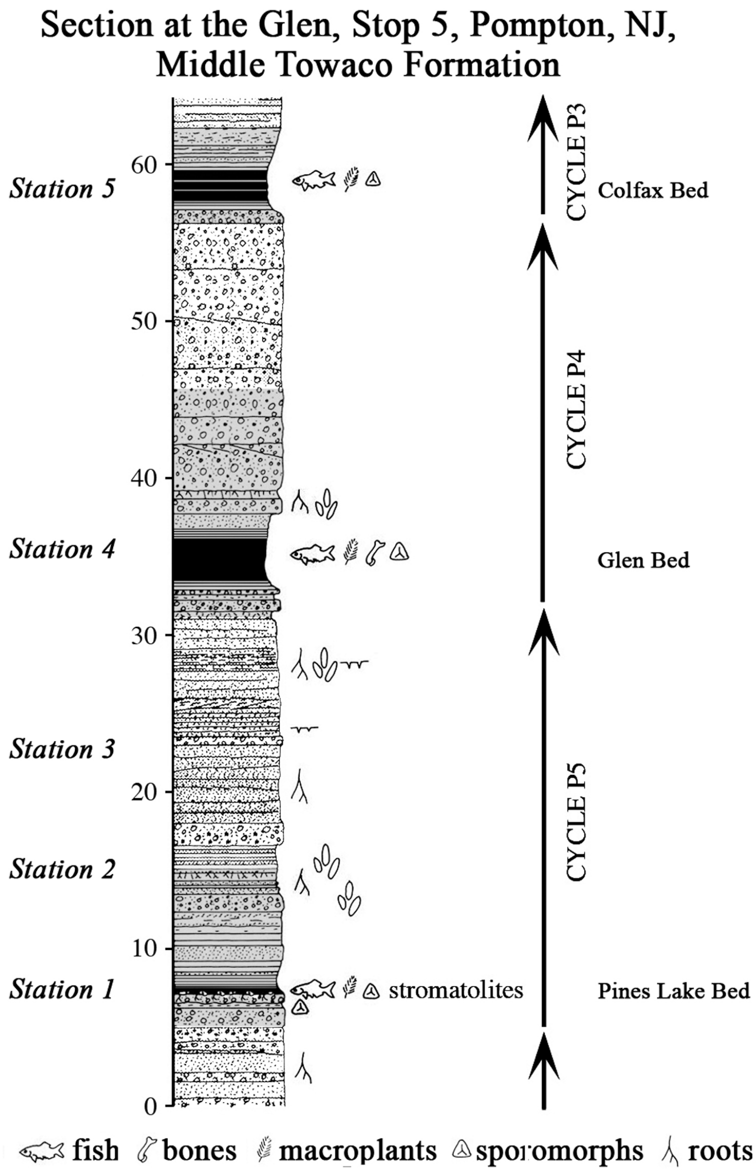


Figure 22: Section at Stop 5. White denotes red beds. Modified from (Olsen et al., 1989).

<u>Station 2</u> is at about 40° 59.442'N, 74° 16.080'W. We have passed up section through gray mudstones, sandstones and minor conglomerate into reddish thin-bedded siltstones and very fine sandstones with abundant dinosaur footprints. Proceed downstream about 200 ft.

<u>Station 3</u> is at 40° 59.454'N, 74° 16.116'W. Upsection from Station 3 we pass into presumably fluvial red sandstones and minor mudstones comprising the shallow water portion of the lower Van Houten cycle. The upper beds of this red sequence becomes finer grained and more highly organized. This red sequence is the interval mentioned by Redfield (1842) as being quarried by Peter M. Ryerson, Esq., of Pompton, NJ. This interval produced several dinosaur tracks, all brontozooids.

Continue downstream about 500 ft.

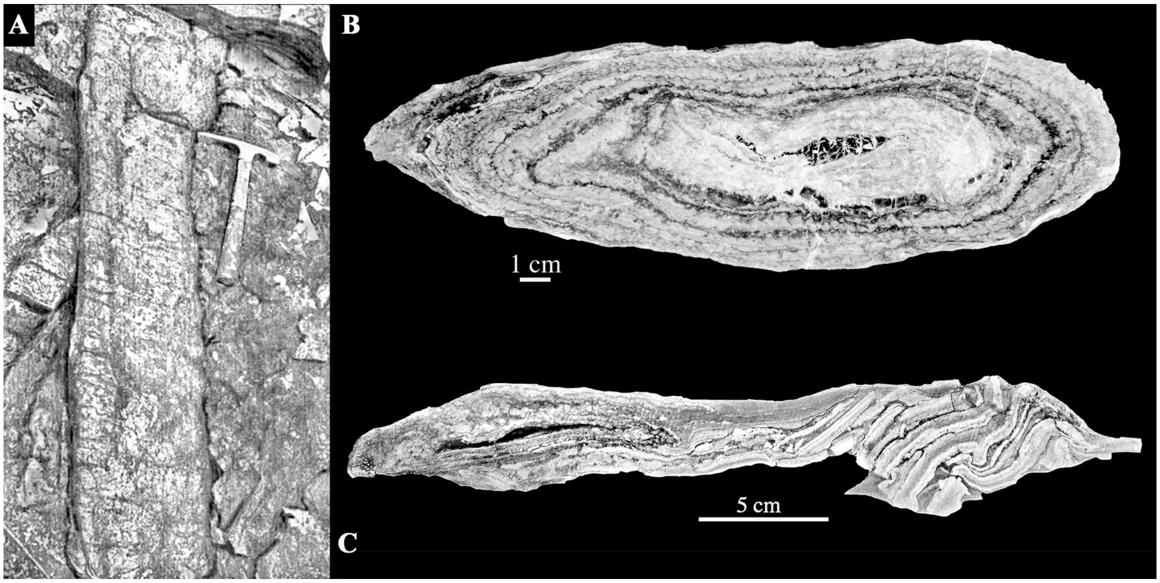
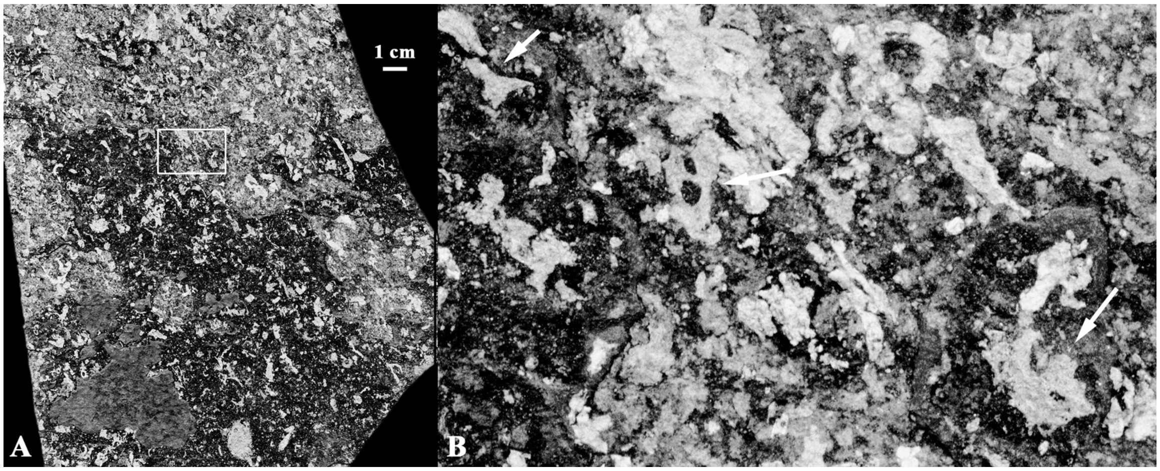


Figure 23: Stromatolites at Station 1: A, *in situ* +1.5 m stromatolite around tree in 1971 at Station 1, Stop 5; B, crossection of same stromatolite with organically preserved tree in center; C, imbricate slumped microbialite mat and associated stromatolite around small tree.

Station 4 is at approximately 40° 59.532'N, 74° 16.164'W at the point that the bluff comes down to near the level of the creek. Here, The Glen Bed is exposed at an accessible level. This is the upper fish-bearing unit mentioned by Redfield (1841). It is also the site of a quarry opened in 1980 by McCune and PEO, and excavated by Amy R. McCune (McCune, 1982, 1986, 1987a, 1986b, 1990, 1996, 1997, 2004).

The lower transgressive portion of this Van Houten cycle (paraphrased from Olsen et al., 1989) is characterized by an overall fining-upward trend from wave-reworked conglomerates into oscillatory rippled sandstones and siltstones. The high-stand sequence begins with oscillatory-rippled siltstones, passes rapidly into laminated siltstones with pinch-and-swell laminae, and then up into a thick microlaminated interval showing silt-carbonate couplets and many sub-millimeter to decimeter thick graded distal lacustrine turbidites. One turbidite has rare 1-2 cm diameter pebbles at its base. The total organic carbon content of these beds is relatively low compared to the finer facies of the Towaco, averaging less than 2% by weight. Complete, well-preserved fossil fishes are abundant in these beds (see below). The transition into overlying non-microlaminatcd silts is abrupt, as is the correlative disappearance of fish.

A peculiar feature of several decimeters of this sequence is the presence of white to tan irregular flattened objects in some of the finest microlaminated beds. Many shapes have cuspate edges or apparent holes, while others are more rod, vermiform, or even hair-like (Figure 24). The simplest explanation of this material is that it represents devitrified and flattened tephra, largely pumice particles that fell into the lake or were washed in from accumulations along shore. The latter interpretation makes more sense because these particles are so abundant on so many laminae. However, the presence of such foamy pyroclastics implies silicic eruptions which have not yet been documented in the CAMP



.

Figure 24: Devitrified and flattened tephra fragments in microlaminated strata at Station 4. A, slab showing blobs, shards, and filaments; B, inset of A showing details with arrows pointing to vesicular blobs.

Sandstone and conglomerate at the transition into the strongly regressive and low-stand portions of the cycle exhibit features suggesting wave-reworking (Smoot, 2010), including virtually all the features described by LeTourneau (1985a,1985b), such as oscillatory ripples, "fitted fabric" of pebbles and cobbles, oriented plant debris. and well-sorted layers and patches of sand, granules, and pebbles. Also present are small lenses of oscillatory-rippled sandstone nestled in well-sorted conglomerate beds. Organically-preserved roots, and dinosaur footprints are present. Higher in the low-stand interval, the conglomerates appear fluvial in origin, although they remain to be examined in detail. These are comprise the thick sequence of conglomerate (Figure 22) referred to by Redfield (1842) as the, “variegated calcareous conglomerate" of Rogers.

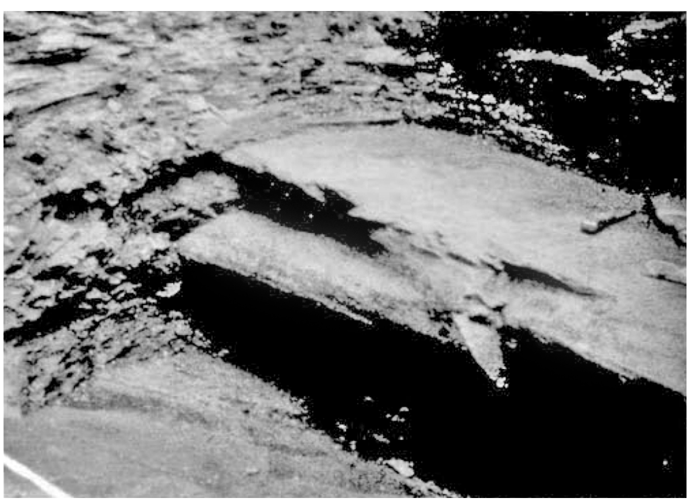


Figure 25: Nappe-like folds associated with bedding plane-subparallel thrust faults in Towaco Formation, The Glen, Stop 5, Station 4. Photo by A.R. McCune from Olsen et al. (1989).

Decimeter-scale folds are prominent in the microlaminated portions of this cycle. They could be subaqueous slump folds, but additional work done by McCune and PEO in the early 1980s, shows they are nappe-like drag folds of thrust faults propagating upward through major portions of the microlaminated unit. These folds show large amounts of thickening in their hinges, which are often cut by small faults. The fault-adjacent limbs are often sheared out, although completely oveturned bedding is locally present. Fish are deformed within these folds, shortened where they are perpendicular to the fold axis and elongated when they are parallel to it (Figure 25). The thrusts themselves are slickensided and sometimes polished, and they sole into bedding plane shear zones. The ductile behavior of the beds and the extremely low thermal maturity of the contained hydrocarbons (Pratt el al., 1988) demands that this thrust faulting and folding be early in the burial history of the units, prior to complete lithification, but after significant burial. An implication of these folds is that the deep-water unit is considerably structurally thickened.

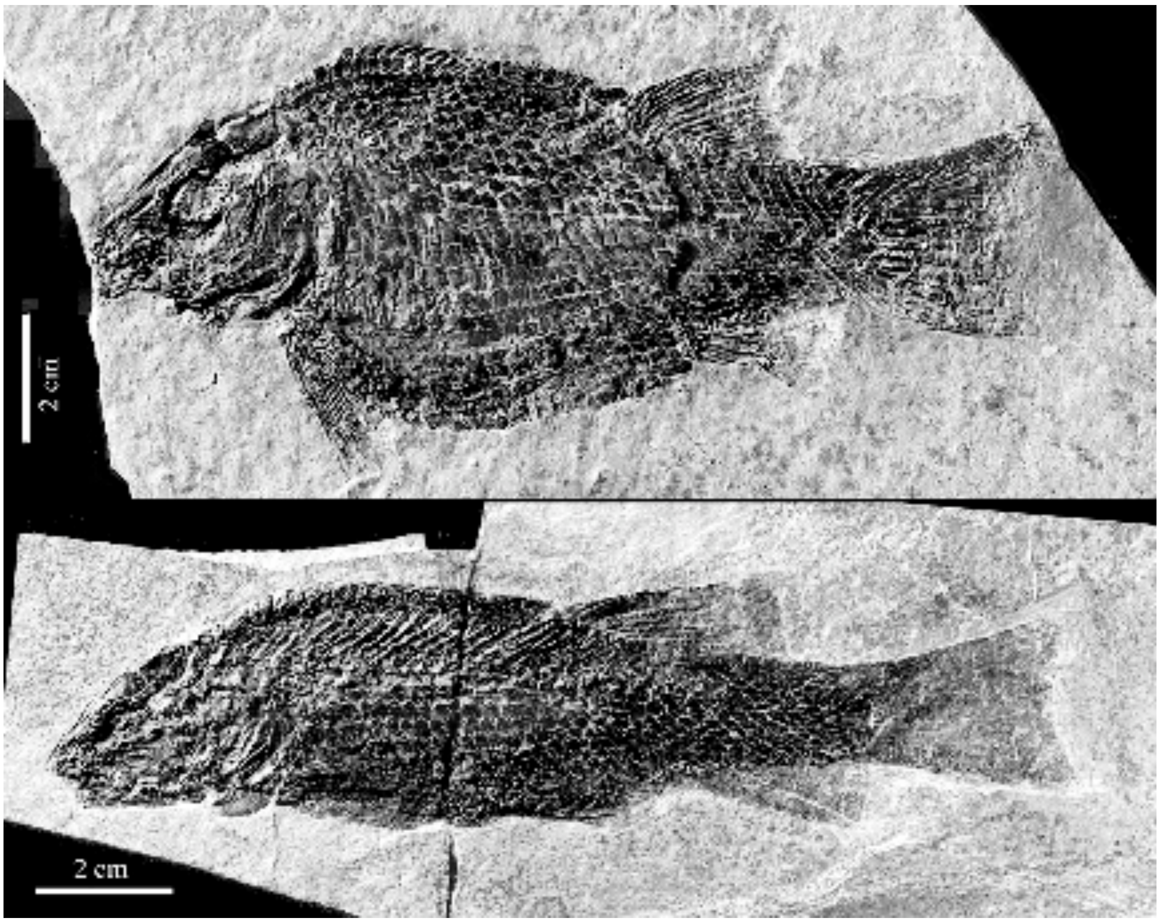


Figure 26: Two examples of the *Semionotus* species flock from the Glen Bed collected by McCune, Stop 5, Station 4.

On the basis of the collection of over 1700 fish made here (Figure 26), McCune described 21 species of *Semionotus* from the microlaminated deep-water part of this cycle, and suggested that a large proportion of them might have evolved *in situ*, endemic to this single lake, analogous to the endemic cichlid fishes of the African great lakes. McCune surveyed more than 2000 museum specimens from 45 localities in eastern North America and showed that six of the species in the Glen Bed were not found in deposits equal in age to or older and thus evolved within the lake level rise and high stand of the Glen Bed, with eight species occurring in older deposits and plausibly supplying the colonizing species. The evidence for seven species was equivocal. McCune (1990) showed that about 5.5% of the specimens have anomalous dorsal ridge scale patterns mixed in with otherwise stereotypic dorsal-ridge-scale patterns, and that dorsal-ridge-scale anomalies are significantly more frequent in older than in younger sediments of the Glen Bed, which she interpreted as being the result of relaxed selection during the early colonization of the lake. Results of McCune's analysis from the Glen Bed are shown in Figure 27.

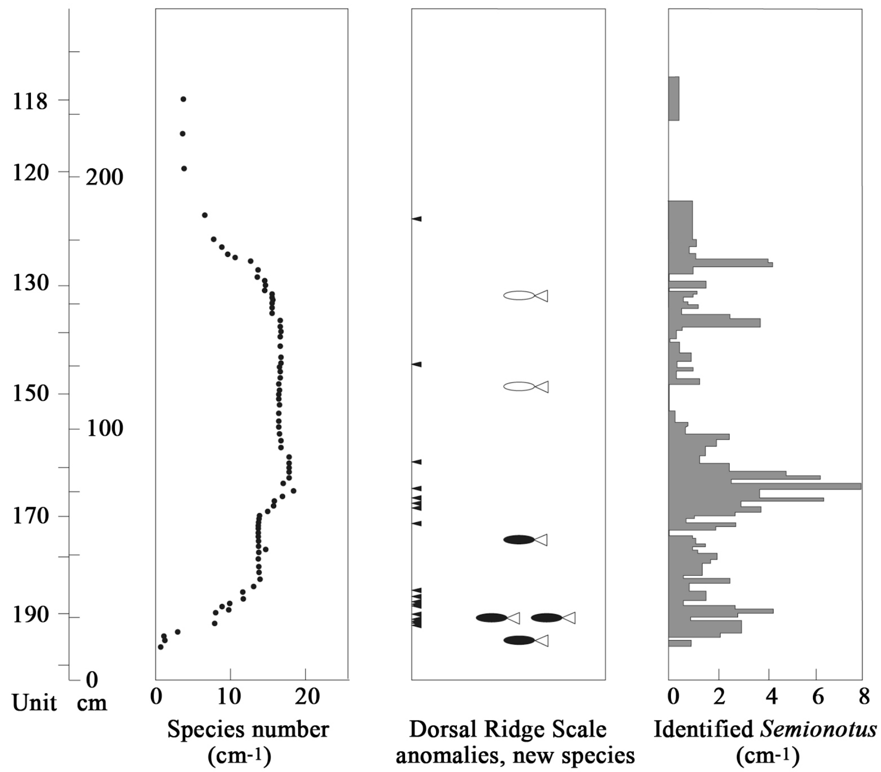


Figure 27: Microstratigraphic distribution of species number and dorsal ridge scales anomalies through 3 m of sediment from the Glen Bed. "Units" are of varying thickness (generally 1–2 cm) with each centimeter of sediment corresponds to about nine years of time (McCune 1990). The right panel plots identified *Semionotus* showing the number of individuals per centimeter of sediment, for which the dorsal ridge scales are visible. The left panel shows the number of species ranges that pass through a given microstratigraphic unit. Arrows in the middle panel indicate layers producing fish with dorsal ridge scales anomalies. First appearances of species are indicated in the same panel by filled fish symbols. Open fish mark singletons; filled outlines are species known from multiple individuals. Adapted from McCune (1990).

Proceed over the ridge in which the quarry is located and walk about 240 ft to the west to the large north-northwest-flowing tributary to the main creek and then proceed about 360 ft south-southwest along the tributary to the large outcrops of the Colfax Bed laminite.

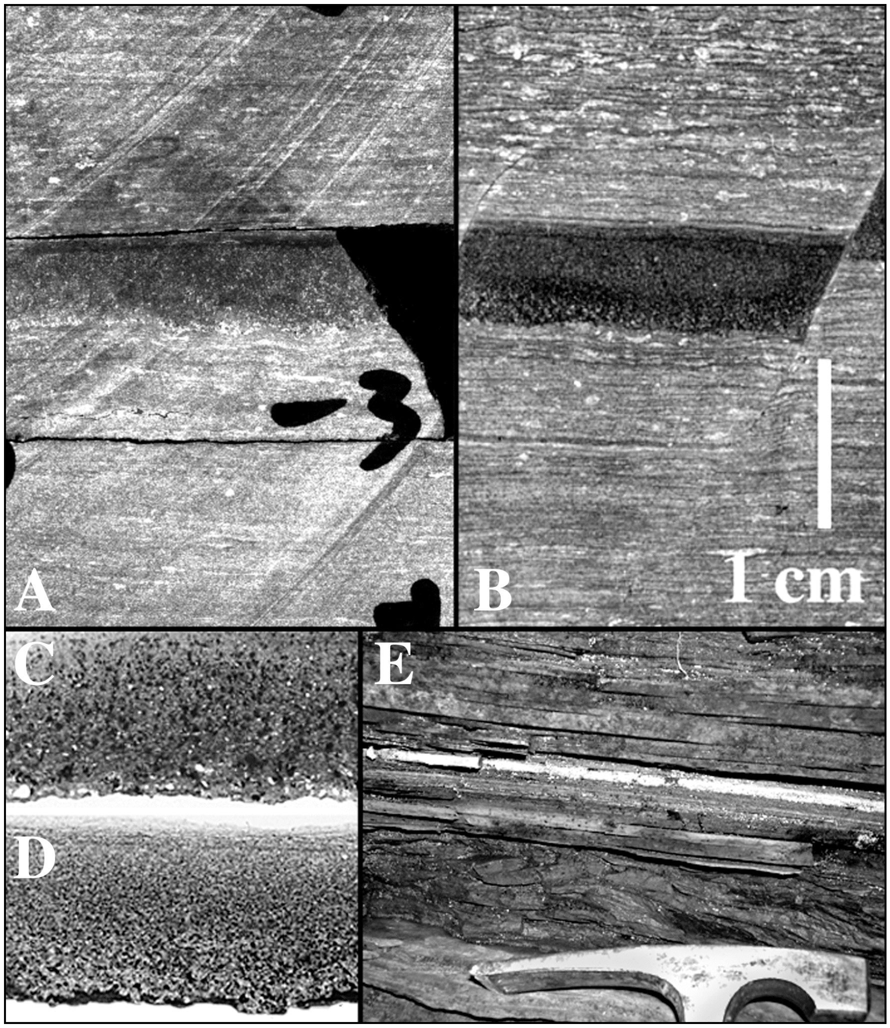


Figure 28: Pompton Tuff: A & B, Pompton Tuff in Towaco ACE core C-128 and slabbed section from the East Berlin Formation (Hartford Basin) "Stevens locality", Parmele Brook, Durham, CT, respectively; C & D, thin sections from the Colfax Bed, Stop 5 (type section) and ACE core C-128, respectively; E, outcrop of the Pompton Tuff (tuff is orange weathering, looking white here), the Colfax Bed, Stop 5, Station 5 (type section).

Station 5 is at about 40° 59.454'N, 74° 16.164'W. Here are outcrops of the Van Houten cycle bearing the Colfax Bed. The Colfax bed is similar to the Glen Bed. It is, however, much less studied. The lower few meters of the laminite have more silt and are thrown into even more folds than are present in the Glen Bed. Fish are not abundant in the Colfax Bed, but this could be because so much of the outcrop is deformed. So far only semionotids have been found.

This is the type section of the Pompton Tuff. After discovery of the small airfall ash in the East Berlin Formation (Olsen et al., 2005), I hypothesized it should be present in the Towaco Formation in the Colfax Bed, based on correlations described by Olsen et al., (1989) and Olsen et al., (1996b). Despite the fact that I had examined the Colfax Bed many times in the last 30 years, I had not noticed any ash layer. However, once knowing where to look, Gustaf I. Olsen and I found it on October 9, 2008 at this site (Figure 28). The ash is thus named after this locality, the Pompton Tuff. Subsequently, I examined the same cycle in the ACE cores and identified the tuff in ACE cores C-128 and PT-14. In the Hartford basin there are abundant conchostracans belonging to the genus *Bublimnadia*. Conchostracans had never been found in the Towaco Formation before but careful examination of the microlaminated interval in the homotaxial position above the Pompton Tuff revealed the same forms (Figure 29). These are representative of the lower trophic levels of consumers.

The extreme similarity in details of the Colfax Bed in terms of lake level phase, conchostracans, and the position of the Pompton Tuff in the correlative Westfield Bed in the Hartford basin would seem unusual in two isolated lakes with separate watersheds. Such similarity would not be surprising if the Colfax and Westfield beds were deposited by the same lake. If they were the deposited by the same lake and this lake extended to the Culpeper basin, it would be larger than the American or African great lakes, or Lake Baikal, as previously discussed.

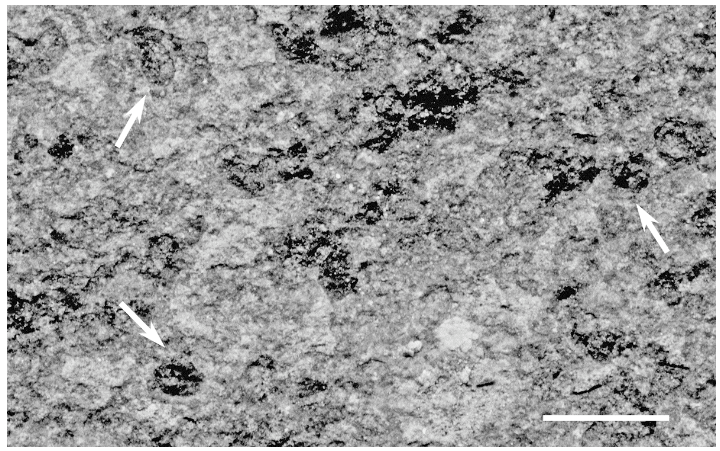


Figure 29: Conchostracans of the genus *Bulblimnadia* from the Colfax Bed, Stop 5, Station 5.

108.8 mi. Head north on Green Knolls Dr/Pines Lake Dr W toward Brook Terrace (80 ft).

108.8 mi. Turn left at Brook Terrace.

109.1 mi. Turn right at Terhune Dr. (US-202).

110.5 mi. Turn left to stay on Ramapo Valley Rd (US-202).

112.4 mi. Take the ramp onto I-287 N.

121.2 mi. Take the I-87 S/I-287/New York Thruway exit toward Tappan Zee Br/New York City.

121.8 mi. Merge onto I-287 E/I-87 S.

125.2 mi. Extensive exposures of Passaic Formation conglomerates.

133.4 mi. Exposures of Palisade diabase.

133.6 mi. Take exit 11 toward US-9W/Nyack/S Nyack.

133.8 mi. Turn left at NY-59 E.

134.2 mi. Turn right at S Highland Ave.

134.4 mi. Continue onto US-9W S/Hillside Ave. Continue to follow US-9W S following strike of the Palisade sill.

141.1 mi. Turn left into entrance of LDEO. Proceed to parking area.

END OF FIELD TRIP

REFERENCES

Anderson, D. L., 2000, The thermal state of the upper mantle; no role for mantle plumes. Geophysical Research Letters v. 27(22), p. 3623-3626.

Austin, J. A. Jr., Stoffa, P. L., Phillips, L. D., Oh, J., Sawyer, D. S., Purdy, G. M., Reiter, E., Makris, J.,1990, Crustal structure of the Southeast Georgia embayment-Carolina Trough: Preliminary results of a composite seismic image of a continental suture (?) and a volcanic passive margin. Geology v.18, p. 1023-1027.

Bailey, R. J, & Smith, D. G., 2008, Quantitative tests for stratigraphic cyclicity. Geological Journal v. 43, p. 431–446.

Barata, C. F. & Caputo, M. V., 2007, Geologia do petróleo da Bacia do Solimõles. O "estado da arte". 4° PDPETRO, Campinas, SP, 21-24 de Outubro de 2007 (ABPG), 1.1.0147, 1-10.

Batygin, K. & Laughlin, G., 2008, On the Dynamical Stability of the Solar System. The Astrophysical Journal v. 683(2), p. 1207–1216.

Benson, R.N., 2003, Age estimates of the seaward-dipping volcanic wedge, earliest oceanic crust, and earliest drift-stage sediments along the North American Atlantic continental margin, in Hames, W., McHone, G., Renne, P., Ruppel, C., (eds.), The Central Atlantic Magmatic Province: Insights from Fragments of Pangea, Geophysical Monograph 136, p. 61-75.

Blackburn, T., Bowring, S., Olsen, P., Kent, D., Rasbury, T., McHone, J. G., 2009, New high-precision U-Pb zircon dating of central Atlantic magmatic province: implications for the Triassic-Jurassic extinction and the astrochronological timescale. Geological Society of America Abstracts with Programs v. 41(7), p. 421.

Bölau, E., 1952, Neue Fossilfunde aus dem Rhät Schones und ihre paläogeographisch-ökologische Auswartung. Geologiska Föreningens i Stockholm Förhandlingar v. 74, p. 44–50.

Brusatte, S.L., Benton, M.J., Ruta, M., Lloyd, G.T., 2008a, Superiority, competition, and opportunism in the evolutionary radiation of dinosaurs. Science v. 321, p. 1485–1488.

Brusatte, S.L., Benton, M.J., Ruta, M., Lloyd, G.T., 2008b, The first 50 mya of dinosaur evolution: macroevolutionary pattern and morphological disparity. Biology Letters v. 4, p. 733–736.

Channell, J. E. T., Kozur, H. W., Sievers, T., Mock, R., Aubrecht, R., Sykora, M. 2003, Carnian-Norian biomagnetostratigraphy at Silické Brezova (Slovakia): correlation to other Tethyan sections and to the Newark Basin. Palaeogeography, Palaeoclimatology, Palaeoecology 191 (2003) 65-109.

Cirilli, S., Marzoli, A., Tanner, L., Bertrand, H., Buratti, N., Jourdan, F., Bellieni, G., Kontak, D., Renne, P.R. 2009. Latest Triassic onset of the Central Atlantic Magmatic Province (CAMP) volcanism in the Fundy Basin (Nova Scotia): New stratigraphic constraints. Earth and Planetary Science Letters 286 (3-4), 514-525.

Coltice, N., Bertrand, H., Rey, P., Jourdan, F., Phillips, B. R., Ricard, Y., 2009, Global warming of the mantle beneath continents back to the Archaean. Gondwana Research v. 15, p. 254-266.

Cohen, A.S., Stone, J., Beuning, K., Park, L., Reinthal, P., Dettman, D, Scholz, C.A., Johnson, T., King, J. W., Talbot, M., Brown, E., Ivory, S., 2007, Ecological Consequences of Early Late-Pleistocene Megadroughts in Tropical Africa. Proceedings of the National Academy of Sciences v. 104,p. 16422-16427.

Diacu, F. & Holmes, P., 1996, Celestial encounters: the origins of chaos and stability. Princeton University Press, 237 p.

Deenen, M. H. L., Ruhl, M., Bonis, N.R. Krijgsman,W., Kuerschner, W.M., Reitsma, M., van Bergen, M.J., 2010, A new chronology for the end-Triassic mass extinction. Earth and Planetary Science Letters v. 291, p. 113–125.

Dickens, G. R., Castillo, M. M., Walker, J. C. G., 1997, A blast of gas in the latest Paleocene; simulating first-order effects of massive dissociation of oceanic methane hydrate. Geology v. 25(3), p. 259–262.

Dyni, J. R. & Hawkins, J. E., 1981, Lacustrine turbidites in the Green River Formation of northwestern Colorado. Geology v. 9, p. 235-238.

Ernst, R. E., Grosfils, E. B., Mege, D., 2001, Giant dyke swarms on Earth, Venus and Mars. Annual Review of Earth and Planetary Sciences v. 29, p. 489–534,

Faust, G. T., 1975, A review and interpretation of the geologic setting of the Watchung Basalt flows. New Jersey. U.S. Geological Survey Professional Paper 864, p. A1-A42.

Fedosh, M. S., & Smoot, J. P., 1988, A cored stratigraphic section through the northern Newark basin, New Jersey. U. S. Geological Survey Bulletin 1776, p. 19-24.

Fowell, S. J. & Traverse, A. ,1995, Palynology and age of the upper Blomidon Formation, Fundy basin, Nova Scotia. Review of Palaeobotany and Palynology v. 86, p. 211-233.

Geissman, J. W., Olsen, P. E., Kent, D. V., 2010, Site selected for Colorado Plateau coring. Eos Transactions American Geophysical Union, Supplement v. 91(14), p. 128.

Goldberg, D. S., Kent, D. V., Olsen, P. E., 2009, Potential on-shore and off-shore reservoirs for CO_2 sequestration in Central Atlantic Magmatic Province (CAMP) Basalts. Proceedings of the National Academy of Sciences: doi:10.1073/pnas.0913721107.

Gradstein, F.M. and Sheridan, R.S., 1983. On the Jurassic Atlantic Ocean and a synthesis of results of Deep Sea Drilling Project Leg 76; in Sheridan, R.S., Initial Reports of the Deep Sea Drilling Project., U. S. Government Printing Office v. 76, p. 913-945.
Hallam, A., Wignall, P.B., 1997, Mass Extinctions and their Aftermath. Oxford Univ. Press, Oxford. 320 p.
Hansen, V. I. & Olive, A., 2010, Artemis, Venus: The largest tectonomagmatic feature in the solar system? Geology v. 38, p. 467-470.
Hesselbo, S.P., Robinson, S.A., Surlyk, F. Piasecki, S. 2002. Terrestrial and marine extinction at the Triassic-Jurassic boundary synchronized with major carbon-cycle perturbation: a link to initiation of massive volcanism? Geology v. 30, p. 251-254..
Hill, R. I., 1991 , Starting plumes and continental break-up. Earth and Planetary Science Letters v. 104, p. 398-416.
Hillebrandt, A. V., Krystyn, L. & Kuerchner, W. M. 2007. A candidate GSSP for the base of the Jurassic in the Northern Calcareous Alps (Kuhjoch section, Karwendel Mountains, Tyrol, Austria). International Subcommission on Jurassic Stratigraphy, Newsletter v. 34(1), p. 2-20.
Holbrook, W.S., Reiter, E. C., Purdy, G. M., Sawyer, D., Stoffa, P. L., Austin, Jr., A., Oh, J. and Makris, J., 1994b, Deep structure of the U.S. Atlantic continental margin, offshore South Carolina, from coincident ocean bottom multichannel seismic data. Journal of Geophysical Research v. 99(B5), p. 9,155-9,178.
Hozik, M. J., 1992, Paleomagnetism of igneous rocks in the Culpeper, Newark, and Hartford/Deerfield basins, in Puffer, J. H., & Ragland, P. C. (eds.), Eastern North American Mesozoic magmatism. Geological Society of America Special Paper 268, p. 279-308.
Irving, E. & Banks, M.R., 1961, Paleomagnetic results from the Upper Triassic lavas of Massachusetts. Journal of Geophysical Research v. 66, p. 1935-1939.
Johnson, T. C., Scholz, C. A., Talbot, M. R., Kelts, K., Ricketts, R. D., Ngobi, G., Beuning, K., Ssemmanda, I., McGill, J. W., 1996, Late Pleistocene Desiccation of Lake Victoria and Rapid Evolution of Cichlid Fishes. Science v. 273(5278), p. 1091-1093.
Kelemen, P.B., and Holbrook, W.S., 1995, Origin of thick, high-velocity crust along the U.S. East Coast margin. Journal of Geophysical Research v. 100, p. 10,077-10,094.
Kent, D. V. & Olsen, P. E., 2008, Early Jurassic magnetostratigraphy and paleolatitudes from the Hartford continental rift basin (eastern North America): Testing for polarity bias and abrupt polar wander in association with the central Atlantic magmatic province. Journal of Geophysical Research v. 113, B06105, doi:10.1029/2007JB005407.
Kidder, D. L. & Worsley, T. R., 2010, Phanerozoic large igneous provinces (LIPs), HEATT (haline euxinic acidic thermal transgression) episodes, and mass extinctions. Palaeogeography, Palaeoclimatology, Palaeoecology v. 295, p. 162–191
Klitgord, K. D. & Schouten, H., 1986, Plate kinematics of the central Atlantic, in Vogt, P.R. & Tucholke, B.E., (eds.), The Geology of North America, v. M., The Western North Atlantic Region, Geological Society of America, p. 351-378.
Kominz, M. A., Beavan, J., Bond, G. C., McManus, J., 1991, Are cyclic sediments periodic? Gamma analysis and spectral analysis of Newark Supergroup lacustrine strata. Kansas Geological Survey Bulletin 233, p. 231-252.

Kominz, M.A. and Bond, G.C., 1990, A new method of testing periodicity in cyclic sediment: Application to the Newark Supergroup. Earth and Planetary Science Letters v. 98, p. 233-244.

Laskar, J., 1999, The limits of Earth orbital calculations for geological time scale use, Philosophical Transactions of the Royal Society of London A v. 357, p. 1735-1759 (1999)

Laskar, J. & Gastineau, M., 2009, Existence of collisional trajectories of Mercury, Mars and Venus with the Earth. Nature v. 459, p. 817-819.

Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A. C. M., Levrard, B., 2004, A long-term numerical solution for the insolation quantities of the Earth. Astronomy & Astrophysics v. 428, p. 261-285.

LeTourneau, P.M., 1985a, Alluvial fan development in the Lower Jurassic Portland Formation, central Connecticut: implications for tectonics and climate, in Robinson, G.R., Jr. & Froelich, A. J. (eds.), Proceedings of the Second U. S. Geological Survey Workshop on the Early Mesozoic Basins of the Eastern United States: U.S. Geological Survey Circular 946, p. 17·26.

LeTourneau, P. M., 1985b, The Sedimentology and Stratigraphy of the Lower Jurassic Portland Formation, Central Connecticut. M.A. Thesis, Wesleyan University, Middletown, CT, 247 p.

LeTourneau, P. M. & Huber, P., 2006, Early Jurassic eolian dune field, Pomperaug basin, Connecticut and related synrift deposits: Stratigraphic framework and paleoclimatic context. Sedimentary Geology v. 187, p. 63–81.

MacArthur, R. H. & Wilson, E. O., 1967, The Theory of Island Biogeography. Princeton University Press, Princeton, NJ.

Malinconico, M. L., 2003, Estimates of eroded strata using borehole vitrinite reflectance data, Triassic Taylorsville rift basin, Virginia: Implications for duration of syn- rift sedimentation and evidence of structural inversion in LeTourneau, P. M., and Olsen, P. E. (eds.), The Great Rift Valleys of Pangea in Eastern North America (Volume 1, Tectonics, Structure, and Volcanism): Columbia University Press, New York, p. 80-103.

Malinconico, M. L., 2009, Synrift to early postrift basin-scale groundwater history of the Newark basin based on surface and borehole vitrinite reflectance data, in Herman, G. C. and Serfes, M. E. (eds.), Contributions to the Geology and Hydrogeology of the Newark Basin. New Jersey Geological Survey Bulletin 77, p. C1-C38.

Manspeizer, W., 1980, Rift tectonics inferred from volcanic and clastic structures, in W. Manspeizer (ed.), Field Studies in New Jersey Geology and Guide to Field Trips, 52nd Ann. Mtg. New York State Geol. Assoc., Newark College of Arts and Sciences, Newark, Rutgers University, p. 314-350.

Marzoli, A., Knight, K. B., Cirilli, S., Buratti, N., Vérati, C., Nomade, S., Renne, P. R., Youbi, N., Martini, R., Allenbach, K., Neuwerth, R., Rapaille, C., Zaninetti, L., Bellieni, G., 2004, Synchrony of the Central Atlantic magmatic province and the Triassic-Jurassic boundary climatic and biotic crisis. Geology v. 32, p. 973–976.

Matter, J. M., Takahashi, T., Goldberg, D., 2007, Experimental evaluation of in situ CO_2-water-rock reactions during CO_2 injection in basaltic rocks: Implications for geological CO_2 sequestration. Geochemistry, Geophysics, Geosystems 8:Q02001 doi: 10.1029/2006GC001427.

Mattinson, J.M., 2005, Zircon U-Pb chemical abrasion (“CA-TIMS”) method: Combined annealing and multi-step dissolution analysis for improved precision and accuracy of zircon ages. Chemical Geology v. 220, p. 47–56.

May, P. R., 1971, Pattern of Triassic diabase dikes around the North Atlantic in the context of predrift position of the continents. Geological Society of America Bulletin v. 82, p. 1285-1292.

McCune, A. R., 1982, Early Jurassic Semionotidae (Pisces) from the Newark Supergroup: systematics and evolution of a fossil species flock. Ph.D. dissertation. Yale University, New Haven, CT, 371 p.

McCune, A. R., 1986. A revision of *Semionotus*, with redescriptions of valid European species. Palaeontology v. 29, p. 213-233.

McCune, A. R., 1987a, Toward the phylogeny of a fossil species flock: Semionotid fishes from a lake deposit in the Early Jurassic Towaco Formation, Newark Basin. Yale Peabody Museum of Natural History Bulletin 43, p. 1-108.

McCune, A. R., 1987b, Lakes as laboratories of evolution: endemic fishes and environmental cyclicity. Palaios v. 2, p. 446-454.

McCune, A. R., 1990, Evolutionary novelty and atavism in the *Semionotus* complex – relaxed selection during colonization of an expanding lake. Evolution v. 44(1), p. 71-85.

McCune, A. R., 1996, Biogeographic and stratigraphic evidence for rapid speciation in semionotid fishes. Paleobiology v. 22 (1), p. 34-48.

McCune, A. R. 1997, How fast is speciation: molecular, geological and phylogenetic evidence from adaptive radiations of fishes, in Givnish, T. & Sytsma, K., editors, Molecular Evolution and Adaptive Radiation. Cambridge University Press, p. 585-610.

McCune, A. R., 2004, Diversity and speciation of semionotid fishes in Mesozoic rift lakes. in Adaptive Speciation, Dieckmann, U., Doebeli, M., Metz, J. A. J., Tautz, D. (eds.), Cambridge University press, p. 362-379.

McCune, A. R., Thomson, K. S. & Olsen, P. E., 1984, Semionotid fishes from the Mesozoic Great Lakes of North America, in Evolution of Fish Species Flocks, in Echelle, A. A. & Kornfield, I. (eds.), Orono, ME, USA, University of Maine Press, p. 22-44.

McElwain, J. C., Beerling, D. J., Woodward, F. I., 1999, Fossil plants and global warming at the Triassic-Jurassic boundary. Science v. 285, p. 1386–1390.

McElwain, J. C., Wagner, P. J., Hesselbo, S. P., 2009, Fossil plant relative abundances indicate sudden loss of Late Triassic biodiversity in East Greenland. Science v. 324, p. 1554–1556.

McHone, J. G., 1996, Broad-terrane Jurassic flood basalts across northeastern North America, Geology v. 24(4), p. 319-322.

McHone, J. G., 2000, Non-plume magmatism and rifting during the opening of the central Atlantic Ocean. Tectonophysics v. 316(3-4), p. 287-296.

McLaughlin, D. B., 1933, A note on the stratigraphy of the Brunswick Formation (Newark) in Pennsylvania. Michigan Academy of Science, Arts, and Letters v. 18, p. 59–74.

McLaughlin, D. B., 1944, Triassic stratigraphy in the Point Pleasant district, Pennsylvania. Pennsylvania Academy of Science Proceedings v. 18, p. 62–69.

McLaughlin, D. B., 1946a, The Triassic rocks of the Hunterdon Plateau, New Jersey. Proceedings of the Pennsylvania Academy of Science v. 20, p. 89–98.
McLaughlin, D. B., 1946b, Continuity of strata in the Newark Series: Michigan Academy of Science, Arts, and Letters, v. 32, p. 295–303.
McLaughlin, D. B., 1959, Mesozoic rocks, in Willard, B., et al., Geology and mineral resources of Bucks County, Pennsylvania: Pennsylvania Geological Survey Bulletin C-9, p. 55–114.
Medina, F., Olsen, P., Et-Touhami, M., Bouaouda, M-S., Hafid, M., 2007, Permian-Triassic rifting and inverted Jurassic-Cretaceous coastal basin in western High Atlas, MAPG First Congress and Exhibition Marrakech, 28-30 October 2007, Field Trip A1, Guidebook, Moroccan Association of Petroleum Geologists, Marrakech, Morocco. 85 p.
Milani, E. J. & Zalán, P. V., 1999, An outline of the geology and petroleum systems of the Paleozoic interior basins of South America. Episodes, v. 22(3), p.
Milner, A. R. C. & Kirkland, J. I., 2007, The case for fishing dinosaurs at the St. George dinosaur discovery site at Johnson Farm. Utah Geological Survey, Survey Notes v. 39(3), p. 1-3.
Morton, N. 2008a. Selection and voting procedures for the base Hettangian. International Subcommission on Jurassic Stratigraphy, Newsletter v. 35(1), p. 67.
Morton, N., 2008b, Details of voting on proposed GSSP and ASSP for the base of the Hettangian Stage and Jurassic System. International Subcommission on Jurassic Stratigraphy, Newsletter 35 (1), 74.
Morton, N., Warrington, G., Bloos, G., 2008, Foreword. International Subcommission on Jurassic Stratigraphy, Newsletter v. 35(1), p. 68-73.
Morton, N., 2010, Progress in defining Jurassic stages – a learning process. Earth Science Frontiers v. 17 (special edition), p. 66-67.
Mutter, J. C., Talwani, M. Stoffa, P. L., 1982, Origin of seaward-dipping reflectors in the oceanic crust off the Norwegian margin by "subaerial sea-floor spreading". Geology v. 10, p. 353-357.
Muttoni, G., Kent, D. V., Olsen, P. E., DiStefano, P., Lowrie, W., Bernasconi, S., Hernandez, F.M., 2004, Tethyan magnetostratigraphy from Pizzo Mondello (Sicily) and correlation to the Late Triassic Newark astrochronological polarity time scale, Geological Society of America Bulletin v. 116(9/10), p. 1043–1058.
Muttoni, G., Kent, D. V., Rigo, M., Nicora, A., Olsen, P. E., Jadoul, F., Galli, M. T., 2009, Rhaetian magnetobiostratigraphy from the Southern Alps (Italy): constraints on Triassic chronology. Palaeogeography, Palaeoclimatology, Palaeoecology v. 285 (1-2), p. 1-16.
Mayr, E., 1963, Animal Species and Evolution (Harvard Univ. Press, Cambridge, MA), 453 p.
Oh, J., Austin, J.A., Phillips, J.D. Coffin, M.F., Stoffa, P.L., 1995, Seaward-dipping reflectors offshore the southeastern United States: Seismic evidence for extensive volcanism accompanying sequential formation of the Carolina trough and Blake Plateau basin. Geology v. 23, p. 9-12.
Oliveira, E. P., Tarney, J., Joao , X. J.,1990, Geochemistry of the Mesozoic Amapa and Jari dyke Swarms, Northern Brazil: plume-related magmatism during the opening of

the Central Atlantic. In: A.J. Parker, P.C. Rickwood and D.H. Tucker, Editors, Mafic Dykes and Emplacement Mechanisms, Balkena, Rotterdam (1990), pp. 173–183.

Olsen, P. E., 1980a, Triassic and Jurassic formations of the Newark basin, in Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: New York State Geological Association, 52nd Annual Meeting, Newark, New Jersey, Rutgers University, p. 2-39.

Olsen, P. E., 1980b, Fossil great lakes of the Newark Supergroup in New Jersey. in W. Manspeizer (ed.), Field Studies in New Jersey Geology and Guide to Field Trips, 52nd Ann. Mtg. New York State Geol. Assoc., Newark College of Arts and Sciences, Newark, Rutgers University, p. 352-398.

Olsen, P. E., 1980c, The Latest Triassic and Early Jurassic Formations of the Newark Basin (Eastern North America, Newark Supergroup): Stratigraphy, Structure, and Correlation: New Jersey Academy of Science Bulletin v. 25, p. 25-51.

Olsen, P. E., 1997, Stratigraphic record of the early Mesozoic breakup of Pangea in the Laurasia-Gondwana rift system. Annual Reviews of Earth and Planetary Science v. 25, p. 337-401.

Olsen, P. E., Fairfield, H. M., Hemming, S., 1999, Stratigraphic and Geochemical Evidence of the Past Distribution of CAMP Basalts. . Eos, Transactions, American Geophysical Union, Supplement v. 80(17), p. S318.

Olsen, P. E., Kent, D V., Cornet, B., Witte, W. K., Schlische, R. W., 1996a, High-resolution stratigraphy of the Newark rift basin (Early Mesozoic, Eastern North America). Geological Society of America, v. 108, p. 40-77.

Olsen, P. E., Kent, D. V., Geissman, J. W., 2008, One hundred million years of climatic, tectonic, and biotic evolution in continental cores. Eos Transactions American Geophysical Union, Supplement v. 89(12), p. 118.

Olsen, P. E., Kent, D. V., Sues, H.-D., Koeberl, C., Huber, H., Montanari, A., Rainforth, E. C., Fowell, S. J., Szajna, M. J. & Hartline, B. W. 2002a, Ascent of dinosaurs linked to an iridium anomaly at the Triassic-Jurassic boundary. Science v. 296, p. 1305-1307.

Olsen, P. E., Kent, D. V., Et-Touhami, M., and Puffer, J. H., 2003, Cyclo-, magneto-, and bio-stratigraphic constraints on the duration of the CAMP event and its relationship to the Triassic-Jurassic boundary, in Hames, W.E., McHone, J.G., Renne, P.R, Ruppel, C. (eds.), The Central Atlantic Magmatic Province: Insights From Fragments of Pangea, Geophysical Monograph Series v. 136, p. 7-32.

Olsen, P. E., Kent, D. V., Whiteside, H., 2010, Implications of the Newark Supergroup-based astrochronology and geomagnetic polarity time scale (Newark-APTS) for the tempo and mode of the early diversification of the Dinosauria. Earth and Environmental Science Transactions of the Royal Society of Edinburgh (in press).

Olsen, P. E. & McCune, A. R., 1991, Morphology of the *Semionotus elegans* species group from the Early Jurassic part of the Newark Supergroup of Eastern North America with comments on the family Semionotidae (Neopterygii). Journal of Vertebrate Paleontology 11(3), 269-292.

Olsen, P. E., McCune, A. R. and Thomson, K. S., 1982, Correlation of the early Mesozoic Newark Supergroup by Vertebrates, principally fishes. American Journal of Science v. 282, p. 1-44.

Olsen P. E, Schlische R. W, Fedosh M. S., 1996b, 580 ky duration of the Early Jurassic flood basalt event in eastern North America estimated using Milankovitch cyclostratigraphy, in Morales, M. (ed.) The Continental Jurassic, Museum of Northern Arizona Bulletin 60, p. 11-22.

Olsen, P. E., Schlische, R. W., Gore, P. J. W. (and others), 1989, Field Guide to the Tectonics, stratigraphy, sedimentology, and paleontology of the Newark Supergroup, eastern North America. International Geological Congress, Guidebooks for Field Trips T351, 174 p.

Olsen, P. E., Smith, J. B., and McDonald, N. G., 1998, Type material of the type species of the classic theropod footprint genera *Eubrontes*, *Anchisauripus*, and *Grallator* (Early Jurassic, Hartford and Deerfield basins, Connecticut and Massachusetts, USA). Journal of Vertebrate Paleontology v. 18(3), p. 586-601.

Olsen, P. E., Whiteside, J. H., LeTourneau, P. M., Huber, P., 2005, Jurassic cyclostratigraphy and paleontology of the Hartford basin. In B.J. Skinner and A.R. Philpotts (eds.), 97th New England Intercollegiate Geological Conference, Department of Geology and Geophysics, Yale University, New Haven, Connecticut, p. A4-1 – A4-51.

Pálfy, J. & Mundil, R. 2006. The age of the Triassic/Jurassic boundary: new data and their implications for the extinction and recovery. Volumina Jurassica v. 1, p. 294.

Prévot, M. & McWilliams, M., 1989, Paleomagnetic correlation of Newark Supergroup volcanics. Geology v. 17, p. 1007-1010.

Puffer, J. H., 1992, Eastern North American flood basalts in the context of the incipient breakup of Pangea. in Eastern North American Mesozoic magmatism, edited by J. H. Puffer and P. C. Ragland, Geological Society of America Special Paper 268, p. 95-118.

Puffer, J.H., 2003, Geochemistry of Pangean and Rodinian continental flood basalts, in LeTourneau, P. M. and Olsen, P. E., (eds.), The Great Rift Valleys of Pangea in Eastern North America, Volume 1, Tectonics, Structure, and Volcanism: New York, Columbia University Press, Chapter 11. p. 155-171.

Puffer, J. H. & Volkert, R. A., 2001, Pegmatoid and gabbroid layers in Jurassic Preakness and Hook Mountain basalts, Newark Basin, New Jersey. The Journal of Geology v. 109, p. 585-601.

Rasbury, E. T., Dewet, C. B., Nienstedt, J., 2003, U-Pb age of stromatolites calcite from the Triassic Passaic ormation of the Newark basin. Geological Society of America Abstracts with Programs v. 35(6), p. 508.

Redfield, W. C., 1842, ART. XIII.-Notice of newly discovered Fish Beds and a Fossil Foot Mark in the Red Sandstone Formation of New Jersey. American Journal of Science and Arts (1820-1879) v. 44(1), p. 134-136.

Ruhl, M., Deenen, M. H. L., Abels, H. A., Bonis, N. R., Krijgsman, W., Kürschner, W. M., 2010, Astronomical constraints on the duration of the early Jurassic Hettangian stage and recovery rates following the end-Triassic mass extinction (St. Audrie's Bay/East Quantoxhead, UK). Earth and Planetary Science Letters v. 295(1-2), p. 262-276.

Russell, I. C., 1880, On the former extent of the Triassic formation of the Atlantic states: American Naturalist v. 14, p. 703-712.

Sanders, J., 1963, Late Triassic tectonic history of northeastern United States. American Journal of Science v. 261, p. 501–524.

Schettino, A. & Turco, E., 2009, Breakup of Pangaea and plate kinematics of the central Atlantic and Atlas regions. Geophysical Journal Intinternational v.178, p. 1078–1097.

Schlische, R.W., 2003. Progress in understanding the structural geology, basin evolution, and tectonic history of the eastern North American rift system, in The Great Rift Valleys of Pangea in Eastern North America,Vol.1: Tectonics, Structure, and Volcanism, pp. 21–64, eds LeTourneau, P.M. & Olsen, P.E., Columbia University Press, New York, USA.

Schlische, R.W., Withjack, M.O. & Olsen, P.E., 2003. Relative timing of CAMP, rifting, continental breakup, and inversion: tectonic significance, in The Central Atlantic Magmatic Province: Insights from Fragments of Pangea, Geophysical Monograph,Vol. 136, pp. 33–59, eds Hames, W.E.,McHone, G.C., Renne, P.R.&Ruppel, C.R., American Geophysical Union, Washington, DC, USA.

Schlische, R.W., 2003, Progress in understanding the structural geology, basin evolution, and tectonic history of the eastern North American rift system. in LeTourneau, P.M., and Olsen, P.E. (eds.), The Great Rift Valleys of Pangea in Eastern North America--Volume 1--Tectonics, Structure, and Volcanism: New York, Columbia University Press, p. 21-64.

Schlische, R.W., Withjack, M.O., 2005, The early Mesozoic Birdsboro central Atlantic margin basin in the Mid-Atlantic region, eastern United States: Discussion. Geological Society of America Bulletin v. 117, p. 823-828.

Schoene, B., Guex, J., Bartolini, A., Schaltegger, U., Blackburn, T. J., 2010, Correlating the end-Triassic mass extinction and flood basalt volcanism at the 100 ka level Geology v. 38(5), p, 387-390.

Scofield, R. P. & Ashwell, K. W. S., 2009, Rapid somatic expansion causes the brain to lag behind: the case of the brain and behavior of New Zealand's Haast's Eagle (*Harpagornis moorei*). Journal of Vertebrate Paleontology v. 29(3), p. 637-649.

Sereno, P.C., 1991, Basal archosaurs: phylogenetic relationships and functional implications. Society of Vertebrate Paleontology Memoir 2, p. 1–53.

Sereno, P. C., 2005, Stem Archosauria—TaxonSearch. URL http://www.taxonsearch.org/Archive/stem-archosauria-1.0.php.

Smoot, J. P., 2010, Chapter A: Triassic Depositional Facies in the Newark Basin, in Herman, G. C. & Serfes, M. E. (eds.), Contributions to the Geology and Hydrogeology of the Newark Basin, New Jersey Geological Survey Bulletin 77, p. A1-A110.

Smoot, J.P. & Lowenstein, T.K., 1991, Depositional environments of non-marine evaporites, in Melvin, J.L. (ed.), Evaporites, Petroleum and Mineral Resources, Developments in Sedimentology 50, Elsevier, Amsterdam, The Netherlands, p. 189-347.

Sundman, K. E., 1912, Memoire sur le probleme de trois corps. Acta Mathematica v. 36 p. 105–179.

Svensen, H., Planke, S., Malthe-Sorenssen, A., Jamtveit, B., Myklebust, R., Rasmussen Eidem, T., et al., 2004, Release of methane from a volcanic basin as a mechanism for initial Eocene global warming. Nature v. 429, p. 542–545.

Svensen, H., Planke, S., Polozov, A. G., Schmidbaue, N., Corfu, F., Podladchikov, Y. Y., Jamtveit, B., 2009, Siberian gas venting and the end-Permian environmental crisis. Earth and Planet Science Letters v. 277, p. 490–500.

Talwani, M., Ewing, J., Sheridan, R. E., Holbrook, W. S., Glover, L. III., 1995, The EDGE experiment and the U.S. Coast Magnetic anomaly, in Talwani, M., and Torne, M. (eds.), NATO/ARW Series book, Rifted Ocean-Continent Boundaries, Banda, p. 155-181.

Tolan T. L., Reidel, S. P. Beeson, M. H. Anderson, J. L. Fecht, K. R. Swanson, D. A., 1989, Revisions to the estimates of the area extent and volume of the Columbia River Basalt Group, in Reidel, S. P. & Hooper, P. R. (eds.), Volcanism and Tectonism in the Columbia River Flood-Basalt Province, Geological Society of America Special Paper 239, The Geological Society of America, Boulder, Colorado, p. 1–20.

Tollo, R. P. & Gottfried, D., 1992, Petrochemistry of Jurassic basalt from eight cores, Newark basin, New Jersey, *in* Puffer. J. H. and Ragland, P. C. (eds.), Eastern North American Mesozoic Magmatism, Geological Society of America Special Paper 268, p. 233-260.

van Dam, J. A., Abdul Aziz, H., Sierra, M. Á. Á., Hilgen, F. J., van den Hoek Ostende, Lourens, L. L., Mein, P., van der Meulen, A. J., Pelaez-Campomanes, P., 2006, Long-period astronomical forcing of mammal turnover. Nature v. 443, p. 687-691.

van de Schootbrugge B., Tremolada, F., Rosenthal, Y., Bailey, T. R., Feist-Burkhardt, S., Brinkhuis, H., Pross, J., Kent, D. V., Falkowski, P.G., 2007, End-Triassic calcification crisis and blooms of organic-walled ‘disaster species’. Palaeogeography, Palaeoclimatology, Palaeoecology v. 244, p. 126–141.

Van Houten, F. B., 1962, Cyclic sedimentation and the origin of analcime-rich upper Triassic Lockatong Formation, westcentral New Jersey and adjacent Pennsylvania: American Journal of Science v. 260, p. 561–576.

Van Houten, F. B., 1964, Cyclic lacustrine sedimentation, Upper Triassic Lockatong Formation, central New Jersey and adjacent Pennsylvania, in Mermiam, O. F., ed., Symposium on cyclic sedimentation: Kansas Geological Survey Bulletin 169, p. 497–531.

Van Houten, F. B., 1969, Late Triassic Newark Group, northcentral New Jersey and adjacent Pennsylvania and New York, in Subitzki, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions, New Brunswick, New Jersey (Geological Society of America, Field Trip 4): Atlantic City, New Jersey, Rutgers University Press, p. 314–347.

Van Houten, F. B., 1980, Late Triassic part of Newark Supergroup, Delaware River section, west central New Jersey, in Manspeizer, W., ed., Field studies of New Jersey geology and guide to field trips: New York State Geological Association, 52nd Annual Meeting, Newark, New Jersey, Rutgers University, p. 264–269.

Vogt, P.R., 1973, Early events in the opening of the North Atlantic, in Implications of Continental Rift to the Earth Sciences, Tarling, D.H. & Runcorn, S.K. (eds.), Academic Press, New York, p. 693-712.

Wang, Q., 1991, The global solution of the n-body problem. Celestial Mechanics and Dynamical Astronomy v. 50(1), p. 73-88.

White and McKenzie, 1989. R.S. White and D. McKenzie , Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. Journal of Geophysical Research v. 94, p. 7685-7729.
Whiteside, J.H., 2004, Arboreal stromatolites: a 210-million year record, in Lowman, M.D., and Rinker, B. (eds.), Forest Canopies (Physiological Ecology Series), 2nd edition. Academic Press, p. 147-149.
Whiteside, J. H., Olsen, P. E., Eglinton, T., Cornet, B., McDonald, N. G., Huber, P., 2010b, Pangean great lake paleoecology on the cusp of the end-Triassic extinction. Palaeogeography, Palaeoclimatology, Palaeoecology in review.
Whiteside, J. H., Olsen, P. E., Eglinton, T., Montluçon, D., Brookfield, M. E. & Sambrotto, R. N. 2010a. Molecular-level carbon isotopes from Earth's largest flood basalt province directly link eruptions to the end-Triassic mass extinction. Proceedings of the National Academy of Sciences v. 107(15), p. 6721-6725.
Whiteside, J. H., Olsen, P. E., Kent, D,V., Fowell, S. J., Et-Touhami, M., 2007, Synchrony between the CAMP and the Triassic-Jurassic mass-extinction event? Palaeogeography, Palaeoclimatology, Palaeoecology v. 244(1-4), p. 345-367.
Wilson, E. O., 1961, The Nature of the Taxon Cycle in the Melanesian Ant Fauna. The American Naturalist v. 95(882), p. 169-193.
Withjack, M. O. & Schlische, R. W., 2005, A review of tectonic events on the passive margin of eastern North America. in Post, P., ed., Petroleum Systems of Divergent Continental Margin Basins: 25th Bob S. Perkins Research Conference, Gulf Coast Section of SEPM, p. 203-235.

Geochemical Characterization of New York City Schist Formations

John H. Puffer[1], Alan I. Benimoff[2], and Jeffrey Steiner[3]

[1]Department of Earth & Environmental Sciences, Rutgers University, Newark, New Jersey, 07102

[2]Department of Engineering Science and Physics, College of Staten Island, 2800 Victory Boulevard, Staten Island, NY 10314

[3]Department of Earth and Atmospheric Sciences, City College of New York, 138th and Convent Avenue, New York City, New York 10031.

Abstract

New geochemical analyses of bedrock from Manhattan, New York support a more widespread distribution of the Manhattan Schist than indicated on recent geologic maps of New York City including Schuberth (1967); Baskerville (1994); Merguerian and Merguerian (2004) and Brock and Brock (2001). The 39 samples of schist from outcrops and roadcuts across all of Central Park, Manhattan plot as a single coherent population on all major element and trace element variation diagrams. Plots of K_2O/Na_2O vs. SiO_2 and K_2O/Na_2O vs. SiO_2/Al_2O_3 on the tectonic setting discrimination diagrams of Roser and Korsch (1986) are consistent with a passive margin to active continental margin depositional setting. Major elements also plot within the compositional field of the Martinsburg Formation, an Ordovician, largely continental shelf meta-pellite and a probable stratigraphic correlative of Manhattan Schist. In addition, samples mapped as Hartland Formation collected west of Central Park along the Hudson River plot within the continuum of the Central Park samples. An arc detritus depositional setting as generally proposed for Hartland deposition is notably un-represented throughout Manhattan. However, east of Manhattan, particularly Pelham Bay, an arc signature is seen. Our geochemical data together with ambiguous Hartland/Manhattan schist petrology underscores the need to establish a reliable mappable criteria capable of distinguishing Hartland from Manhattan schists.

Introduction

There are currently at least four competing geologic maps of New York City that present highly differing formation distributions, particularly Manhattan Schist and Hartland Formation. In addition each of four competing maps including Schuberth (1967); Baskerville (1994); and Merguerian and Merguerian (2004) and Brock and Brock (2001) present highly differing tectonic and stratigraphic interpretations. Our chapter is not an attempt at re-mapping New York City but simply offers some overdue geochemical characterizations of Manhattan and Hartland formations that may be used to refine future mapping.

The status of New York City geology has only become more confusing during recent years. Currently there is no consensus on the age, the provenance of the sedimentary protoliths, or the formation identification of most schist outcrops in Manhattan despite good exposure and extensive studies by countless geologists. We will, therefore, be unable to provide a consensus answer to the most basic questions at some of the outcrops that we visit on this field trip, such as “What formation is this and how old is it?” However, we will be able to provide some of the first geochemical data on each of the outcrops we visit together with some geochemical based stratigraphic correlations and some geochemical constraints on the provenance of the sedimentary provenance.

The Status of NYC Mapping and Stratigraphy

It is beyond the scope of this report to contribute to complex stratigraphic issues or map detail. However, it may be instructive to point out a few of the current choices (Figure 1).

1. Hall (1968, 1976, 1980)

Hall (1968) and a series of papers published with co-authors subdivided the Manhattan Schist into member A (an autochthonous member) and members B and C (allochthonous members). Member A overlies the Inwood Marble unconformably and is a gray schist containing quartz, biotite, muscovite, garnet, and plagioclase. Near the base of member A some dolo-marble layers appear with diopside and phlogopite. Member C overthrusts member A and is thought by Hall (1968) to be Early to Middle Cambrian in age. Member C is gray schist containing biotite, muscovite, quartz, plagioclase, garnet, kyanite, sillimanite, tourmaline and magnetite with minor gneiss. Mamber B is amphibolite but is not continuous.

The Hartland Formation was mapped from southwestern Connecticut through Westchester County, New York by Hall (1968). He subsequently (Hall 1976) divided the Hartland into four members (an amphibolite member, a schist-gneiss-amphibolite member, a gray gneiss member, and a schist and granulite member). However these member designations are best applied to the White Plains New York to Glenville, Connecticut area and have not been applied elsewhere. Hall (1980) describes in some detail the overthrusting of Hartland Formation over Manhattan Schist.

2. Ratcliffe and Knowles (1969)

The Middle Ordovician age of the lower Manhattan Schist (member A) was determined by Ratcliffe and Knowles (1969) on the basis of pelmatozoan fragments they discovered in a meta-limestone interbedded with the base of the Manhattan Schist. They also correlate the lower Manhattan Schist with the Annsville Phyllite of New York, a thick meta-pellite that may also correlate with the Martinsburg shale of New Jersey.

3. Baskerville (1994)

The map of New York City created by Baskerville (1994), describes the Manhattan Schist (without members) as an allochthonous Lower Cambrian formation that was thrust over the autochthonous Walloomsac Formation (previously mapped as Manhattan A) along the Inwood Hill Thrust-fault. Baskerville (1994) also thrusts the Hartland Formation (a Middle Ordovician to Lower Cambrian) unit (with no members) over the Manhattan Schist along Cameron’s Line.

4. Brock and Brock (2001)

The map of New York City created by Brock and Brock (2001), is a major departure from previous mapping. It is based in part on a 570 Ma zircon date found in the Ned Mountain Formation that they interpret as forcing Manhattan Schist and Bronx Zoo-type Hartland Formation of “identical” chemical composition into the Late Neoproterozoic. This assumes that the host of the zircon was an ortho-amphibolite perhaps a meta-basalt and not a para-amphibolite originally composed of detritus eroded off of Late Neoproterozoic diabase exposed throughout the New Jersey Highlands and Reading Prong (Volkert and Puffer 1995). The Ned Mountain Formation includes the rocks exposed along the East River mapped by Baskerville as the Cambrian-Ordovician Ravenswood Granodiorite.

The Ned Mountain and Manhattan Schist are thrust over the Walloomsac Schist according to Brock and Brock (2001) and then Bronx Zoo type Hartland Formation is thrust over the Manhattan Schist. “True” Hartland Formation from an exotic (presumably arc) source is confined to schists associated with serpentinites (including the Staten Island Serpentinites and Castle Point Serpentinite) exposed along the Hudson River and to schists exposed well about nine Km east of Manhattan such as those of Pelham Bay.

5. Merguerian and Sanders (1994), Merguerian and Merguerian (2004)

Mapping and stratigraphic interpretations by Merguerian and Sanders (1994) and Merguerian and Merguerian (2004), agrees in most respects with earlier work by Hall. However they proposed that Manhattan Schist was thrust over Ordovician, autochthonous Walloomsac Schist instead of Ordovician, autochthonous Manhattan A. Then the Cambrian-Ordovician, allochthonous Hartland Formation was thrust over Cambrian-Ordovician, allochthonous Manhattan Schist along Cameron's Line deep into Manhattan through Central Park.

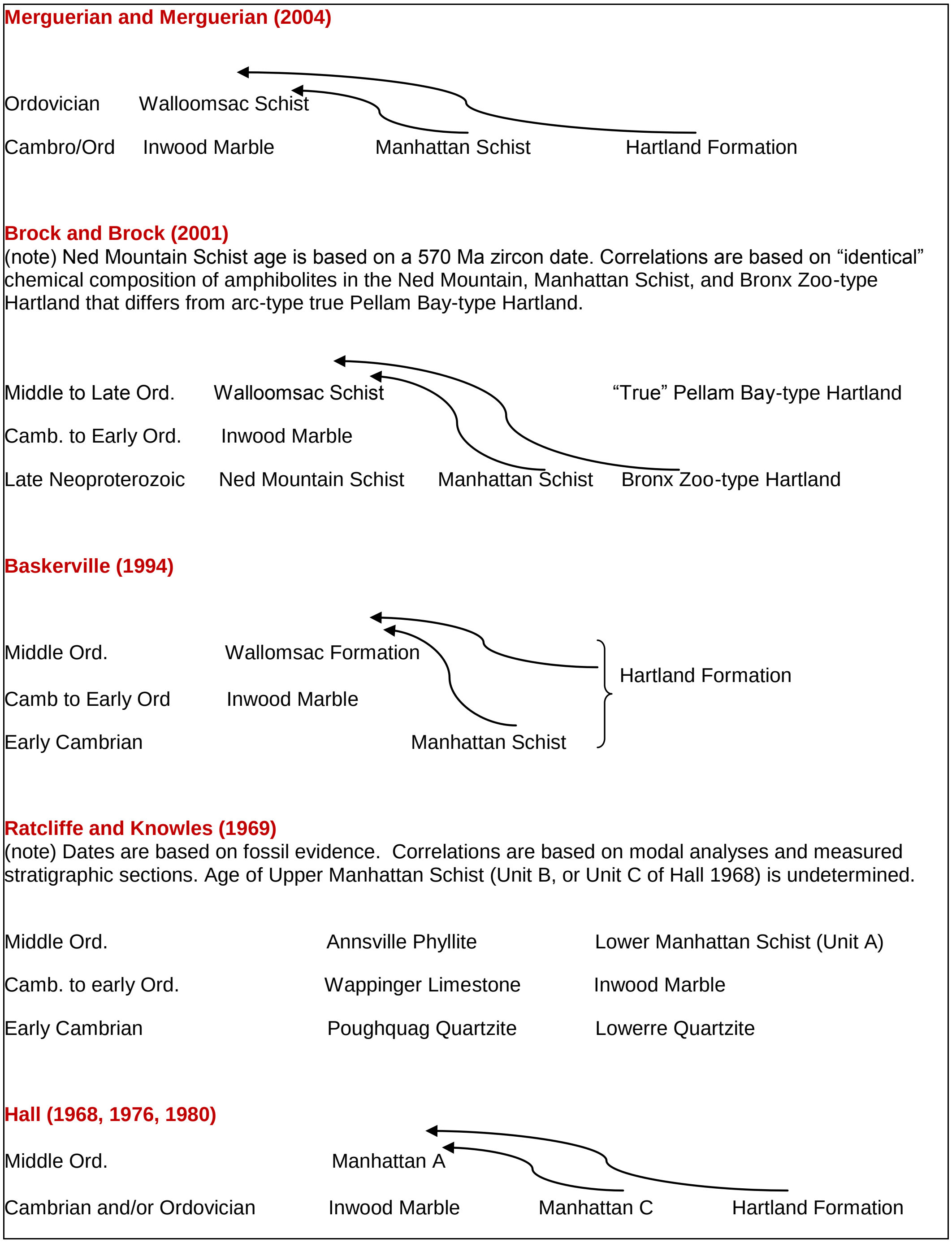


Figure 1. Stratigraphic relationships among New York City rock units according to various authors.

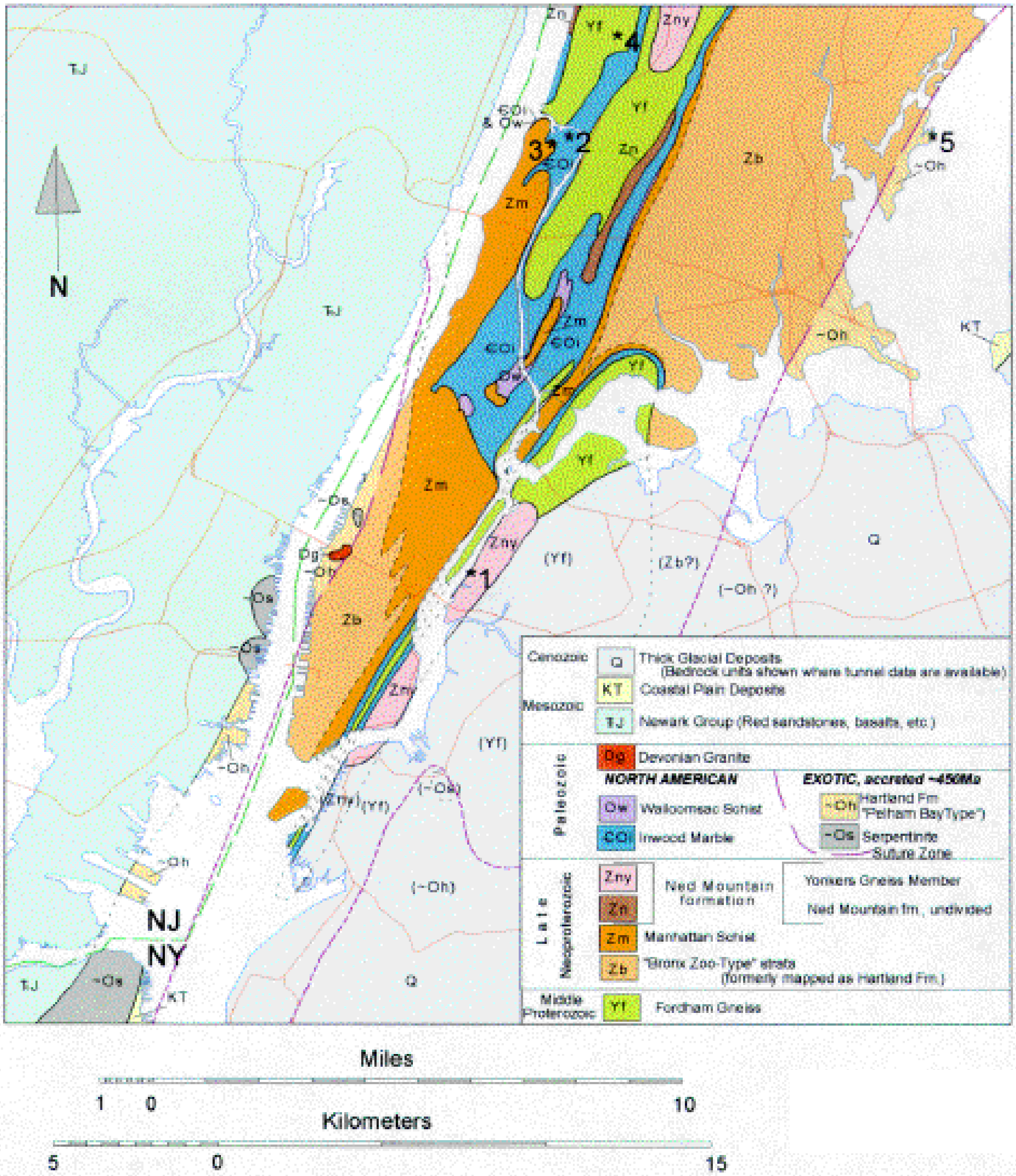


Figure 2. Geologic map of New York City (Brock and Brock, 2001)

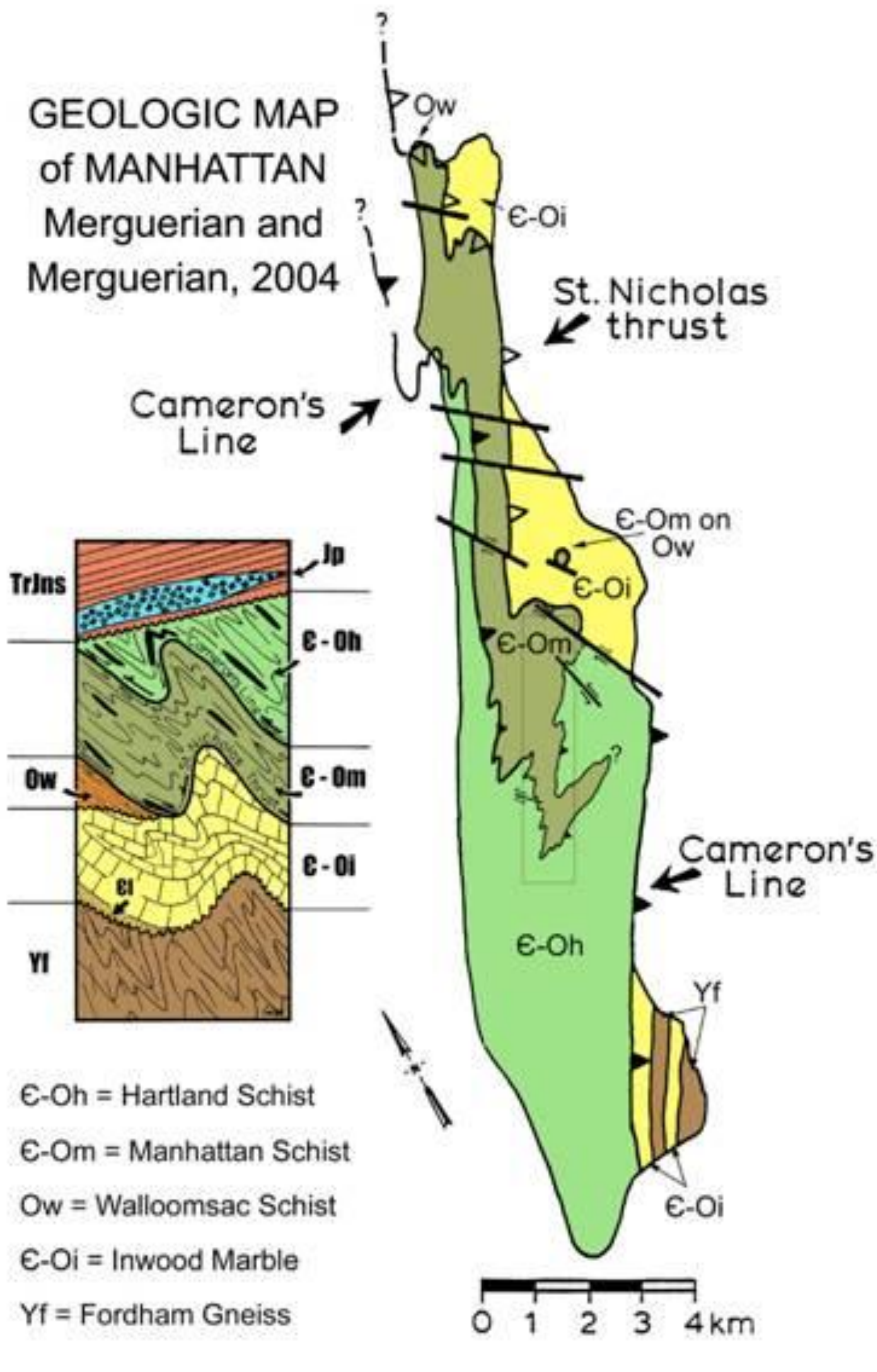


Figure 3. Geologic Map of Manhattan (Merguerian and Merguerian, 2004).

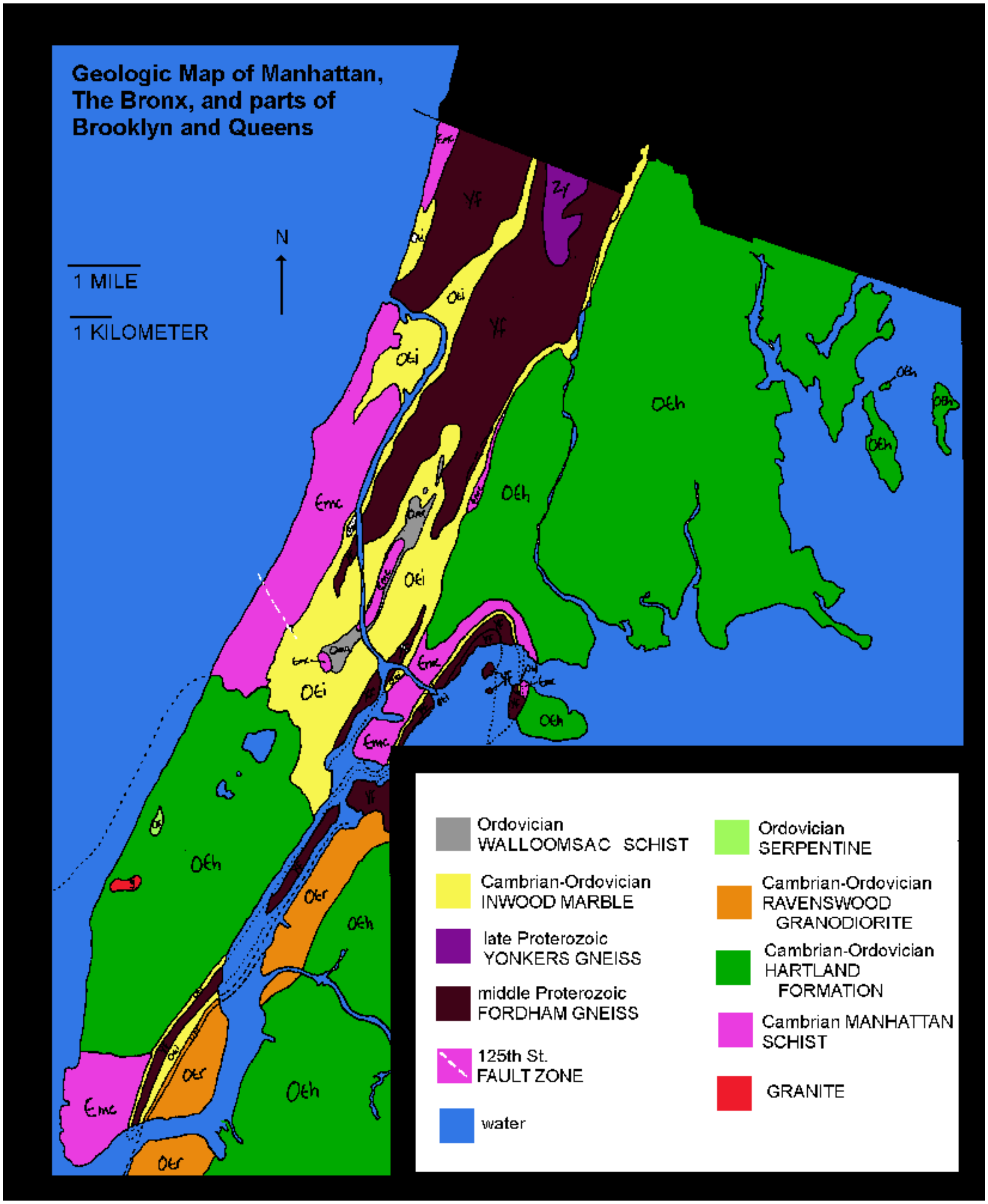


Figure 4. Geologic Map of Manhattan, The Bronx, and parts of Brooklyn and Queens by Baskerville (1994) as modified by the American Museum of Natural History (2010), New York City Geology, (Amnh.org).

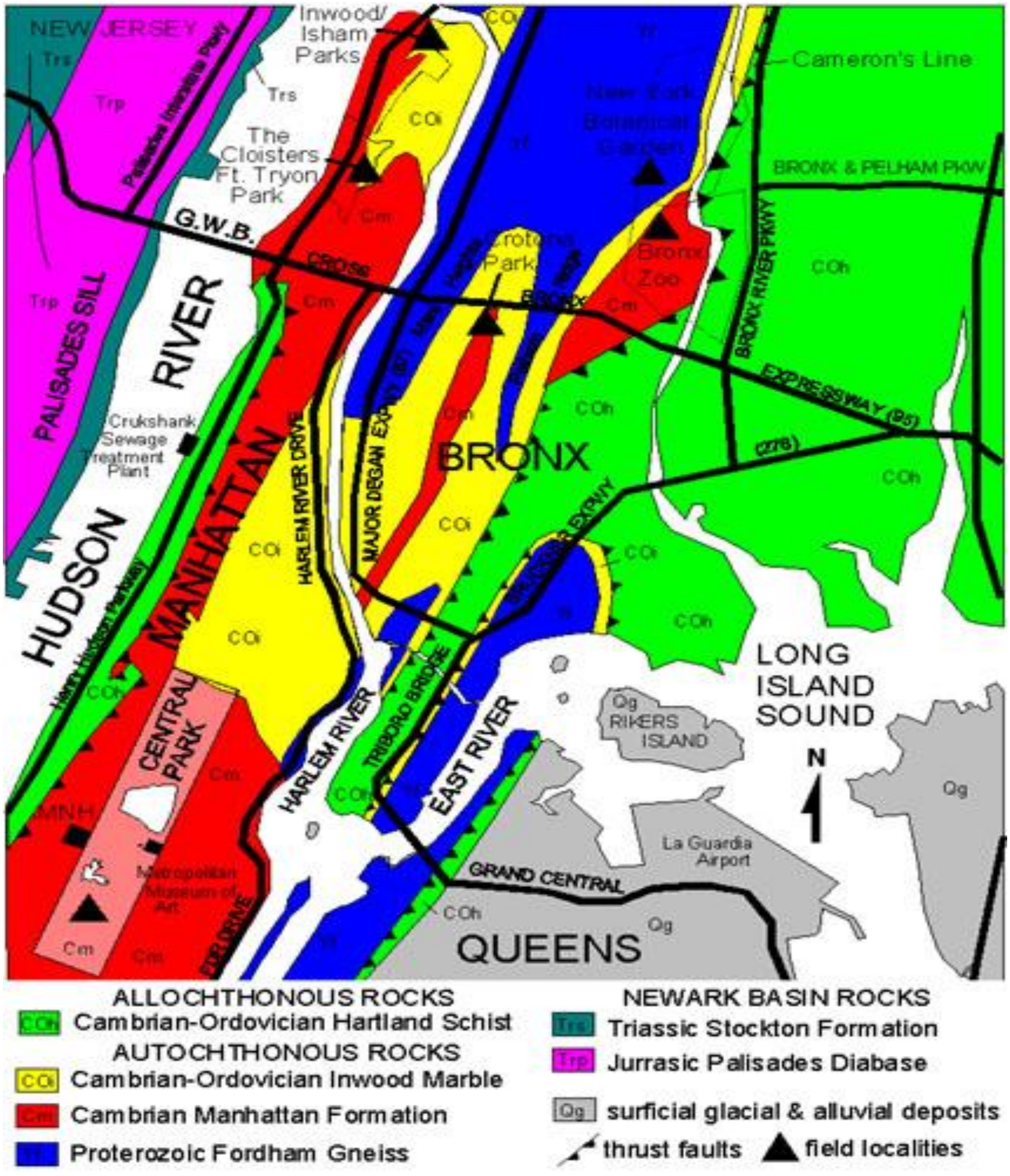


Figure 5. Geologic Map of northern New York City area by Schuberth (1967) as modified by the United States Geological Survey (2010) (3dparks.wr.usgs.gov).

Where is Cameron's Line ?

A recent Wikipedia definition of Cameron's Line is "… an Ordovician suture fault in the Northeast United States which formed as part of the continental collision known as the Taconic orogeny around 450 Mya. Named after Eugene F. Cameron, who first described it in the 1950's it ties together the North American continent, the prehistoric Taconic Island volcanic arc, and the bottom of the Iapetus Ocean." The NYC portion of the Northeast the "American continent" is the Manhattan Schist, the "volcanic arc" is the Hartland Formation and the "bottom of the Iapetus Ocean" is represented by a series of serpentinite units such as the Staten Island and Castle Point serpentinites . If the serpentinites of the NYC area were a continuous unbroken unit they would define Cameron's Line. However since this is not the case it is located at the boundary of the Manhattan Schist and the Hartland Formation along the stretches where serpentinite is missing.

One reason why it may be important to find Cameron's line is because as defined by Wikipedia it is a major fault and earthquakes typically occupy fault-plains. Since it appears that Cameron's line cuts through one of the most densely populated urban centers on earth it is all the more important. Strong local earthquakes are rare but not unheard of. It is somewhat likely that the focus of the next quake to strike NYC will be on Cameron's line although this is a subject that needs considerable additional study.

Petrological Characterization of Manhattan and Hartland Formations

A major obstacle faced by previous studies of New York City is the lack of any reliable petrologic criteria that can be used to distinguish the Hartland Formation from the Manhattan Schist in the field. Both formations are biotite-schists with similar accessory minerals including muscovite- garnet amphiboles, sillimanite, and magnetite. Some samples of Manhattan Schist also contain kyanite or tourmaline although these are not reliable or mappable characteristics. A few geologists including Merguerian and Merguerian (2004) have found weathered outcrop color useful. They find that the weathered surface of most outcrops of Manhattan Schist are reddish brown in contrast to the gray color of most Hartland weathered surfaces although exceptions are not rare.

Contrasting Provenances of Hartland and Manhattan Protoliths

There is a consensus among each of the principal geologists who have studied the New York City area that the protolith of the Manhattan Schist was pelitic sediment eroded off the North American continental craton and deposited on the continental shelf. The timing of metamorphic conversion into schist, however is in dispute. Bed-rock sources of sediment included widespread Proterozoic potassic granites, quartz-oligoclase gneiss including those of the New Jersey Highlands and shallow Neoproterozoic alkalic diabase intrusions and probable basaltic extrusions (Volkert and Puffer, 1995; Puffer, 2004)

In contrast, the Hartland Formation is generally interpreted as an exotic arc terrain composed of calc-alkaline volcanic rock and vocanogenic sediments that was accreted onto Laurentia during the Taconic Orogeny during the Ordovician period.

One objective of this project, therefore, is to examine the extent to which any geochemical distinction can be identified among sediments eroded off such contrasting sources.

Sampling and Analysis of New York City Schists

Sampling for this project was confined to schists, the chief component of NYC bedrock. Although amphibolites are also widespread throughout the NYC area they are petrogenically ambiguous and beyond the scope of this study. The MgO/CaO ratio of NYC schists is much higher than any common volcanic rock and are therefore is interpreted as meta-sediments. In contrast, NYC amphibolites resemble meta-diabase which may have intruded into the bedrock after it was deposited and resemble common meta-volcanic rocks which may have extruded from sources that differ from the sedimentary source of the schists.

Each schist sample was collected at a different large outcrop. Each sample, to the extent possible, represents the most common lithology of the outcrop from which it was sampled. Float samples were not collected.

Each sample collected from Boro Hall Park, and Central Park tables 1-2 was analyzed with a Rigaku x-ray fluorescence wavelength dispersive spectrophotometer. Samples from Riverside Park, the campus of The City College, and Pelham Bay were analyzed commercially at ALS Chemex using ICP-MS techniques. Care was taken to avoid damage to outcrops and wherever possible small loose samples were lifted from existing cracks then cleaned and trimmed of weathered material.

1. Central Park, Manhattan

Thirty two schist samples representing 32 of the largest outcrops in Central Park were collected and analyzed (Table 1). The analytical data appearing in Table 1 has not been previously published although averages were published by Puffer et al (1994) that included some amphibolite samples. As a test of Taterka's (1987) placement of Cameron's Line (the plate suture at the base of the Hartland Terrain) through the center of Central Park, 19 schist samples from the northern half of Central Park were collected north of his line placement and geochemically compared with 20 samples from the southern half. Puffer et al (1994). It was anticipated that if the schist samples collected from the northern half were Manhattan Schist they might display some geochemical distinction when compared to Hartland Schist samples from the southern half. Sample locations appear on the Map of Central Park by Puffer et al (1994).

Table 1, Chemical Composition of Schist Samples from Central Park

A. 18 samples from north of Taterka's placement of Cameron's Line (Manhattan Formation ?)

Sample #	CK-5	CK-8	CK-11	CK-13	CK-19	CK-20	CK-21	CK-22	CK-30	CK-32	CK-33	CK-34
SiO_2	70.08	63.41	68.42	61.98	59.91	58.07	66.34	53.08	62.41	52.19	55.95	53.27
TiO_2	0.99	1.22	0.81	1.2	1.24	1.18	1.03	1.47	1.28	1.49	1.4	1.79
Al_2O_3	12.77	16.14	13.56	15.08	16.39	20.21	14.11	20.34	16.69	21.36	18.09	16.53
FeO_t	5.51	7.61	5	8.1	7.19	7.73	5.07	8.06	6.87	8.92	7.29	10.81
MnO	0.09	0.1	0.08	0.15	0.07	0.14	0.05	0.12	0.13	0.19	0.09	0.19
MgO	2.92	3.35	2.51	3.86	3.78	3.91	2.56	4.29	3.44	4.2	4.01	4.35
CaO	1.58	1.13	1.89	1.85	1.31	1.29	1.01	0.93	1.63	1.12	1.99	2.01
Na_2O	2.68	1.52	3.26	2.13	1.62	2.78	1.28	1.74	2.34	2.01	1.92	1.89
K_2O	2.54	3.62	2.24	3.46	3.99	4.31	3.41	4.73	3.26	4.82	4.31	5.09
P_2O_5	0.14	0.24	0.12	0.25	0.23	0.13	0.18	0.14	0.17	0.1	0.18	0.22
LOI	1.32	1.7	1.98	3.2	3.35	1.42	3.56	4.5	1.76	2.6	3.42	2.97
Total	100.62	100.04	99.87	101.26	99.08	101.17	98.6	99.4	99.98	99	98.65	99.12
Cr	80	105	85	151	131	91	50	110	80	80	121	90
Cu	30	109	109	95	101	94		97	101			
Ni	31	59	62	42	64	51	15	60	40	50	49	43
Rb	97	112	89	114	118	119	83	114	96	120	110	112
Sc	9	14	9	16	12	10		15				
Sr	192	160	218	226	172	200	163	131	158	212	201	122
V	144	180	123	178	165	171	92	194	163	171	138	179
Y	26	25	24	24	26	25	26	26	24	21	25	22
Zr	319	248	263	243	225	203	344	199	268	244	183	206

Sample #	CK-37	Blockh	EMeadow	Red	3	2
SiO_2	62.36	66.5	64	49.43	47.01	75.72
TiO_2	1.09	1.1	1.12	1.38	1.78	0.66
Al_2O_3	18.8	14.9	12.95	20.01	19.42	10.9
FeO_t	6.21	4.99	6	10.79	11.58	3.72
MnO	0.09	0.08	0.08	0.11	0.14	0.08
MgO	3.21	2.2	3.04	3.21	4.63	1.5
CaO	0.99	1.31	1.92	0.17	1.01	1.32
Na_2O	1.24	2.12	1.92	0.31	1.62	2.25
K_2O	3.7	2.71	3.1	6.35	6.21	1.23
P_2O_5	0.23	0.19	0.17	0.07	0.34	0.12
LOI	1.8	2.61	2.96	5.68	5.29	2.18
Total	99.72	98.71	97.26	97.51	99.03	99.68
Cr	98	80	150	102	210	49
Cu	112					
Ni	40	18	20	15	60	37
Rb	107			153	148	77
Sc	14					
Sr	78	151	135		85	155
V	179	88	120	210	195	67
Y	27			26		
Zr	240	454	338	158		

B. 13 samples from south of Taterka's placement of Cameron's Line (Hartland Formation ?)

Sample #	CK-1	CK-2	A1	B1	B2	B3	C1	D1	G1	H	I2	Z1	7
SiO_2	57.22	73.94	61.11	55.85	47.85	60.21	61.42	58.35	68.4	61.92	65.31	53.22	56.83
TiO_2	1.73	1.18	1.12	1.17	1.48	0.95	1.35	1.14	0.79	1.04	0.99	1.36	1.41
Al_2O_3	14.88	10.69	19.86	19.8	23.01	17.85	17.11	18.99	14.52	16.99	16.32	18.41	17.61
FeO_t	8.21	4.49	6.07	7.91	9.08	6.28	8.44	8.11	4.71	7.03	5.82	8.29	7.71
MnO	0.13	0.08	0.08	0.11	0.11	0.14	0.13	0.12	0.08	0.12	0.09	0.12	0.1
MgO	4.88	2.35	2.99	5.56	3.92	4.41	3.33	1.19	1.97	4.42	2.79	4.38	4.2
CaO	2.21	1.82	1.97	0.52	1.29	0.3	0.61	0.62	1.02	0.22	1.18	2.09	1.03
Na_2O	2.47	2.63	2.52	1.01	1.73	0.48	0.96	1.45	1.24	0.68	1.87	2.27	1.25
K_2O	3.85	1.42	2.99	5.94	6.69	5.34	5.37	5.63	3.61	5.54	3.86	4.46	5.09
P_2O_5	0.4	0.21	0.22	0.13	0.23	0.18	0.22	0.11	0.09	0.07	0.12	0.23	0.27
LOI	3.7	1.5	1.48	2.88	3.21	2.73	1.39	3.32	3.21	1.94	2.9	4.39	3.97
Total	99.68	100.31	100.41	100.88	98.6	98.87	100.33	99.03	99.64	99.97	101.25	99.22	99.47
Cr	145	149	85	195	110	90	95	52	45	89	80	83	133
Cu	92	107	106	90	87		92	90		94	107		
Ni	50	30	39	63	43	60	42	32	36	71	37	54	51
Rb	102	65	99	179	158	177	131	173	113	174	123	131	150
Sc	16	8	15	15	16	14	14	14		13	14		
Sr	187	107	171	42	132	10	22	72	100	2	129	123	109
V	230	75	169	177	227	112	207	171	85	124	172	160	150
Y	24	24	25	28	26	26	26	25	25	25	27	26	
Zr	268	456	230	178	185	139	163	213	278	163	269	223	

2. Boro Hill Park, Manhattan

Fifteen schist samples from Boro Hall Park, at locations that appear on a map by Cadmus et al. (1996), were collected and chemically analyzed (Table 2). Preliminary analyses of the same samples plus some amphibolite analyses appear on Table 1 of Cadmus et al (1996). Samples M-5 through M-9 and H-1 through H-4 were collected along Third Avenue road cuts at the southern edge of the park at locations spaced at least 10 m apart. As a test of Baskerville's (1989) placement of Cameron's Line through the center of Boro Hill, 8 schist samples from the southwestern portion of Boro Hill Park were collected west of his line placement and geochemically compared with 7 samples from the east of his placement of Cameron's Line. It was anticipated that if the schist samples collected from the western half were Manhattan Schist they might display some geochemical distinction when compared to Hartland Schist samples from the eastern half. Baskerville (1989) proposed that the swale separating the east side of the park from the west side defines Cameron's Line.

Table 2, Chemical Composition of Schist Samples from Boro Hall Park

A. Boro Hall west of Baskerville's placement of Cameron's Line (Manhattan Formation?)

Sample #	MS-1	MS-3	MS-4	MS-6	MS-7	M-7	M-8	M-9
SiO_2	67.31	61.24	53.86	61.72	55.72	60.16	58.84	58.32
TiO_2	0.87	0.68	1.09	1.2	1.51	0.9	1.09	1.02
Al_2O_3	12.48	15.68	16.46	19.25	16.8	16.4	17.02	17.87
FeO_t	5.32	7.01	7.03	6.26	7.78	6.25	7	7.02
MnO	0.05	0.08	0.07	0.06	0.1	0.1	0.04	0.06
MgO	1.81	1.73	2.62	1.72	2.45	3.72	4.34	4.02
CaO	3.02	4.82	5.92	0.85	1.24	4.16	3.9	3.95
Na_2O	1.18	0.88	1	0.49	1.34	0.95	0.6	0.65
K_2O	2.33	1.8	4.33	4.17	5.11	4.26	3.28	3.32
P_2O_5	0.2	0.19	0.21	0.23	0.27	0.22	0.22	0.23
LOI	5.21	5.11	4.92	4.85	5.83	3.42	4.01	4.32
Total	99.78	99.53	97.51	100.8	98.15	100.54	100.34	100.78
ppm								
Ni	16	18	15	15	18	67	66	65
Rb	133	148	160	148		89	83	78
Sr	630	85	622	320		760	219	358
Zr	257	188	177	256		216	208	202

B. Boro Hall Park east of Baskerville's placement of Camerons Line (Hartland?)

Sample #	H-1	H-2	H-3	H-4	Hart-1	Hart-3	Hart-5
SiO_2	51.45	62.51	40.32	62.06	54.61	58.75	58.95
TiO_2	1.62	1.61	2.25	1.63	1.74	1.29	1.93
Al_2O_3	16.43	13.83	24.04	15.77	19.48	19.82	15.07
FeO_t	9.92	8.48	13.4	9.13	10.04	8.67	8.67
MnO	0.1	0.14	0.15	0.17	0.16	0.13	0.14
MgO	5.52	4.03	5.74	3.65	2.38	2.1	3.07
CaO	2.5	1.42	0.75	1.04	0.82	0.79	0.36
Na_2O	2.36	1.91	0.91	1.6	0.58	0.46	0.33
K_2O	4.53	3.92	6.58	3.49	4.89	3.45	6.19
P_2O_5	0.31	0.12	0.32	0.25	0.19	0.2	0.22
LOI	5.02	2.64	5.31	2.11	4.52	4.22	4.75
Total	99.76	100.61	99.77	100.9	99.41	99.88	99.67
Ni	73	65	79	73	17	18	20

Rb	92	86	104	81	145	128	176
Sr	328	277	132	192	223	221	175
Zr	146	314	232	358	287	180	219

3. West-side, Manhattan, along Hudson River

Thirteen schist samples were collected along the west side of Manhattan of which 5 were chemically analyzed (Table 3). From south to north analyzed samples were collected at a city park on 53rd Street and 11th Ave (Sample H3), at a large outcrop just north east of the Boat Basin near 82nd Street and 11th Ave, at another large outcrop in Riverside Park at 91st Street along the Hudson River, and on the campus of The City College at 135th Street. Of these samples only the sample from The City College has been consistently mapped as Manhattan Schist. Each of the remaining west-side samples are highly controversial.

Table 3, Chemical Composition of Schist Samples from the West Side of Manhattan

location	82nd & 11	91st & H	91st & H	53rd & 11	City College
Sample #	H1D	H2A	H2C	H3	M1
SiO2	61.92	83.51	46.29	63.76	58.13
TiO2	0.96	0.62	1.51	0.81	0.95
Al2O3	18.9	7.76	23.98	15.4	18.52
FeOt	7.87	2.81	13.68	6.42	7.76
MnO	0.1	0.05	0.17	0.11	0.22
MgO	1.81	0.59	3.36	3.33	3.26
CaO	0.59	0.96	0.35	2.61	3.46
Na2O	0.81	1.46	0.62	3	2.47
K2O	4.61	1.24	6.92	3.02	3.21
P2O5	0.13	0.06	0.05	0.27	0.12
LOI	1.8	0.56	2.71	0.81	1.67
Total	99.5	99.62	99.64	99.54	99.77
Ba	0.1	0.03	0.12	0.05	0.08
Cr	79	26	118	42	99
Sr	119	125	92	246	292
Zr	269	483	282	215	194

4. Pelham Bay, Bronx

In order to collect schist samples from a New York City location that has been consistently mapped as Hartland Formation we chose Pelham Bay. Similar reasoning was provided by Brock and Brock (2001) who identified the schist there as "True Hartland Formation" and as Pelham Bay-type Hartland Formation. Seven schist samples were collected there at large representative outcrops of which four were chemically analyzed (Table 4). Each of these sample locations were mapped by Seyfert and Leveson (1968) as part of the "Felsic Unit" which represents about 85% of the exposed rock of the Bay. Some of the Felsic Unit is described as felsic gneiss but the samples that we collected each contain at least 20 volume percent biotite and display a schistose texture. Table 4 also includes the average chemical composition of 2 "sillimanite schists" and the average of 2 "plagioclase-biotite gneisses" (interpreted here as schists with 35 % biotite) chemically analyzed by Seyfert and Leveson (1968).

Additional samples of schist and gneiss from the Brooklyn – Queens water tunnel complex were also sampled by Jeff Steiner and chemically analyzed (Table 4).

Table 4, Chemical Composition of Schist Samples from Pelham Bay Park and Brooklyn

	Seyfert and Leveson (1968)		Pelham Bay Park				Brooklyn Tunnel Complex	
			new	new	new	new	gneiss	schist
Sample #	4	14	OB-1	OB-3	OB-4	OB-5	b12	b13
SiO2	48.1	50.7	65.2	54.8	69.4	68.5	73.03	55.93
TiO2	1.7	1.6	0.87	1.2	0.79	0.94	0.19	1.42
Al2O3	23.1	20.3	14.15	19.1	13.3	14.1	11.99	14.58
Fe2O3	2.6	2.7	7.55	9.21	4.79	6.43	4	12.11
FeO	9.5	6.8						
MnO			0.4	0.17	0.07	0.14	0.09	0.06
MgO	4.4	3.5	2.07	3.07	1.57	1.23	1.89	5.01
CaO	2	4	1.07	2	2.39	3.4	0.79	1.89
Na2O	2.3	5.1	2.67	2.19	2.82	2.65	1.55	2.88
K2O	4.5	3.6	4.54	4.9	1.96	1.29	6.23	3.44
P2O5	n.d.	n.d.	0.1	0.14	0.16	0.39	0.02	0.09
LOI	n.d.	n.d.	0.39	0.79	0.68	0.2	0.5	1.52
Total	98.2	98.3	99.01	97.57	97.93	99.27	100.28	98.93
Ba			623	949	445	342	1158	524
Cr			60	70	40	50	0	275
Cu			7		17	16	11	71
Ni			39	37	10	14	0	100
Rb			163	151	62	48	102	120
Sr			111	106	97	159	75	129
V			69	97	64	79	0	305
Y			27	42	29	37	226	52
Zr			179	169	365	356	617	220

Bktl 12 run 45 BTL 13 run

Geochemical Results:

1. Geochemical resemblance of schist samples from Manhattan to an active continental margin

About half of the samples collected throughout Manhattan were collected at locations mapped by various authors as Hartland Formation and about half were collected at locations mapped as Manhattan Formation. However, anticipated bimodal distributions are not apparent on any plot of element distributions and both sample populations display overlapping distributions of all elements analyzed. In particular, plots of the K_2O/Na_2O vs. SiO_2 composition of schist samples from Manhattan (Tables 1, and 3) onto Figure 6 define a single population within the "active continental margin" field of Roser and Korsch (1986). Sedimentary rock collections from well defined tectonic settings on a global basis were plotted onto a K_2O/Na_2O vs. SiO_2 discrimination diagram and successfully separated by Roser and Korsch (1986) into 3 fields with minimal overlap. The "passive continental margin" tectonic setting is described as sediment deposited in plate interiors at stable continental margins or intracratonic basins. Sediment sources are dominated by recycled quartz-rich sediment derived from adjacent continental terrains. Sediment deposited into an "active continental margin" setting is described as derived from a tectonically active continental margin on or adjacent to active plate boundaries. Sediment are dominated by quartzo-feldspathic continental derived trench deposits deposited into an accretionary wedge or complex active margin basins. The "oceanic island arc" field of Figure 6 represents quartz poor volcanogenic sediments derived from oceanic island arcs and deposited in a variety of tectonic settings including forearc, intraarc, and backarc basins and trenches (Roser and Korsch, 1986).

An active continental margin setting is consistent with several tectonic models that describe sedimentation preceding the Taconic Orogeny including the "Cross Sections of Eastern North America" (USGS, 2003). An active continental margin setting is also consistent with the consensus description of Manhattan Schist sedimentation.

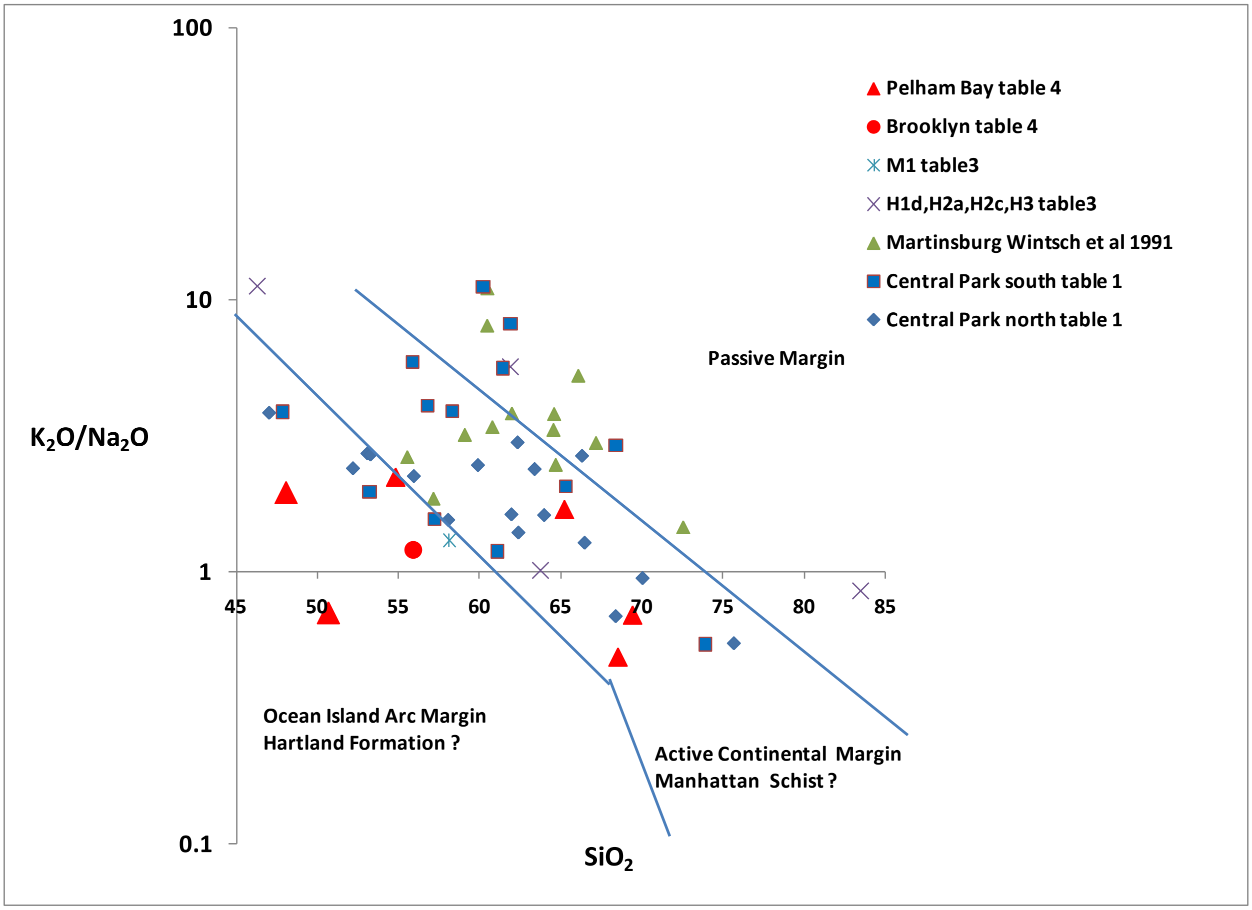


Figure 6. Plot of schist samples from Manhattan (tables 1-4) plotted onto diagram developed by Roser and Korsch (1986).

2. Geochemical resemblance of schist samples from Manhattan to Martinsburg shale

The Martinsburg Formation (Group) represents a second met sediment that was probably deposited in an active continental margin setting. McBride (1962) on the basis of his interpretation of sedimentary structures typical of the Martinsburg Formation suggested a turbidity current depositional setting and interpreted the chemical composition of the Martinsburg as indicative of a sedimentary to low-grade metamorphic rock provenance with granitic rocks as a secondary source. He also interpreted the turbidity currents as having flowed down the sub-sea slope of Appalachia toward the southeast.

The geochemical data of Wintch et al (1991) can be used to compare the Martinsburg with the schists of Manhattan. They collected 48 mudstones and 26 greywacke samples from a 3500 m thick section through the Martinsburg at Lehigh Gap Pennsylvania, 65 km west of Manhattan. They found that during greenschist-facies metamorphism of greywacke/metagraywacke assemblages Na_2O was lost and K_2O was gained. They also found that these changes were balanced by opposite changes in adjacent mudstone/slate assemblages resulting in minimal formation-wide net changes in either component. The K2O/Na2O and SiO2 data of Wintsch et al (1991) is plotted onto Fig. 6 and overlaps the field of schist from Manhattan. Figure 6, therefore, indicates that the schists of Manhattan and the metasediments of the Martinsburg Formation were probably deposited in a similar depositional setting and supports a correlation of two very thick Ordovician metasediment formations exposed 65 km from each other on the opposite sides of the Newark rift basin.

3. Geochemical resemblance of schist samples from Pelham Bay and Queens New York to volcanogenic arc detritus

In contrast to the schists of Manhattan, a schist sample collected from Brooklyn/Queens and the schist samples collected at Pelham Bay Park (including those chemically analyzed by Seyfert and Leveson, 1968) plot within or close to the “oceanic island arc” field of Figure 6. Although there is some overlap into the “active continental margin” field this result is consistent with the consensus view that the metasediments of Pelham Bay Park were derived from an island arc source.

More geochemical data is clearly needed but it appears that the Roser and Korsch (1986) diagram or similar as yet to be developed criteria may provide a basis for distinguishing between Manhattan Schist and Hartland Formation.

Field Trip Stops:

Stop 1. Central Park

No rock collecting is permitted so don’t even bother to bring your rock hammer. However feel free to pick up loose samples and examine them with your hand lens. We will spend one hour at this stop. The bus cannot wait for stragglier.

The outcrops near 59^{th} Street. along the southern portion of Central Park have been described as Cambrian Manhattan Formation by Schuberth (1967), (Fig. 1); as Cambrian-Ordovician Hartland Formation by Baskerville (1994), Fig. 2); as Cambrian-Ordovician Hartland Schist by Merguerian and Merguerian (2004), (Fig. 3); and as Late Proterozoic Manhattan Schist by Brock and Brock (2001), (Fig. 4). Still other options have been published and we encourage lively debate, but please leave your rock hammers in the bus.

Most of the outcrops of Central Park are roche moutonnée that were cut by a thick ice sheet into bedrock during the Pleistocene. Glacial polish and striations are clearly visible.

The large outcrop near 59^{th} Street is a biotite-muscovite-garnet-plagioclase-sillimanite schist that is intersected by common pegmatites and amphibolite layers that are approximately conformable to the rock foliation.

Please note the mineralogy and texture of this rock because we will compare it with outcrops from further north into Central Park and elsewhere on the trip. But please be mindful of the time and try not to wander too far north into the park.

Stop 2. Riverside Park (Boat Basin)

The outcrops near 82nd Street just north east of the Boat Basin have been described as Cambrian Hartland Formation by Schuberth (1967), (Fig. 1); as Cambrian-Ordovician Hartland Formation by Baskerville (1994), Fig. 2); as Cambrian-Ordovician Hartland Schist by Merguerian and Merguerian (2004), (Fig. 3); and as Ordovician Pelham Bay-Type Hartland Formation by Brock and Brock (2001), (Fig. 4). The exact outcrop location appears on the map by Baskerville (1994) and includes strike and dip data as do all of his outcrop locations.

Stop 3. City College Campus and St. Nicholas Park (lunch)

Box lunches will be provided for all that have ordered them. While you are eating please examine any of the excellent exposures of Manhattan Schist throughout St. Nicholas Park adjacent to the campus of The City College but please don't wander too far off.

There is a general consensus that the outcrops exposed on the campus of the City University of New York (The City College) and along the St. Nicholas Park hill-side adjacent to the campus are Manhattan Schist (Figs 1-4). Samples observed here can, therefore, be thought of as a Manhattan Schist standard for comparison with other field trip stops.

St. Nicholas Avenue at the base of St. Nicholas Park marks the boundary between the Manhattan Schist on the west and the Inwood Marble on the East. The topography marking the highlands, locally Sugar Hill, has long been considered the result of differential erosion that sculpted the softer marble unit at the base. However, the downward-stepping outcrops that can easily be traced in the Park may represent an imbricate structural system that down-drops the schistose unit eastward.

Stop 4. Pelham Bay (Bronx, NY) Park in Beach Parking Lot , walk to north end of the beach and outcrops are at shoreline at Latitude 40.870095°; Longitude -73.783833°

The structural geology and petrology of Pelham Bay Park has been described by Seyfert and Leveson (1968). They used the New York State geological map by Fisher et al (1961) as their source of stratigraphy which had designated the Pelham Bay rocks as "…undivided schists and gneisses of unknown age". Seyfert and Leveson (1968) divided the Pelham Bay rocks into a "Felsic Unit" consisting of felsic quartz-plagioclase-biotite-gneisses and biotite-sillimanitic schists and a "Mafic Unit" consisting of amphibolites, plagioclase-biotite gneiss and minor calcite-rich layers and plagioclase-rich layers. They also describe several pegmatites within both units.

Some of this wide range of lithologies including pegmatites, pegmatite border zones, and the unusual calcite-rich layers were chemically analyzed with accompanying mineral modes (Seyfert, 1968). However, about 85% of their map of Pelham Bay Park consists of the "Felsic Unit". It is difficult to determine which lithology typifies the Felsic Unit but seems to fall somewhere close to an undefined boundary between schist and gneiss In general the Felsic Unit is largely a biotite- plagioclase-quartz-sillimanite-muscovite-garnet rock that mineralogically overlaps most of the schists of Manhattan. Microcline is notably absent or only a minor component of most of the rock but is a major component of some of the pegmatites. The microcline rich pegmatites appear to be discordant intrusions that intersect the foliation of the host rocks. A plagioclase rich generation of pegmatites is approximately equally common but is generally concordant to the foliation of host rocks that are also plagioclase rich and depleted in microcline.

Chemical analyses of the felsic unit members including plagioclase-biotite gneiss containing 64 % plagioclase and 35 % biotite and sillimanite schist containing 27 % plagioclase 43 % biotite and 11.5 % sillimanite (Seyfert, 1968) appear in Table 4.

Most of the exposed rock of Pelham Bay is a lighter shade of gray and not as reddish compared to Manhattan Schist. However, we find that fresh broken surfaces are difficult to distinguish from Manhattan Schist and are equally difficult to distinguish petrographically in thin section. We therefore invite all field trippers to carefully compare Pelham Bay rock with Manhattan rock. If any of you find any consistent mappable distinctions please share them with the rest of us.

One possible distinction may be contrasting plagioclase/biotite ratios. In general the plagioclase/biotite ratio of the average Pelham Bay rock is higher than typical of the schists of Manhattan. This elevated ratio is reflected in the contrasting (although overlapping) Na/K ratio of the analyzed rocks (Fig. 6). The plagioclase/biotite ratio may, therefore, be one of the few ways to map Hartland/Manhattan boundaries although, again, overlapping values do not permit definitive conclusions. As previously stated, sodic enrichment is an important characteristic of calc-alkaline lithologies such as the andesitic volcanic-arc lithology proposed for the protolith of the Hartland Formation.

REFERENCES

American Museum of Natural History, 2010, New York City Geology, (Amnh.org).

Baskerville. C. A., 1994, Bedrock and engineering geology maps of New York County and parts of Kings and Queens Counties, New York and parts of Bergen and Hudson counties, New Jersey: U.S. Geological Survey Miscellaneous Investigations Series Map I-2306 (2 sheets; colored maps on scale of 1/24,000)

Berkey, C. P., 1910, Areal and structural geology of southern Manhattan Island: New York Academy of Sciences Annals, v. 19, no. 11, part 2, p. 247-282.

Berkey, C. P., 1933, Engineering geology of the City of New York, p. 77-123 *in* Berkey, C. P., *ed*., Guidebook 9, New York Excursions, New York City and vicinity: International Geological Congress, 16th, United States, 1933, Washington, D. C., United States Government Printing Office, 151 p.

Britton, N. L., 1881, On the geology of Richmond County, New York: New York Academy of Sciences Annals, v. 2, p. 161-182 (map); abstract, Columbia School of Mines Quarterly, v. 2, p. 165-173.
Cozzens, Issachar, 1843, A geological history of Manhattan or New York Island: New York, NY, 114 p. (includes map).

Brock, P. J.C., and Brock, P. W.G., 2001, Bedrock geology of New York City: More than 600 m.y. of geologic history: http://pbisotopes.ess.sunysb.edu/reports/NYC/index.html,11 p.

Cadmus, D.; Hodgson, R.; Gatto, L. M.; and Puffer, J. H., 1996, Geochemical Traverse across Cameron's Line, Boro Hall Park, Bronx, New York, in Hanson, G.N., (editor), Geology of Long Island and Metropolitan New York, Long Island Geologists, SUNY Stony Brook, p. 26-32.

Eicher, J. C.; Ageitos, D. M.; Seeber, F. C.; Kaplan, C. H.; Buehler, S. C., and Puffer, J. H., l994, Geochemistry of the amphibolites of Central Park, New York City: in Hanson, G.N., (editor), Geology of Long Island and Metropolitan New York, Long Island Geologists, SUNY Stony Brook, p. 44-45.

Emerson, B. K., 1898, Geology of Old Hampshire County Massachusetts, comprising Franklin, Hampshire, and Hampden Counties: United States Geological Survey Monograph 29, 790 p.

Fisher, D. W., Isachsen, I. W., and Rickard, L. V., *editors and compilers*, 1970, Geological map of New York: New York State Museum and Science Service Map and Chart Series Number 15, scale 1:250,000.

Gratacap, L. P., 1887, Serpentine rock of Staten Island, N.Y.: Staten Island Natural Science Association Proceedings, v. 1, p. 55.

Gratacap, L. P. 1909. Geology of the City of New York. Henry Holt and Company, New York. 232pp.

Hall. L. M., 1968, Times and origin and deformation of bedrock in the Manhattan prong: in Zen, E-an, White, W.S. Hadley, J.B., and Thompson, J.B., eds., Studies of Appalachian geology: northern and maritime: Interscience Publishers, New York, p. 117-127.

Hall, L. M., 1976, Preliminary correlation of rocks in southwestern Connecticut: Geological Society of America Memoir 148, p. 337-349.

Hall, L. M. 1980, Basement-cover relations in western Connecticut and southeastern New York: in Wones, D.R.., ed., Proceedings: The Caledonides in the U.S.A., Memoir 2, p. 229-306.

Isachsen, Y. W., 1980, Continental collisions and ancient volcanoes: the geology of southeastern New York: New York State Geological Survey, Museum and Science Service, Educational Leaflet 24, Albany, New York, 13 p.

Julien, A. A. 1903. Genesis of the amphibole schists and serpentinites of Manhattan Island, New York. Bull. Geo. Soc. Amer. 14: 421-494.

Mather, W. W., 1843, Geology of New York. Part I. Comprising the geology of the First Geological District: Albany, NY, Carroll & Cook, Printers to the Assembly, 653 p., 46 pl.

Mcbride, E. F., 1962, Flysch and associated beds of the Martinsburg Formation (Ordovician), Central Appalachians, Journal of Sedimentary Research v. 32, p.

Merguerian, Charles, 1983, Tectonic significance of Cameron's Line in the vicinity of the Hodges Complex--an imbricate thrust model for western Connecticut: American Journal of Science, v. 283, p. 341-368.

Merguerian,, Charles. and Sanders, John E. 1994, Staten Island and Vicinity, Section of Geological Sciences 1994 Trips On-The Rocks Trip 33, The New York Academy of Sciences, 2 East 63rd Street, New York, New York, 152 pp.

Merguerian, Charles, and Merguerian, Mickey, 2004, Geology of Central Park -- From Rocks to Ice : *in* Hanson, G. N., *chm*., Eleventh Annual Conference on Geology of Long Island and Metropolitan New York, State University of New York at Stony Brook, NY, Long Island Geologist Program with Abstracts, 24 p.

Merguerian, Charles; and Moss, Cheryl, J., 2005, Newly discovered ophiolite scrap in the Hartland Formation of midtown Manhattan: in Hanson, G. N., chm., Twelfth Annual Conference on Geology of Long Island and Metropolitan New York, 16 April 2005, State University of New York at Stony Brook, NY, Long Island Geologists Program with Abstracts, 7 p.

Merrill, F. J. H., 1898, The origin of serpentinites in the vicinity of New York: New York State Museum Annual Report 50, v. 1, p. 32-44.

Merrill, F. J. H.; Darton, N. H.; Hollick, Arthur; Salisbury, R. D.; Dodge, R. E.; Willis, Bailey; and Pressey, H. A., 1902, Description of the New York City district: United States Geological Survey

Geologic Atlas of the United States, New York City Folio, No. 83, 19 p. (Includes colored geologic map on a scale of 1:62,500.)

Puffer, J. H.; Ageitos, D. M.; Buehler, S. C.; Eicher, J. C.; Kaplan, C. H.; and Seeber, F. C., l994, Geochemistry of the schists of Central Park, New York City: in Hanson, G.N., (editor), Geology of Long Island and Metropolitan New York, Long Island Geologists, SUNY Stony Brook, New York, p 86-92.

Puffer, J. H., 2004, Contrasting geochemistry, tectonic settings, and environmental impacts of Neoproterozoic, Paleozoic, and Mesozoic dike swarms of Northern New Jersey and their correlatives: *in* Puffer, J. H. and Volkert R. A. (editors), Neoproterozoic, Paleozoic, and Mesozoic Intrusive Rocks of Northern New Jersey and Southeastern New York, Geologic Association of New Jersey Field Guide and Proceedings, v. 21, p. 6-26.

Schnock, E. M., 1999, Construction of the Brooklyn Tunnel: p. 91-100 *in* Mega Projects – Means, methods and other construction issues, American Society of Civil Engineers, Metropolitan Section, Annual Seminar Proceedings, Cooper Union, NY, February 1999, 140 p.

Schuberth, Christopher, 1968, Geology of New York City and Environs, Garden City, New York; Natural History Press.

Seyfert, C. K., and Leveson, D. J., 1968 Structure and petrology of Pelham Bay Park, Trip G., New York State Geological Association Guidebook to Field Excursions, 40th Annual Meeting at Queens College, City University of New York, Flushing New York, p. 175-195.

Taterka . B.D., 1987, Bedrock geology of Central Park, New York City (M.S. Thesis): Contribution 61, Department of Geology and Geography, Univ. of Mass., Amherst Mass., 84 p.

United States Geological Survey (USGS), 2010. Northern Manhattan Island. http://3dparks.wr.usgs.gov/nyc/highlands/manhattan.htm

Volkert, R.A. and Puffer, J.H., l995, Late Proterozoic diabase dikes of the New Jersey Highlands – A remnant of Iapetan rifting in the North-Central Appalachians: United States Geological Survey Professional Paper, 1565-A, 22 pp.

Results (or Consequences) of Remediation at Lemon and Mill Creek Sites, Staten Island, New York

Peter G. Crowley, William G. Wallace, Department of Biology, College of Staten Island, 2800 Victory Boulevard, Staten Island, NY 10314

Introduction:

Lemon Creek and Mill Creek are two sites impacted by metals contamination. (See Figure 1.0) Both locations share common history in that they were active industrial sites for over one hundred years. Manufacturing operations ended at both locations and the buildings demolished. Development plans for the sites are still pending.

The S.S. White Dental Supply Company factory was located near Lemon Creek on the southern shore of Staten Island facing Raritan Bay. This facility manufactured dental instruments and amalgams from 1937 to 1986. Suspected contaminants include copper, lead, zinc, and mercury.

Mill Creek (AKA Nassau Creek) is a small tidally flushed creek located 3 km from the opening of the Raritan Bay. It is located adjacent to a facility that was previously used to recover metals from old telephone transmission cables. This facility closely resembled a smelting operation. The Tottenville Copper Company was located adjacent to Mill Creek in 1900. By 1941, the plant was operating as the Nassau Smelting and Refining Company. In 1971, it became the Nassau Recycling Corp where copper, lead and zinc were recovered from old telephone cables. This creek received effluent from the smelting facility for approximately 40 years (the mid 1930's through the late 1970's). As a result, the site received a Superfund designation for contamination. (EPA ID: NYD086225596)

The Mill Creek site is located on the western shore of Staten Island facing New Jersey. Mill Creek is located along the Arthur Kill Waterways known for frequent oil discharges by passing vessels.

For more than a century, chronic pollution from industrial facilities along the Arthur Kill has severely degraded its' water quality (Crawford, etal., 1994). Furthermore, anthropogenic pollution has caused the fertile marshlands to deteriorate. Due to the potential toxicity and transference up the food chain, chemical pollution continues as a major concern in coastal ecosystems (Goto, Wallace, 2010).

Remediation efforts were conducted at both locations in 2000. These efforts consisted of removing contaminated soil. Both sites are still under investigation to determine the present contamination levels.

Live worms and grass shrimp samples were collected at both sites in 2007. Tissues obtained from these samples were used to compare body burdens of metals before and after remediation.

Shrimp and worms are used to assess the biological health of a wetlands site where contamination is suspected. Worms collected for this study live amidst the contaminated sediments. Shrimp live higher in the water column and are relatively mobile. Shrimp and worms take up pollutants internally during normal feeding. Once digested the pollutant is retained in tissues. Animals higher in the food chain have the ability to "accumulate" toxins by eating contaminated prey.

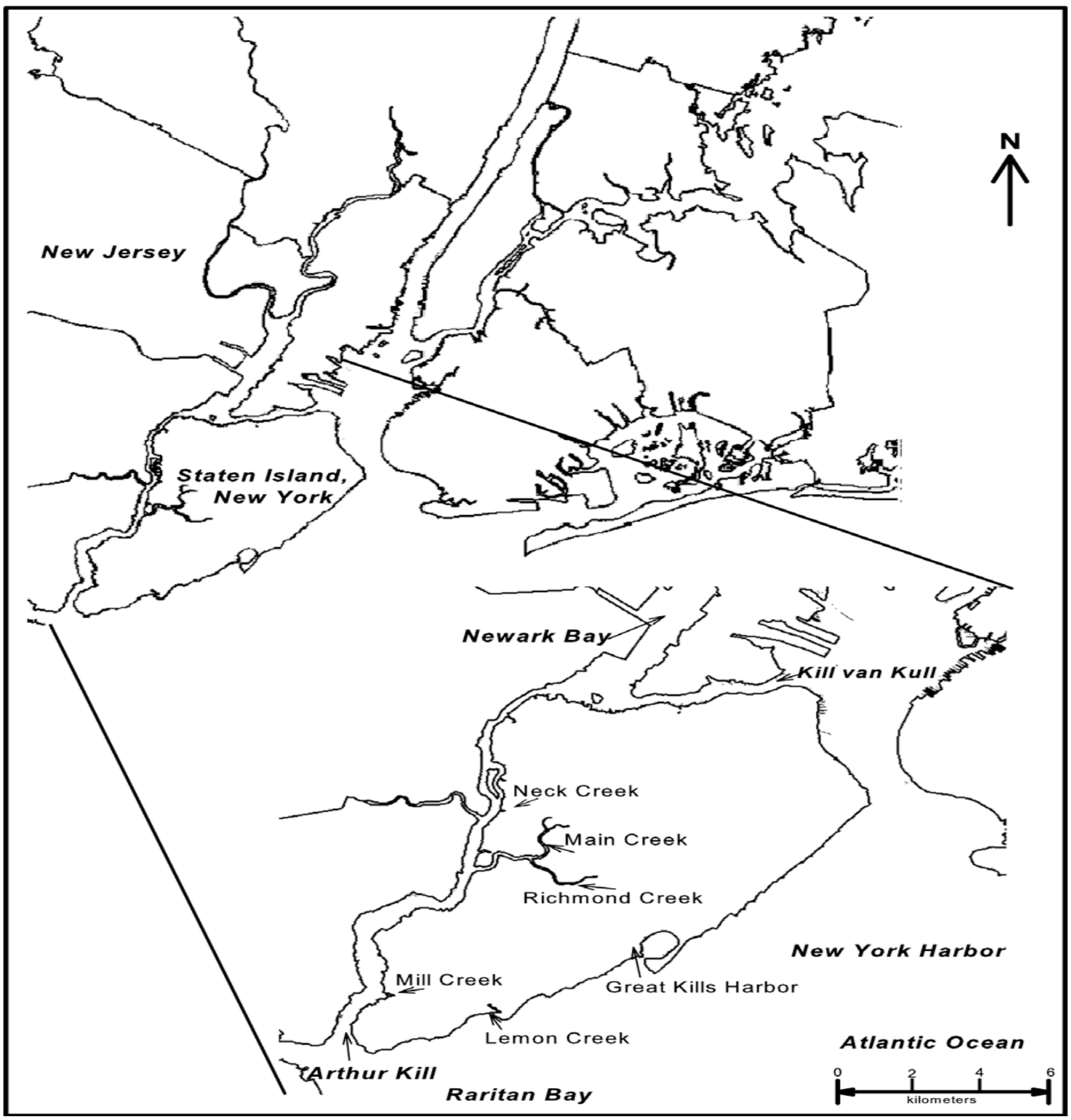


Figure 1. Map of study sites along the Arthur Kill Waterways and along the south shore of Staten Island (Goto, Wallace, 2010).

Brief Geological History of the Region

The earth has witnessed four major glacial ages in the past 2 million years. The most recent glacial age, the Wisconsin, began approximately 90,000 years ago. When it ended about 22,000 years ago, the Wisconsin glacier covered all of what is now New York City. The glacier's southern boundary is outlined by a terminal moraine that crosses Staten Island (See Figure 2 below).

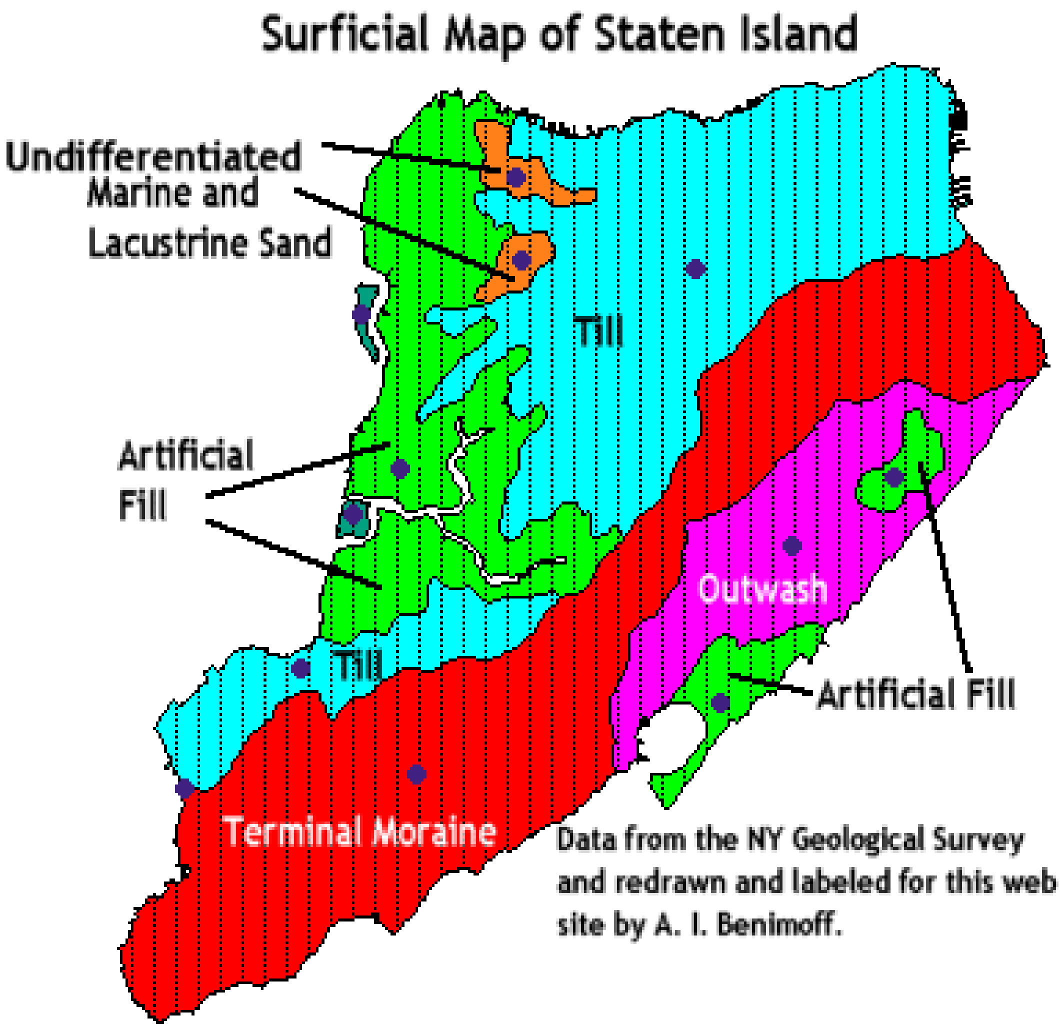


Figure 2. Terminal moraine shows glacier's southern boundary (Benimoff, Ohan, 2003).

As the glacial ice melted, a blanket of loose, unconsolidated, poorly sorted glacial till was left behind. Glacial melt water also deposited outwash plain sediment south of the terminal moraine. (Benimoff, Ohan, 2003)

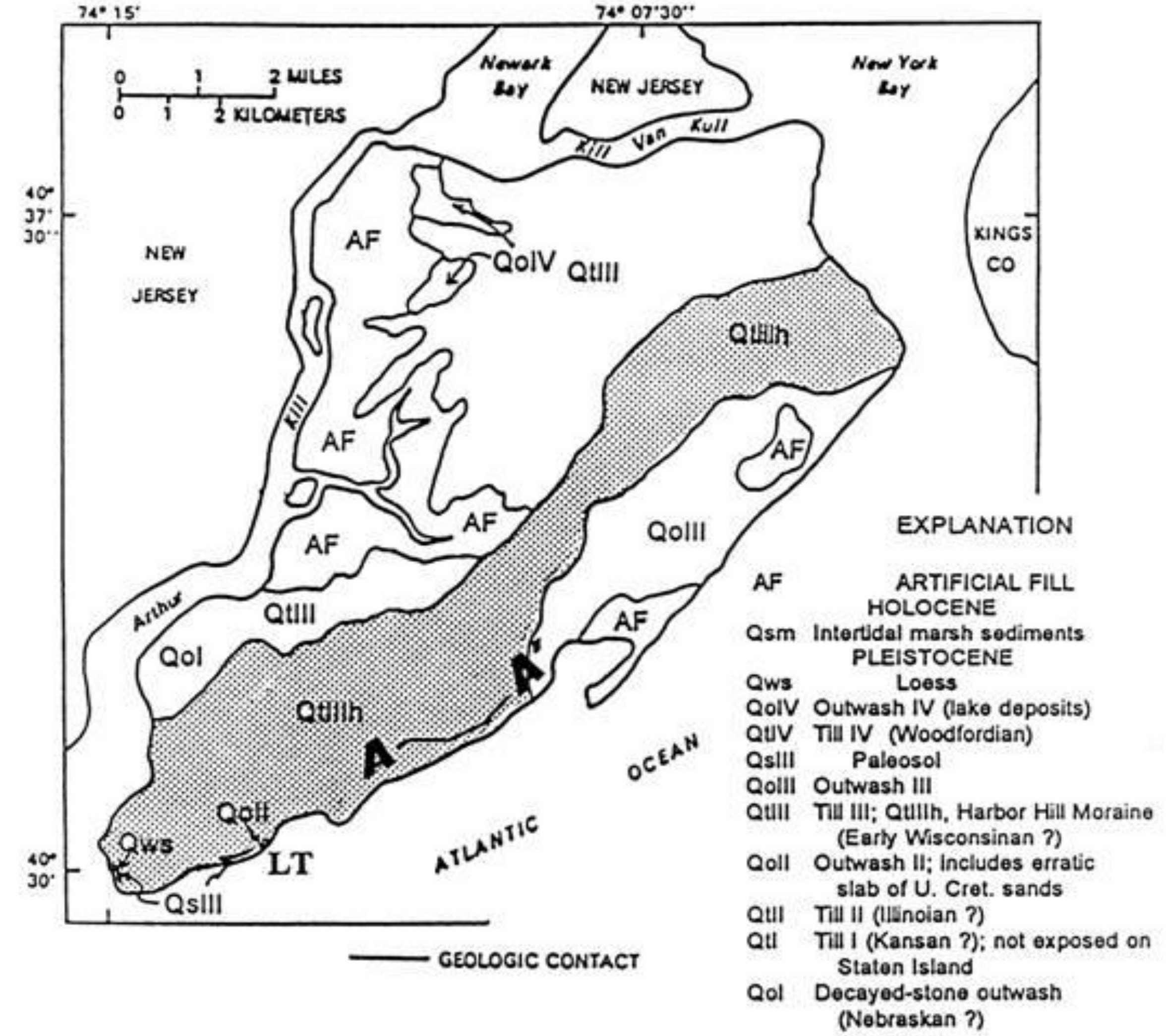


Figure 2.1 Map of surficial deposits on Staten Island modified from Sanders and Merguerian, (1994). Map produced Sanders using outline from Soren, (1988) and Caldwell (1989). Shaded area is the Harbor Hill Moraine; LT on diagram is a light tower on the shore of Princes Bay.

The Vital Role of Salt Marshes

Salt marshes play a vital role in supporting all forms of life. Marshes like Lemon Creek act as natural filtration systems improving water quality by trapping pollutants that would otherwise contaminate bays and oceans. Salt marshes are also among the richest wildlife habitats. Marshes support wildlife by providing a productive habitat for many species.

When the last glaciers melted, the oceans rose to their current levels. Sediments washed from land were deposited offshore in narrow sandy strips. These formed long islands parallel to the shoreline. The pounding ocean surf and a relatively calm landward shore bound these barrier islands.

The presence of saltwater makes survival difficult for most vegetation. Marshes further reduce erosion by trapping sediments that would otherwise be washed away. A resilient species of cord grass (*Spartina alterniflora*) was able to colonize the beaches, and mudflats despite being covered twice daily by ocean tides. This cord grass is still abundant along the Atlantic coast.

As this specialized grass spreads, its' stems trap floating debris. Sediments and particles of decaying matter slowly build up forming a nutrient rich mud that supports a complex food web. For example, the fiddler crab (*Uca*) and ribbed mussels (*Geukeasia demisea*) have formed a mutually beneficial relationship with the cord grass. Crabs and mussels benefit by feeding on decaying matter trapped by cord grass roots. The cord grass benefits from the burrowing activity of the fiddler crab, which aerates the soil. The mussels' excretion provides necessary nitrogen for the cord grass to thrive.

At the end of each season, the cord grass dies creating a spongy peat. Each year the peat layer raises the surface of the marsh enabling it to colonize new territory. This makes it possible for new plants with less salt tolerance to colonize the peat. This increases plant diversity that may invite other plant or animal species to join. The death of the cord grass also permits the formation of separate plant communities, an intertidal marsh, and a salt meadow.

Lemon Creek

This site is located just northeast of the light tower in Figure 2.1 inside Lemon Creek Park's 77 acres of narrow oak filled slopes that surround a sizable tidal marsh area. The creek and salt marsh represent the last remnants of a wetland system that flows through Bloomingdale Woods. Lemon Creek is one of a few relatively undisturbed tidal marshes on Staten Island. Our viewing site will be on a bridge on Hylan Boulevard, which runs across Lemon Creek. (40˚31'03.66"N; 74˚12'04.01"W from Google Earth).

NYC Dept. of Parks and Recreationwww.nycgovparks.org)

Mill Creek

Mill Creek is located near the confluence of the Arthur Kill and the Raritan Bay has been polluted from historic smelting activities. The Tottenville Copper Company, a metal processing plant was established adjacent to Mill Creek in 1900. By 1941, the plant was operating as the Nassau Metals and Refining Company. Thirty years later the plant was renamed the Nassau Recycling Plant whose chief purpose was to recycle copper and zinc from old telephone cables.

Mill Creek runs through the property formerly occupied by the metals smelting and refining plant. The location is less pristine than the Lemon Creek site chosen for this tour. Our viewing site will be on Arthur Kill Road looking west toward New Jersey. (40˚31'12.15"N; 74˚14'23.24"W from Google Earth) The Arthur Kill, a 25-mile long tidal strait, runs northeast to southwest from Newark Bay to Raritan Bay and separates Staten Island from New Jersey. Despite improvements in recent years, levels of trace metals and PCB's in some sediment remain high. Accidental discharge of petroleum products into the Arthur Kill is a chronic problem due to the number of oil transfer facilities. Although major oil spills occur infrequently, many small oil spills occur each year. These events are compounded due to the low flushing and increased residence time of water. Urban runoff containing petroleum products also adds to the burden on this system. Mill Creek is a typical marsh within the Arthur Kill system. Mill Creek is located just south of the Outerbridge Crossing on the New York side.

Results of this study:

As of the date this field guide was written, the results of this investigation are still be evaluated. These results will be discussed during the field visits and hard copies will be available during the NYSGA conference in September 2010.

Road Log.

Leaving main entrance to CSI, 2800 Victory Blvd. Staten Island, New York 10314 to Mill Creek, near 4996 Arthur Kill Road, Staten Island, New York 10307. Total distance is approximately 9.41 miles, (MapQuest)

1. Leave through front gate at CSI. Turn left onto Victory Blvd. toward Richmond Avenue 0.5 miles
2. Make left turn onto Richmond Avenue. Proceed along Richmond Avenue past the Staten Island Mall, which will be on your left. Section of Fresh Kills Landfill will be on your right. 3.2 miles
3. Stay on your right as you go past the mall. Take Drumgoole Road West leading toward Korean War Veterans Parkway. 0.3 miles
4. Take Korean War Veterans Parkway South toward Outerbridge Crossing. (Do not get on the bridge!) 2.7 miles
5. Take the Maguire Avenue/Bloomingdale Road exit. Follow arrows to Arthur Kill Road. 0.2 miles
6. Proceed to Drumgoole Road West. 0.4 miles
7. Continue on Drumgoole Road. Note, Drumgoole Road West becomes Veterans Road East 0.5 miles
8. Turn left onto Englewood Avenue. 0.1 mile
9. Turn left onto Veterans Road West 1.0 miles
10. Turn left onto Arthur Kill Road 0.6 miles
11. Viewing location is a small bridge over Mill Creek. Vehicles may be parked off road on the right facing New Jersey.

Directions from Mill Creek to Lemon Creek Bridge (Hylan Blvd). The distance is approximately 3.07 miles from Mill Creek. (MapQuest)

1. Turn left, proceed north on Arthur Kill Road toward Richmond Valley Road.

2. Turn right onto Richmond Valley Road. 0.2 miles
3. Turn right onto Page Avenue. 1.0 mile
4. Turn left onto Hylan Blvd. 1.6 miles
5. Turn right onto Bayview Avenue. Walk to over on Hylan Blvd.

References:

Benimoff, A.I., and Ohan. A.A. (1996) Highlights of Staten Island Geology, *in* Field Trip Guide for the 68th annual meeting of the New York State Geological Association eds. A.I. Benimoff and A.A. Ohan pages 125-138

Caldwell, D.H., 1989, Surficial geology map of New York, Lower Hudson sheet: New York State Museum Map and Chart Series 40 scale 1:250,000.

Goto, Daisuke;Wallace, William G., 2010, Relative importance of multiple environmental variables in structuring benthic macro faunal assemblages in chronically metal polluted salt marshes, Marine Pollution Bulletin, 60,363-375.

New York City Department of Parks and Recreation "Salt Marshes in New York City Parks-Lemon Creek", www.nycgovparks.org)

Perez, M.H.; Wallace, W.G. 2004, Differences in Prey Capture in Grass Shrimp, *Palaemonetes pugio,* Collected Along an Environmental Gradient, Archives of Environmental Contamination and Toxicology. 46, 81-89.

Sanders, J. E.; and Merguerian, Charles, 1994, Glacial geology of the New York City region, p. 93-200 *in* Benimoff, A. I., *ed*., The geology of Staten Island, New York: Geological Association of New Jersey Annual Meeting, 11th, Somerset, NJ, 14-15 October 1994, Field guide and proceedings, 296 p.

Soren, Julian, 1988 Geologic (sic) and geohydrologic reconnaissance of Staten Island, New York: United States Geological Survey Water Resources Investigations Report 87-4048,22p