



AASG Mid-Conference Field Trip

Hosts:

New Jersey Geological Survey

National Park Service

Sterling Hill Mining Museum

**Geologic Field Trip Guidebook for
102nd Annual Meeting – Association of American State Geologists**

**Delaware Water Gap, Pennsylvania – New Jersey to
Sterling Hill Mining Museum, Ogdensburg, New Jersey**

**Mid-Conference Field Trip
June 29, 2010**

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(Photo by Ron W. Witte).

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Message from the New Jersey State Geologist

Welcome to the AASG 2010 Mid-Conference Field Trip - a day to traverse northern New Jersey and visit classic areas of geological studies in New Jersey's Valley and Ridge and New Jersey Highlands. The trip is co-sponsored by the National Park Service and the Sterling Hill Mining Museum.

The Valley and Ridge is a special place for me. It was the start of my career in Geology. As a scout I was impressed by the vistas and beauty of the Appalachian Trail and Kittatinny Ridge. I hiked down the Pennsylvania side and up the Jersey side and became familiar with the rocks that hold up the Delaware Water Gap. Later while in high school this area was in the news - development plans for the Tocks Island Dam, a water resource and flood control project for the Delaware. The activism to preserve the region eventually resulted in the Delaware Water Gap National Recreation Area that we will visit today.

As an undergraduate Geology major my first visit to an underground mine was to the Sterling Hill Mine. Samples I collected became the focus of my first experience in mineralogy research. Later when I returned to New Jersey as a practicing Geologist, it was coincidentally my first assignment to assist in the shutdown of the mine and preservation of the core. In subsequent years the mine was preserved and now flourishes as a museum. Our final stop today will visit this testament to the rich mining history of New Jersey.

Much of the Highlands region has recently been preserved as the one of the outstanding natural resources of the State. Both the water and land resources have been recognized as important in maintaining the quality of life to all New Jersey citizens. Because of these initiatives much of the natural aspects that first impressed me will be there for future generations. I hope we are able to share some of the special features of the region with you today. Enjoy the trip!

Karl W. Muessig

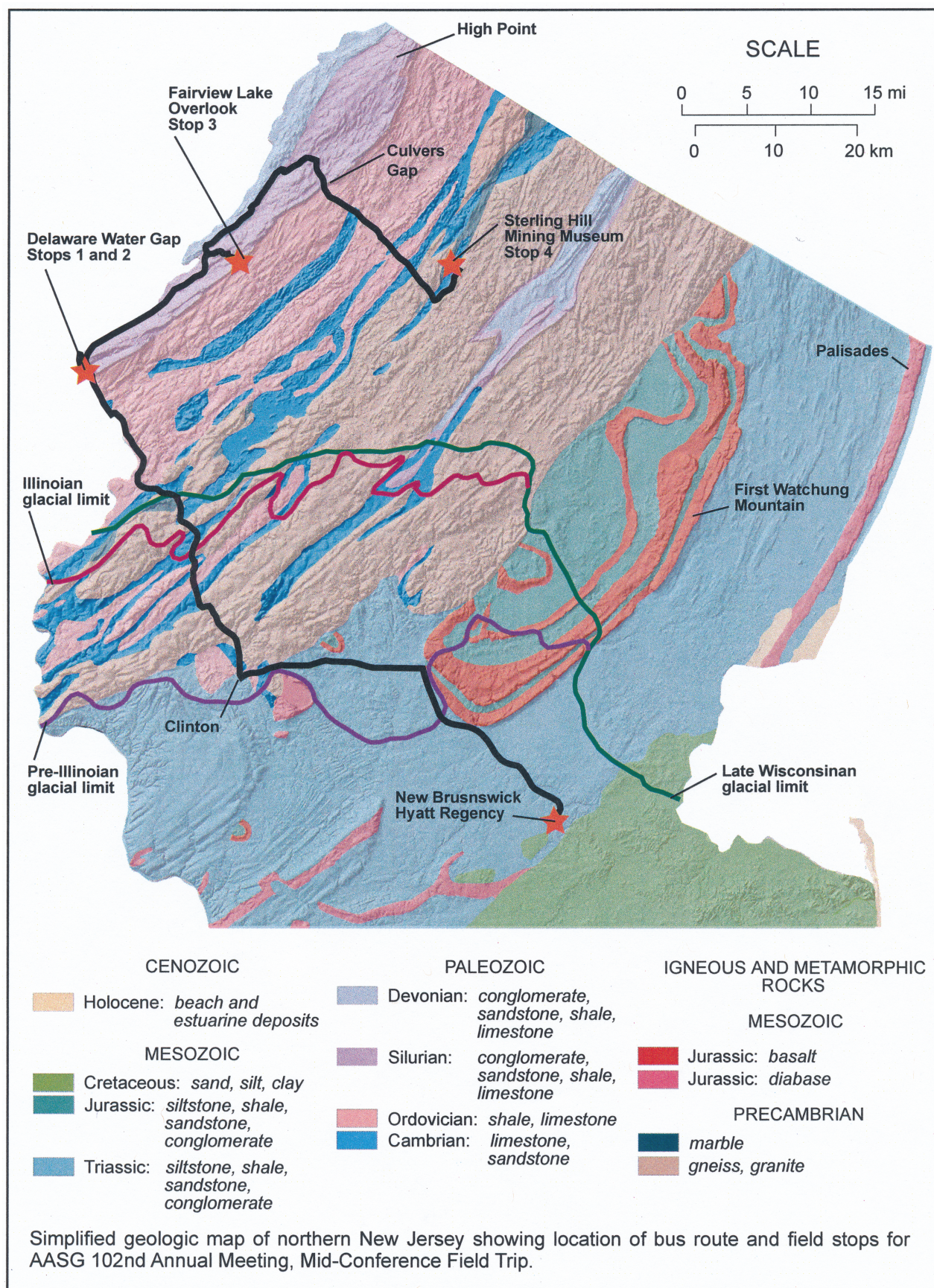
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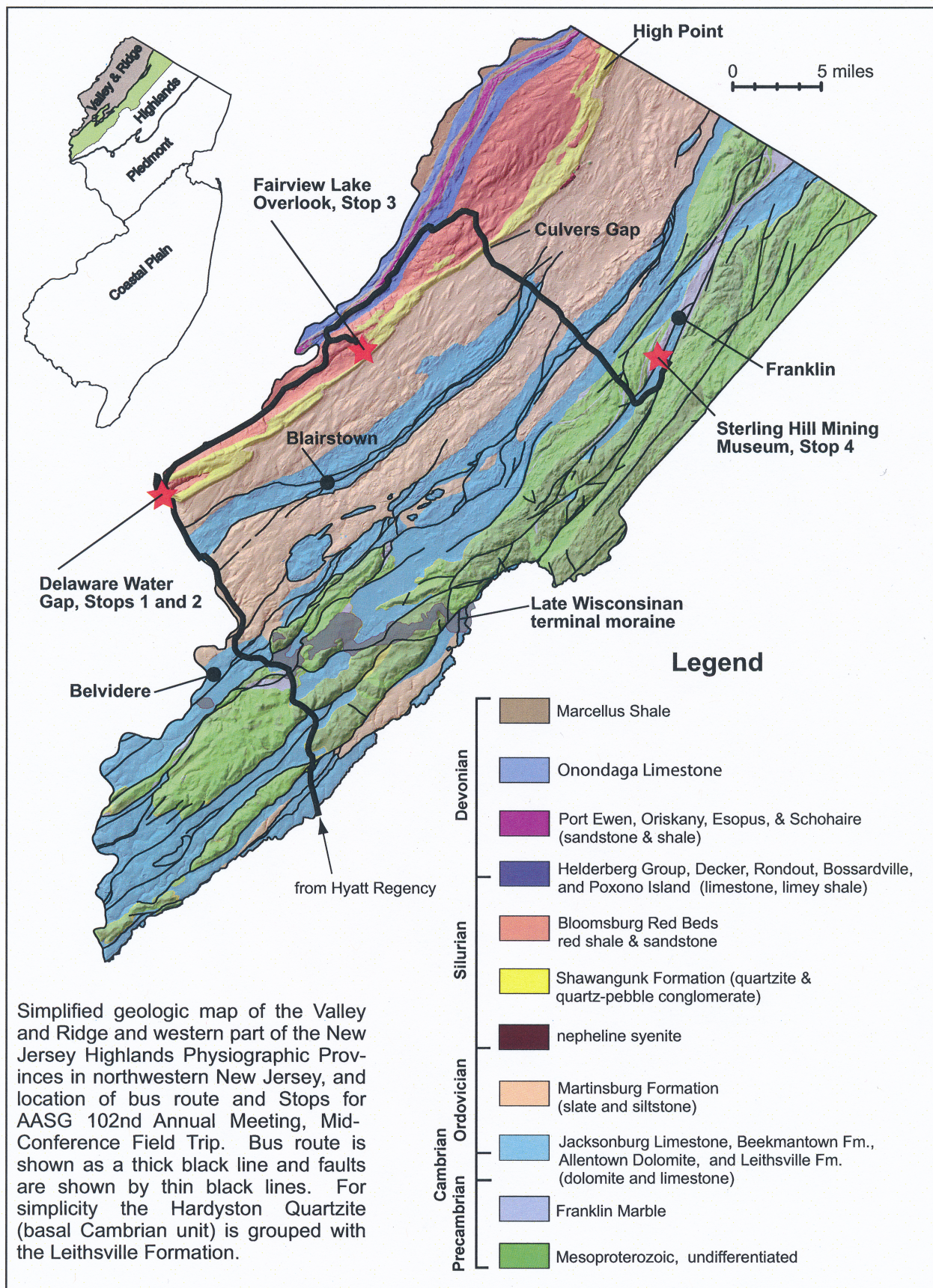
The organizers and editors of the AASG Mid-Conference Field Trip for the 102nd Annual Meeting of Association of American State Geologists are greatly indebted to the many individuals and organizations that made this guidebook possible. Thanks to the field trip leaders and contributors for their hard work and dedication to preserving New Jersey's rich geologic heritage. A special thanks to Jack Epstein (USGS) who enthusiastically agreed to be a field trip leader, even though he was under no obligation to do so. His contributions, based on 40 plus years of experience investigating Paleozoic stratigraphy, structural geology, paleontology, geomorphology, glacial geology, geologic hazards, and wealth of knowledge about local lore, are invaluable in furthering our understanding of the geology in and around Delaware Water Gap.

We are also extremely grateful to the National Park Service for providing transportation costs (motor coaches are an expensive way to travel, but very necessary for this type of trip). Thanks also to John Donahue, Superintendent (DEWA), Deborah Nordeen, Public Relations Officer (DEWA), and DeNise Cooke-Bauer, Geologic Resources (DEWA) for their support in bringing a large geologic field trip to the park. Thanks to Bruce Heise, Geologic Resources Division, NPS (Denver) for providing logistical support and all his work on the *value of geology* in National Parks.

Thanks to Richard and Robert Hauck, owners of the Sterling Hill Mining Museum, for hosting part of the field trip. They have preserved a very important piece of New Jersey's industrial and geologic heritage that would have been lost had it not been for their perseverance and tireless efforts.

We are also grateful to Jane Uptegrove and Mark Godfrey (New Jersey Geological Survey) for driving the support vehicles and providing our guests with lunch and cold drinks. Lastly, many thanks to the staff of the New Jersey Geological Survey who helped with the guidebook and field trip.





GEOLOGIC HISTORY OF NEW JERSEY'S VALLEY AND RIDGE PHYSIOGRAPHIC PROVINCE

by

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INTRODUCTION

Kittatinny Mountain's rocky crest (fig. 1), karst in Paulins Kill Valley, and the watery fens along the Muckshaws are just a few of the geologic landscapes that make up New Jersey's Valley and Ridge physiographic province. The rock and soil that give this land its distinctive form are the products of a geologic history that stretch back in time more than one half a billion years.

The Valley and Ridge province (fig. 2a) in New Jersey encompasses an area about 536 square miles and includes all the land west of the New Jersey Highlands to the Delaware River. It is largely underlain by sedimentary rock deposited in and along the coast of ancient seas from 540 to 400 million years ago. These rocks, tilted, folded, and fractured during the Taconic and Alleghenian orogenies (past mountain building events), form a series of parallel to nearly parallel northeast- to southwest-trending belts that overall become younger going northwest toward the Delaware River (fig. 2b). Due to this congruent alignment and the varying ability of the different rocks to withstand erosion over long periods, the land has a very distinctive, northeasterly to southwesterly topographic grain that consists of long linear valleys and ridges (fig. 2c.). Based on similarity of topography, bedrock, and geographic setting, New Jersey's Valley and Ridge consists of three distinct physiographic areas. These are Kittatinny Valley, Kittatinny Mountain, and Delaware Valley (fig. 2d).



Figure 1. Curving ridge line of Kittatinny Mountain looking southwest from High Point, Sussex County, New Jersey.

PHYSIOGRAPHY

Kittatinny Valley (fig. 2d) forms the broad lowland between the New Jersey Highlands and Kittatinny Mountain. Erosion of dolomite, limestone, slate, siltstone, and sandstone, all of Lower Paleozoic age (fig. 2b) and glacial modification have given the valley a varied topography. Dolomite and limestone, rocks that are susceptible to dissolution, typically underlie the Paulins Kill, Pequest, and Wallkill River valleys. Relief here is as much as 200 feet and rock outcrops are very abundant. The land is very uneven, consisting of many small rocky knolls and ridges, sinkholes, and stream-less valleys (fig. 3a). Springs are common and in places, large, irregular depressions mark the valley floor. The largest of these depressions, Swartswood Lake, Newton Meadows, Crooked Swamp, and Big Springs (fig. 2d) were formed by the extensive dissolution of carbonate bedrock and glacial scour. Glacial lake deposits of silt, clay, and sand, and gravel, laid down during the last ice age, fill these basins, and many parts of the river valleys that flow away from them.

Northeast- to southwest-trending belts of slate, siltstone, and some sandstone underlie the higher land in Kittatinny Valley (fig. 2b). Overall, average elevation here is greater than the carbonate-floored valleys by about 300 feet, with relief as much as 600 feet. Topography consists of undulose, narrow to broad ridges and hills (fig. 3a). Glacial erosion has accentuated this topography, sharpening relief and streamlining ridge crests. Small streams that drain across this region's topographic grain have deeply dissected these uplands.

Kittatinny Mountain (fig. 2d) is a prominent ridge underlain by quartz-pebble conglomerate, quartzite, red sandstone and shale all of Lower Paleozoic age (fig. 2b). The land here is a mosaic of diverse mountainous

Figure 2a - New Jersey's Valley and Ridge Physiographic Province

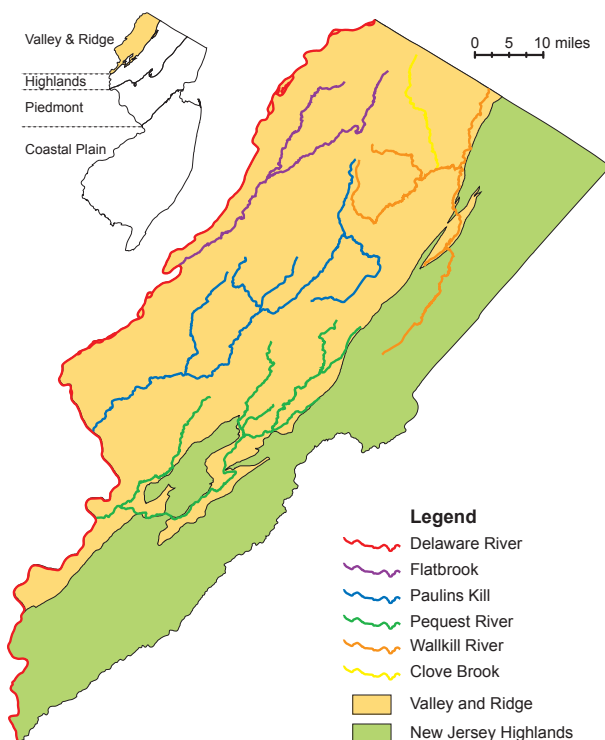


Figure 2c - Color-shaded Relief

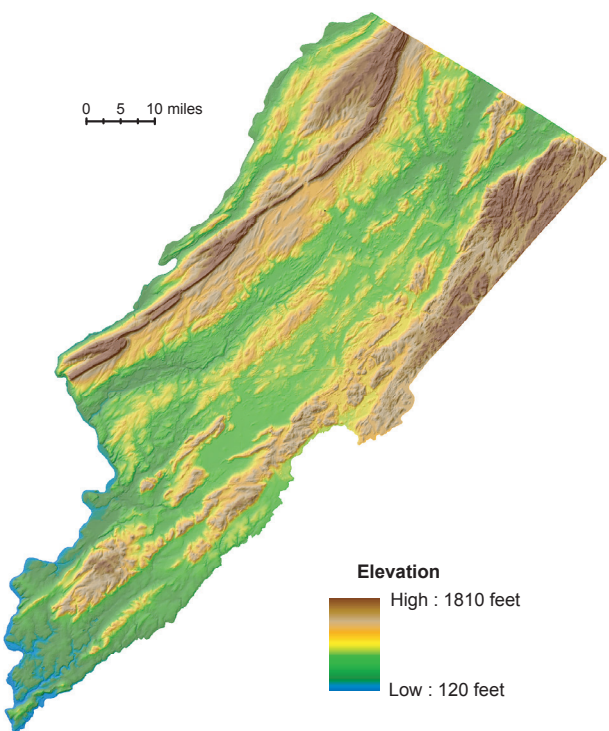


Figure 2b - Simplified Bedrock Geology

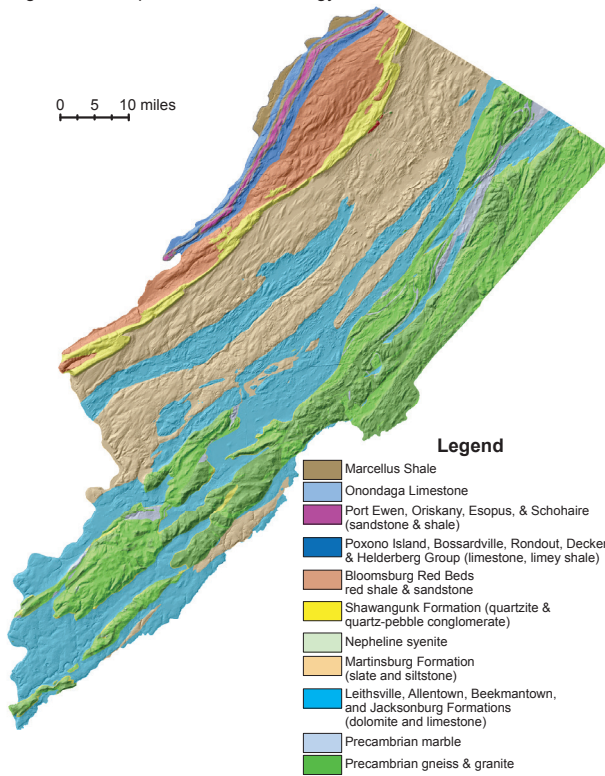
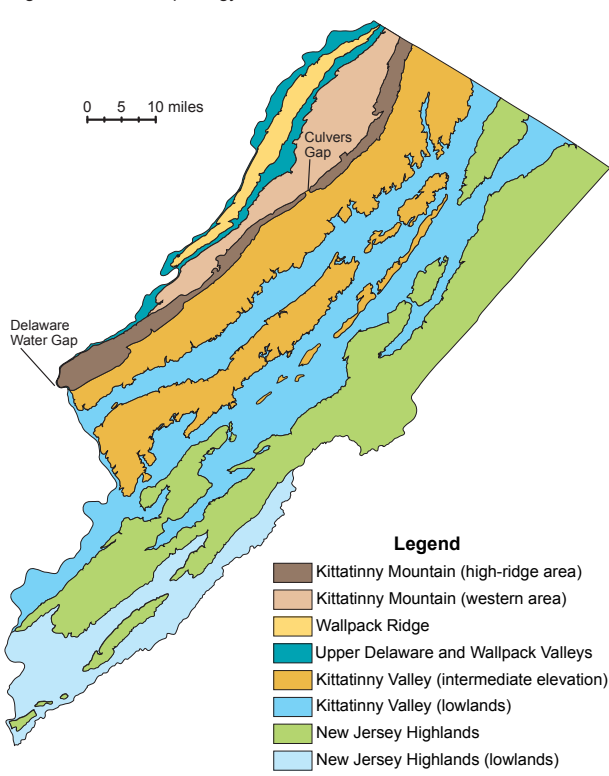


Figure 2d - Geomorphology



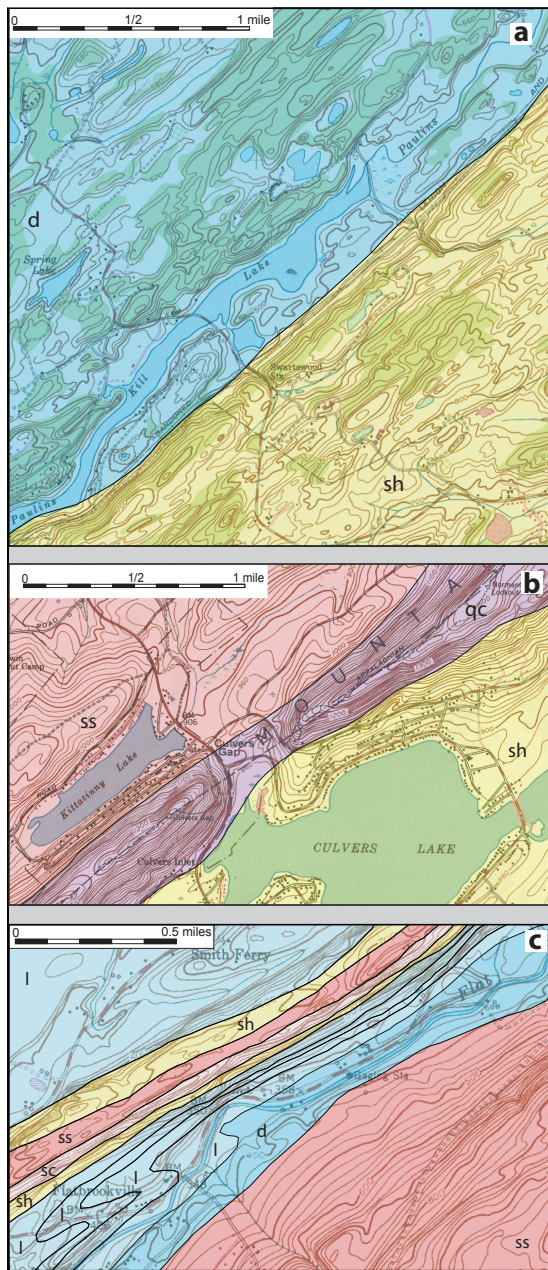
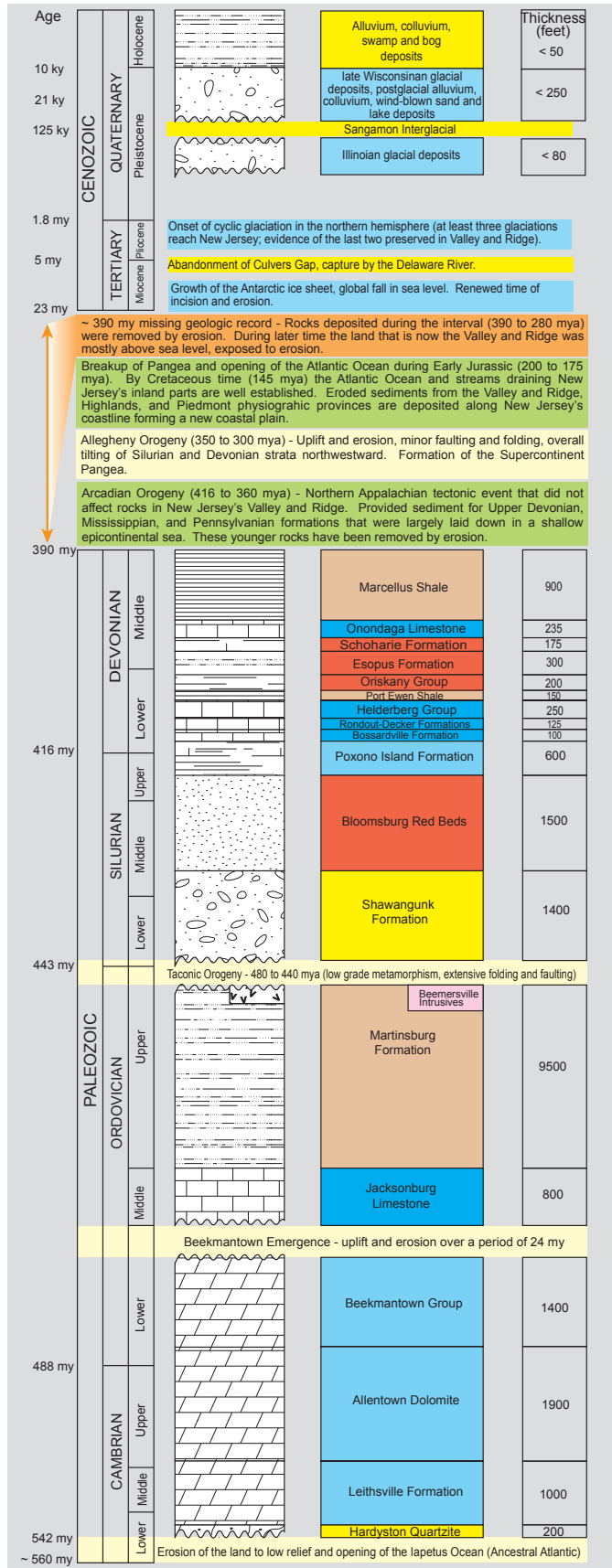


Figure 3, *above*. Topography and simplified geology of areas in Kittatinny Valley (a), Kittatinny Mountain (b), and Wallpack Ridge (c). Congruent alignment of topography is strongly controlled by the southwest trend of rock formations and their resistance to erosion. Key to lithotypes: d-dolomite, l-limestone, sh-siltstone and slate, qc-quartzite and conglomerate, ss-sandstone, sc-sandstone and conglomerate. Figure 4, *right*. Stratigraphic column of New Jersey's Valley and Ridge physiographic province.



terrain, consisting of hilly forest, barren, rocky, ridge crests, and many wetlands. The mountain consists of two topographic sections. The first, informally named the high ridge area (fig. 2d), forms the eastern flank of the mountain. This prominent topographic area is underlain by quartzite and quartz-pebble conglomerate of the Shawangunk Formation. These tough, resistant rocks form slightly sinuous, curvilinear, narrow to broad, parallel ridges along the mountain's crest (fig. 3b). Maximum relief is nearly 1000 feet with the highest point in New Jersey, at 1803 feet above sea level, marked by the High Point Monument. Rock outcrops are very abundant and most exhibit a smoothed and streamlined look, the product of glacial erosion. The mountain's steep southeast face forms a nearly continuous escarpment that is broken by Culvers Gap and a few smaller wind gaps.

The second area, called the western hills (fig. 2d) includes the lower parts of the mountain and slopes that lie west of the high ridge area. Red sandstone and shale of the Bloomsburg Red Beds lay beneath these lower hills. Topography consists of rolling hills, long moderate to steep slopes, and low relief, northeast- to southwest-trending ridges (fig. 3b).

The Delaware Valley area consists of Minisink Valley, Wallpack Valley, and Wallpack Ridge (fig. 2d). The valleys are narrow, deep, and trend northeast to southwest, following belts of weaker rock that are chiefly limestone, shale, and limey shale of Silurian and Devonian age. In most places, thick deposits of glacial outwash and alluvium cover the valley's floors. Wallpack Ridge forms a narrow, rocky cuesta that separates the two valleys, rising as much as 300 feet above their floors. Thinly bedded, gentle to moderately northwest-dipping siliclastic rocks (shale to conglomerate) and limestone underlie the ridge. Sandstone and conglomerate form the ridge's spine with less resistant units of shale and limestone forming the ridge's flanks. Due to the northwest dip of the strata, a high 200 to 300 foot scarp (fig. 3c) marks the southeast face of the ridge.



Figure 5. Shawangunk Formation on the east shore of Crater Lake, Kittatinny Mountain. This rock consists of thin layers of quartzite and quartz-pebble conglomerate. Bedding dips from right to left. Because of its very high quartz content and that the sand grains and larger clasts are tightly held together by silica cement, this formation is highly resistant to weathering.

Major streams (fig. 2a) include the Delaware River and its tributary's Flat Brook, Paulins Kill, and Pequest River, and the Walkill River, which is a tributary to the Hudson River. All trunk streams and main tributaries flow southwestward except the Walkill River, which flows northeastward into New York. Easily eroded bedrock appears to control the course of these larger streams. Pequest, Paulins Kill, Walkill, and Flat Brook valleys overlie dolomite and limestone. The Delaware Valley above Wallpack Bend is largely cut in shale and limestone, while downstream from the bend to the Delaware Water Gap, the valley is largely cut in dolomite. Below the water gap, the river flows southeastward following the trend of cross-joints in the local bedrock. In places, smaller tributaries also follow the trend of bedrock cross-joints and form a congruent drainage pattern. A dendritic (branching) drainage pattern has formed where streams flow over thick glacial sediment.

GEOLOGIC HISTORY – building geologic landscapes

Erosion by running water, dissolution, and glaciation, all working on a resistant framework of folded, and tilted sedimentary rock, shaped New Jersey's Valley and Ridge to its distinctive appearance. Because changes to the landscape brought about by tectonism, magmatism, eustasy, and glaciation often operate at very slow rates, it is convenient to discuss geologic history as a series of connected, singular events. However, these events may span millions of years and they may chronologically overlap each other. The stratigraphic column shown in Figure 4 defines major periods in the Valley and Ridge's history based on the succession and age of its geologic formations, and chronological arrangement of geologic events that are important to the development of the province's geomorphology.

Scale and time are used to define geologic landscapes. Large-scale features, such as orogens (mountain



Figure 6. Well-developed joints (vertical fractures) and cleavage (sub-vertical platy sheets) in the Esopus Formation, Wallpack Ridge. Bedding dips into the photograph.



Figure 7. Allentown Dolomite southwest of Balesville, Kittatinny Valley. This rock consists of a gray, weathering tan, magnesium-calcium carbonate that was deposited in a warm shallow sea about 500 million years ago. Because of its mineralogic makeup it is susceptible to weathering by dissolution. Bedding dips gently from right to left.

dolomite with lesser limestone, is very susceptible to erosion through chemical weathering, even though it is a harder rock than many of the shales found throughout the province. Collectively, quartz-rich rock will form the higher areas, carbonate rock and some shale the lowest with sandstone, siltstone, and slate forming areas of intermediate elevation. Comparison of a shaded relief-map and bedrock map of the Valley and Ridge show a strong correlation between rock type and elevation (fig. 2b). Thus, the location, size and spacing of ridges and valleys are largely due to rock composition and its distribution. For example, the Shawangunk Formation consists largely of quartzite and quartz-pebble conglomerate, rocks that are very resistant to erosion. Kittatinny Mountain is underlain by this formation and it forms the highest land in New Jersey. The formation's outcrop belt is largely coincident with the high ridge section of Kittatinny Mountain.

Induration, which represents the tightness or consolidation of a rock's grains, also controls a rock's ability to withstand erosion. In sedimentary rocks individual grains are typically held together by silica or calcium carbonate cement. Silica-cemented rock will resist erosion to a much greater degree than rock that is cemented together by calcium carbonate. The quartzite and quartz-pebble conglomerate that underlie Kittatinny Mountain (fig. 5) are much more resistant to weathering and erosion than the calcium carbonate-

belts), cover expansive areas, consist of many landforms, and require millions of years to take shape. Small-scale features such as glacial lake deltas cover small areas, typically consist of a single landform, and require only tens to hundreds of years to form. Most of the regional geomorphic elements of the Valley and Ridge, those that are easily seen on a page-sized color-shaded relief map (fig. 1c), require 10^5 to 10^6 years to develop. Features that form over shorter periods (i.e. flood plains, end moraines, drumlins) are also seen on the small-scale map, but because of their lesser size, they typically only overprint and not obscure the regional topography.

In New Jersey's Valley and Ridge there exists a very strong correlation between elevation and lithology with the higher land underlain by more resistant rocks (fig. 2b). Collectively, bedrock (figs. 2b and 4) forms a framework that resists erosion by water, wind, ice and chemical weathering. Its lithology is a relative indicator of weathering susceptibility and therefore erosion. Bedrock in the Valley and Ridge is chiefly sedimentary consisting of siliciclastic and carbonate rock. Siliciclastic rocks consist of conglomerate, sandstone, siltstone, and shale and their low-grade metamorphic equivalents of quartzite and slate. All these rocks except shale and slate are quartz rich, the mineral's abundance a general indicator of resistance to erosion. Quartzites and conglomerates in the Valley and Ridge contain the greatest amount of quartz with lesser amounts in the sandstone followed by siltstone, shale and slate. These latter varieties typically contain a mixture of quartz and less resistant minerals such as feldspar, clays (aluminosilicates), and lithic fragments. Carbonate bedrock, chiefly



Figure 8. Interlayered siltstone (thicker layers) and shale (thin layers) in the Martinsburg formation, Kittatinny Valley. Bedding dips into the picture. Subvertical joints and cleavage (subvertical sheets that dip toward the viewer) cut the outcrop forming steps.

sandstone. Locally, it is conglomeratic at its base. The Hardyston sediments were derived from weathered Precambrian rocks during the waning stages of an erosional cycle that started in Late Precambrian time. These deposits represent river and beach deposits laid down on an ancient coastal plain. The formation lies along the eastern edge of Kittatinny Valley where Paleozoic rocks are in contact with Precambrian Rocks of the Highlands. Because the formation in most places is very thin, it is not a major landscape-forming unit. A few small areas of Precambrian rocks are also found in Kittatinny Valley, but because of their very limited extent they are not discussed.

The Kittatinny Supergroup consists of the Leithsville Formation, Allentown Dolomite (fig. 7), and Beekmantown Group. Collectively these formations are about 4300 feet thick. They consist largely of dolomite with subordinate limestone, chert, and minor quartz sandstone, siltstone and shale. These deposits were laid down in warm, shallow seas in near-shore, tidal, and lagoonal environments. The top of the Beekmantown Group is marked by a major unconformity that represents an extended period of emergence and erosion that lasted millions of years (fig. 4). These formations make up wide belts of carbonate rock that typically underlie lowlands and valleys in Kittatinny Valley. Sinkholes, caves, streamless valleys, sinking streams, and springs, formed by the dissolution of dolomite and limestone, are common here. Great Meadows, Swartswood Lakes area, Big Springs, Newton Meadows, and Crooked Swamp are places where large amounts of rock have been removed by dissolution.

The Jacksonburg Limestone is a transitional unit that lies between the carbonate rocks of the Kittatinny Supergroup and the clastic rocks of the Martinsburg Formation. It can be as much as 800 feet thick and it rests unconformably on the Beekmantown Group. The lower part of the formation is a fossiliferous limestone that shows that there was a rise in sea level following the Beekmantown emergence. The upper part is principally a muddy limestone. Clay in the upper part of the formation reflects proximity to a landmass and a return to a deeper water environment. The Jacksonburg Limestone is found in Kittatinny Valley where it

cemented sandstones in Kittatinny Valley.

Faults, joints, and cleavage create zones of weakness (fig. 6) in rock. These partings act as conduits for agents of mechanical and chemical weathering which greatly make the rock more susceptible to erosion. The discordant and congruent alignment of several smaller streams in Kittatinny Valley and possibly the many gaps that cut through ridges of resistant rock suggest that these features preferentially developed in areas underlain by highly jointed rock.

BEDROCK FORMATIONS

Cambrian and Ordovician (542 to 443 mya)

The Hardyston Quartzite is Lower Cambrian age. It may be as much as 200 feet thick, but typically is much thinner and rests unconformably on the Precambrian rocks of the Highlands. The formation is largely quartzite with lesser amounts of arkosic and dolomitic



Figure 9. Medium- to thick-bedded sandstone and siltstone of the Bloomsburg Red Beds southwest of Millbrook near Van Campens Glen. Beds dip gently northwest (into photo). In places, a southeast dipping spaced cleavage is seen (area to the right of hammer, photo center).

typically lies on the western side of the carbonate belts. Its land-forming characteristics are similar to the Kittatinny Supergroup.

The Martinsburg Formation is as much as 9500 feet thick and it rests conformably on the Jacksonburg Formation. It consists of repetitive layers of shale, slate, siltstone and sandstone that record a deepening oceanic basin. Generally, the finer-grained rocks (shale and slate) are present in larger proportions in the lower part of the formation (fig. 8). The coarsening of the formation upwards reflects an increasing proximity to a continental landmass, the result of the Proto North American plate moving toward an unnamed microcontinent. The Martinsburg Formation is found throughout Kittatinny Valley. It typically forms the higher hills and ridges in Kittatinny Valley. Cleavage, a physical character of the rock that allows it to split along finely spaced planes, is very well developed in this formation (fig. 8). Because of this, the rock is very susceptible to frost shattering, a weathering process that breaks the well-cleaved rock into shale-chip rubble.



Figure 10. Biogenic limestone in the Coeymans Formation, Montague. Outcrop represents reef debris found on the flank of a small patchwork reef. Crinoid fragments are common with a rare stromatoporoid seen near the quarter. Matrix material is chiefly coral fragmnets (Favosities).

The Taconic Unconformity (fig. 4) marks the upper surface of the Martinsburg Formation. It represents a period of 20 to 30 million years during which the upper part of the formation was lifted above sea level and then lowered by erosion. The younger Shawangunk sediments were deposited on this surface. At some time during the Late Ordovician, rocks of the Beemerville intrusive complex were formed. These rocks, about 435 million years old, are of magmatic origin, emplaced during the Taconic orogeny (fig. 4). Because these rocks have such a limited outcrop extent, they are not an important landscape-forming rock in the Valley and Ridge.

Silurian and Devonian (443 to 345 mya)

The Shawangunk Formation is about 1400 feet thick and it rests unconformably on the Martinsburg Formation. It consists of interlayered quartz-pebble conglomerate (fig. 5), and quartzite, rocks that are highly resistant to weathering and erosion. Subordinate interbeds of argillite also occur, especially in the middle part of the formation near Delaware Water Gap. Shawangunk sediments were laid down by braided streams on an alluviated coastal plain and along marine shorelines largely in tidal areas. The Shawangunk Formation forms Kittatinny Mountain, holding up the Mountain's main ridge, escarpment, and several high subordinate ridges.



Figure 11. Esopus Formation, Wallpack Ridge near Flatbrookville. The Esopus is chiefly a fine grained sandstone that with Oriskany Formation forms the higher areas on Wallpack Ridge. Bedding (dips right to left) is largely masked by a well-formed southeast dipping cleavage (dips left to right).

The Bloomsburg Red Beds are about 1500 feet thick, and rest conformably on the Shawangunk Formation. The Red Beds consist of repetitive packages of layered sandstone, shale, and siltstone (fig. 9). Locally the formation is conglomeratic near its base. Bloomsburg sediments were deposited under tidal and lagoonal conditions along the coast

of a large inland sea. Some of the rock layers contain mud cracks, small burrows, root traces, and evidence of soil formation. These features show that the Bloomsburg sediments were subaerially exposed and subjected to weathering during short periods when sea level lowered and the shoreline retreated seaward. The repetition of fining upwards cycles of sedimentation show sea level rose and fell often during deposition of the Bloomsburg Red Beds. Regionally, fossil shells and marine animal burrows in the formation also show a change in depositional environment proceeding from continental, to near shore, to shallow marine. The Bloomsburg Red Beds form a very broad outcrop belt along the west side of Kittatinny Mountain. This area consists of hills and ridges of moderate to steep relief. In places, the northwestward dip of the formation forms a long dip slope that extends downward toward Delaware and Wallpack Valleys.

Upper Silurian through Middle Devonian formations outcrop along a southwest-trending belt in the far western part of the Valley and Ridge. These formations, which represent several cycles of sea level rise and fall, conformably overlie each other, except for a few minor disconformities. In most places these rocks dip uniformly northwestward. Most of these formations are too thin to show on the small scale geologic map fig. 2b where instead they have been grouped based on similar lithology and land-forming characteristics. The lower and oldest group of formations includes the Poxono Island,



Figure 12. Highly jointed, thin to medium bedded nodular limestone in the Onondaga Formation, Delaware Valley.

Bossardville, Rondout, Decker, and the Helderberg Group. These rocks underlie Wallpack Valley and southeastern face of Wallpack Ridge. They largely consist of limestone mixed with varying proportions of clay and silt. In several places (fig. 10), banks of biogenic limestone and patchwork reefs are found. Lying above the limestone is a predominantly clastic sequence that consists of the Port Ewen Shale, Oriskany Group, Esopus and Schohaire Formations. These formations underlie the upper part of Wallpack Ridge with the upper part of the Oriskany and the Esopus Formation (fig. 11) forming the highest-standing ridges and hills. The Onondaga Limestone overlies the clastic rocks and forms a long dip slope on the northwest side of Wallpack Ridge that extends to the Delaware River. In places, sinkholes and springs are found along this slope. These features are products of dissolution and the formation's gentle northwest dip. Overlying the limestone is the Marcellus Shale, a weak shale that floors most of the Delaware Valley upstream from Wallpack Bend to Port Jervis, New York. It is rarely observed in outcrop because in most places it is buried by glacial outwash and alluvium.

YOUNGER ROCKS – (What happened to them?)

The Valley and Ridge's geologic record (fig. 4) reveals a substantial gap in time from about 390 million years ago to 125 thousand years ago. Since geologic formations of these missing ages occur elsewhere in New Jersey and in nearby states, we can assume that some the missing bedrock has been removed by erosion over the many millions of years since they were deposited. Also, because the land that now makes up the Valley and Ridge was at times uplifted and exposed during several major and many lesser tectonic events, some of the younger formations were never deposited.

TECTONIC HISTORY

Rock structure has also played an important role in the development of the Valley and Ridge's topography. Major fold axes and faults trend northeast to southwest, a direction that is coincident with the province's topographic grain. These folds and faults formed during several major mountain-building events that occurred when the North American plate collided with other lithospheric plates that now make up parts of North America, Europe and Western Africa. These deformational events called the Taconic and Alleghenian orogenies (fig. 4) took place over millions of years and were followed by longer periods

of quiescence. The Arcadian Orogeny, an important event in the northern Appalachian Mountains, did not affect New Jersey's Valley and Ridge formations.

Joints (fig. 13) are fractures with very little or no discernable offset. They are the products of tectonic compression or extension and they play a significant role in the development of geologic landscapes. Many joints in the Valley and Ridge cut across the topographic grain of the province. Some stream reaches are accordant with these cross-joints suggesting a relationship between stream course and zones of structural weakness. Joints also present avenues for weathering agents, both mechanical and chemical, providing for increased rates of erosion.

In addition to major mountain-building events, uplift along the continental margin also occurred during several lesser tectonic events during the Cenozoic Era. Uplift may also be attributed to the warping of the Earth's crust due to the tremendous amount of sediment deposited offshore.

PATHWAY TO THE PRESENT-DAY LANDSCAPE

No single point in time represents the birth of New Jersey's Valley and Ridge. Although its rocks are as much as 535 million years old, the topographic expression that defines its namesake came into being at a much younger time. Because streams are the dominant sculptor of the Valley and Ridge's topography, a good place to start a discussion on the province's geomorphic history would be the birth of its streams.

About 306 million years ago (fig. 4), North America collided with Gondwana (combined continents of Europe and Africa) forming part of the supercontinent Pangea. During this collision (named the Allegheny Orogeny) a massive mountain range known as the Appalachians was created. This mountain range, whose roots lie east of New Jersey,



Figure 13. Highly jointed Allentown Dolomite in Kittatinny Valley near Newton. Bedding dips from right to left while closely-spaced joints dip in the opposite direction. Joints typically exhibit preferred orientations with most being orthogonal to bedding. Joints, fractures and faults were formed during the Taconic and Alleghenian orogenies where the Lower Paleozoic Formations were subjected to compressional stress during the collision of continental plates.



Figure 14. Culvers Gap from Smith Hill, Kittatinny Valley.

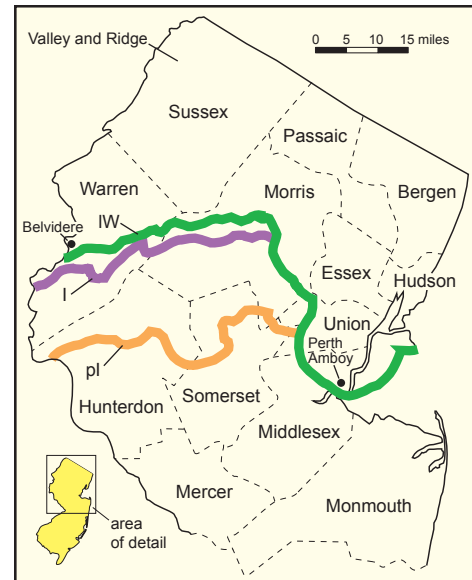
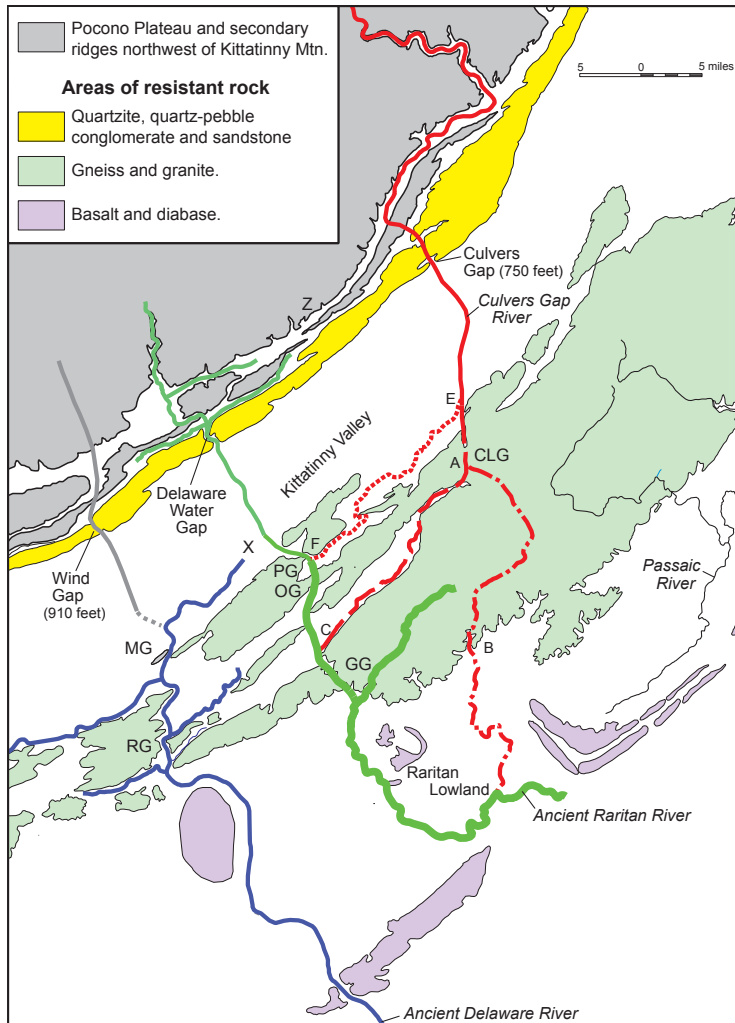


Figure 15, *left*. Reconstruction of the late course of the Culvers Gap River, and several scenarios for its capture by the ancient Delaware River. Key to map names abbreviated in figure : PG - Pequest Gap, OG - Oxford Gap, GG - Glen Gardner Gap, MG - Marble Mountain Gap, RG - Riegelsville Gaps, CLG - Cranberry Lake Gap. Pre-capture course of the Culvers Gap River: A-B Andover-Ledgewood course, A-C Andover-Musconetcong Valley Course, E-F Pequest Valley course. Location of capture: V - Pequest Valley capture, W - Wind Gap capture, X - Pequest Gap capture, Y - Paulins Kill valley capture, Z - Minisink Valley capture. Figure 16, *above*. Limits of glaciations in New Jersey. The trace of the IW limit generally marks the position of the Terminal Moraine. Key: IW - late Wisconsinan, I - Illinoian, pl - pre-Illinoian.

reached its maximum size sometime during the Permian Period. About 200 million years ago during Jurassic time (fig. 4) Pangea broke apart when the North American and African continents pulled away from each other. This resulted in the opening of the Atlantic Ocean. By Late Cretaceous time (99 - 65 mya), the Atlantic Ocean and New Jersey's coastal plain were well established. The Cretaceous formations of New Jersey record streams draining to the Atlantic Ocean. Prior to this streams flowed west toward the continental interior to a large inland sea.

From the time of the Alleghenian Orogeny, New Jersey's Valley and Ridge was an uplifted landmass subject to erosion. New streams associated with the opening of the Atlantic Ocean eventually cut back into the Valley and Ridge and over time sculpted the land into its present form. Eroded sediment derived from inland areas was transported to the Atlantic Ocean where it formed New Jersey's coastal plain and continental shelf.

The period between 65 and 20 million years ago was a relative time of tectonic quiescence in the Valley and Ridge. Streams slowly expanded their drainage networks inland and carried eroded sediment to the New Jersey Coastal Plain. Over time relief lessened as ridge tops and hills were eroded. The topographic expression of the land was much more muted than it is today because the land was covered by a thick mantle of weathered rock called regolith.

Renewed Period of Erosion and Incision

Starting about 23 million years ago, global sea level started to fall due to the growth of the Antarctic ice sheet (fig. 4). The coastal plain formations of New Jersey were most effected because they consist of unconsolidated sediments that were easily eroded and because they lie next to the falling sea. Farther



Figure 17. Compact sandy-silty till, Kittatinny Valley. Clasts chiefly derived from local rock formations (dolomite, shale, and sandstone). Divisions on scale equal one foot.

wind gap found along Kittatinny Mountain and it represents last great change in drainage in New Jersey's Valley and Ridge. The Culvers Gap River may have been part of the ancient-Raritan River drainage basin during the late stages of its history. Before its abandonment of Culvers Gap, the river followed a course through Kittatinny Valley and crossed the New Jersey Highlands to the Raritan lowland (fig. 15). The demise of the Culvers Gap River appears to have resulted from stream capture during the Late Miocene or Early Pliocene (8 to 3 mya). Possibly driven by sea level lowering caused by the growth of the Antarctic ice sheet, the Delaware River or one of its earlier tributaries captured the Culvers Gap River behind Kittatinny Mountain. Apparently, the narrow width and structural weakness of resistant rocks along the Delaware River where it crossed the New Jersey Highlands, gave it an advantage over the Culvers Gap River and its more northerly course through the Highlands.

The hypothetical course area of the Culvers Gap River across Kittatinny Valley is now the drainage divide between the Hudson and Delaware Rivers. This is the result of drainage shifting southwest to the present course of the Delaware River and the subsequent longitudinal adjustment (headward erosion) of streams (Paulins Kill and Pequest) chiefly along belts of carbonate rock.

GLACIAL HISTORY – (The Ice Age)

Over the last two million years continental glaciers, originating in the subarctic regions of Canada, reached the Valley and Ridge Province at least three times (fig. 16). During each glaciation, bedrock ridges, hills and slopes were worn down by the action of moving glacial ice, and valley floors were deeply scoured. Eroded rock debris and soil entrained by the ice sheet was deposited as till (fig. 17), an unsorted mixture of clay- to boulder-sized material. This material was laid down from the glacier on the bedrock surface in sheets, in streamlined hills called drumlins, and in ridges laid down along the former edge of the ice sheet called moraines. Stratified sediment (fig. 18) consisting of a sorted and layered gravel, sand, and silt was deposited by glacial meltwater streams in valleys that

inland the land that makes up the Valley and Ridge province also felt the effects of the changing seas. In response to lowering sea level, streams cut down in bedrock as they adjusted to a lowering base level. Correspondingly, slope retreat accelerated, erosional rates increased, and topographic relief became greater. Geologists estimate that 1100 meters of land has been eroded since the Middle Miocene (16 to 11 mya). If this rate is correct then the present shape of the Valley and Ridge is of a young *geologic* age.

Relict Landforms

Wind gaps are places where rivers used to flow. Cut through ridges of resistant rock, they offer a glimpse of geologic past, enigmatic landforms that represent an earlier time.

Culvers Gap (figs. 2d and 14) is the lowest



Figure 18. Stratified, planar to cross-bedded gravel and sand deposited by a glacial meltwater stream during the last ice age about 20,000 years ago. Foul Rift, Delaware Valley. Divisions on scale equal one foot.

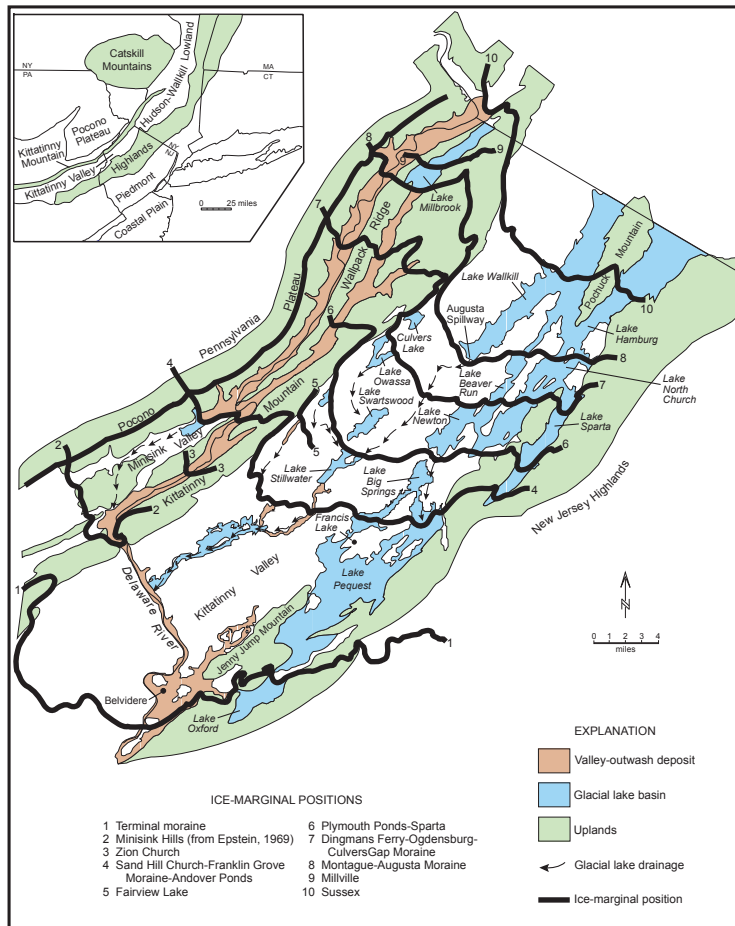


Figure 19. Late Wisconsin ice margins of the Kittatinny and Minisink Valley ice lobes, and location of large glacial lakes, extensive valley-outwash deposits.

geometry of the ice sheet's margin and resulted in regional ice flow turning to the southwest, a direction more in line with the topographic grain of the province. Accordingly, deglaciation was characterized by the systematic northeastward retreat of the Kittatinny Valley and Minisink Valley ice lobes (fig. 19). This interpretation is based on the distribution of ice-marginal meltwater deposits and moraines, and correlative relationships between elevations of delta topset-foreset contacts, former glacial-lake-water plains, and lake spillways. During glacial retreat, meltwater sediment was chiefly laid down in glacial lakes (fig. 19) that occupied valleys now drained by the Pequest River, Paulins Kill, and Wallkill River, and to a lesser extent in small upland basins and valleys. These former lake basins were dammed by stratified drift, moraine, and stagnant blocks of ice, or by the glacier's margin.

Glaciations change or modify surface drainage by erosion of the land and deposition of glacial sediment. Continental ice sheets deeply scoured and widened valleys that were oriented along lines of glacial flow. Due to glacial erosion the bedrock floor of the Delaware Valley lies as much as 200 feet beneath the Delaware River. In places, glacial meltwater streams and glacial-lake drainage carved and rerouted channels in preglacial valleys and also eroded new channels in bedrock and surficial materials in areas where none had previously existed. Also, some preglacial valleys were buried by glacial lake deposits and till. These buried valleys have been only discovered by mapping out the elevation of the rock surface using geologic records from public and private wells, exploratory borings, and remote sensing methods. Meltwater deposits and end moraines laid down in stream valleys may also modify surface drainage through damming and reversing stream gradients, especially in the upper parts or heads of drainage basins. The barbed drainage pattern of Clove Brook (fig. 2a) suggests that at one time it flowed southwestward to the Paulins Kill. Today it flows to the Wallkill River, its change in course due to the drainage modifications caused by glacial drift deposited in the upper part of the Paulins Kill valley during the last ice age.

drained away from the glacier and in glacial lakes that formed where the flow of streams had become blocked by glacial drift or ice.

The most recent ice sheet reached New Jersey during the late Wisconsin glacial about 22,000 to 20,000 years ago. Its deposits provide the clearest record of glaciation. Older glacial deposits were largely eroded or they lie buried beneath the younger glacial sediment.

The farthest advance of the late Wisconsin ice sheet in most places is marked by a terminal moraine (fig. 19). In the Delaware and Pequest valleys till and ice-contact glacial outwash show that terminus of the ice sheet did extend about 0.5 miles south of the moraine. The terminal moraine was deposited over a period of about a 1000 to 1500 years. It represents a time when the ice sheet's margin remained in a constant position neither retreating or moving forward except within the narrow zone marked by the moraine. During deglaciation, the outer part of the ice sheet thinned and its flow, which was largely independent of topography, became much more constrained by the northeast-to-southeast orientation of the large valleys. This changed the



Figures 20. Talus apron on the southeast facing flank of Kittatinny Mountain, Kittatinny Valley. Talus here consists of quartzite and conglomerate joint blocks that were dislodged from Shawangunk outcrops located upslope.

POSTGLACIAL HISTORY

Since the last glacier receded from New Jersey about 17,000 years ago the climate has become increasingly warmer. In many places, swamps and bogs have formed where shallow lakes and ponds have become filled with decayed vegetation and detritus. Rock outcrops have been fractured by the effects of freeze and thaw, their dislodged pieces forming aprons of rocky debris at their bases. Postglacial streams have further cut down in glacial valley-fill materials, widening their valley floors and depositing flood plains.

Initially, cold and wet conditions, and absent to sparse vegetative cover, enhanced erosion of hillslope material by solifluction, soil creep, and slope wash. Mechanical disintegration of rock outcrops by freeze-thaw also provided additional sediment. Some of

this material forms extensive aprons of talus at the base of cliffs on Kittatinny Mountain (fig. 20). A few small boulder fields were formed where boulders, transported downslope by creep, had accumulated at the base of hillslopes and in first-order drainage basins. These fields and other boulder concentrations formed by glacial transport and meltwater erosion, were further modified by freeze and thaw; their stones in places reoriented to form crudely-shaped stone circles. Gradually as the climate warmed, vegetation spread and was succeeded by types that further limited erosion.

The many swamps (fig. 21) and poorly drained areas in the Valley and Ridge are typical of glaciated landscapes. Upon deglaciation, surface water, which had in preglacial time flowed in a well defined network of streams, became trapped in the many depressions, glacial lakes and ponds, and poorly-drained areas that made up the glacial landscape. Between 14,000 and 11,000 years ago, relatively barren lake and pond sediments, which largely consisted of weathered rock and soil washed in from surrounding higher ground, became enriched with organic material. This transition represents a warming of the climate where it became possible for aquatic vegetation to thrive. In addition, the landscape changed from tundra to a mix of small expanses of spruce and hemlock and open land populated by shrubs and grasses. Eventually a closed boreal forest of conifers covered the area. About 10,000 years ago at the start of the Holocene (fig. 4), oak and other hardwoods began to populate the landscape, eventually displacing the conifers. Throughout the Holocene the many shallow lakes and ponds left over from the ice age slowly filled with decayed vegetation eventually forming bogs and swamps. These organic-rich deposits principally consist of peat, muck, and minor rock and mineral fragments. Ponds in areas of karst may also have become filled with marl, which is calcium carbonate precipitated by some aquatic plants.

Swamps and bogs contain sedimentary and organic deposits that can be used to reconstruct past climatic conditions. Because these materials were laid down layer upon layer, they may preserve a climatic record from the time of deglaciation to the present. The identification of pollen and radiocarbon dating of plant and animal material retrieved from swamps by coring provides stratigraphic control on regional and local changes in vegetation, which can be used as a proxy for climatic change. Several studies on bogs and swamps in northwestern New Jersey and northeastern Pennsylvania have established a



Figure 21. "Our Swamp", High Point State Park, Kittatinny Mountain.

dated pollen stratigraphy that nearly goes back to the onset of deglaciation. Paleoenvironments, interpreted from pollen analysis, show a transition from tundra with sparse vegetal cover, to open parkland of sedge and grass with scattered arboreal stands that largely consisted of spruce. During the period from about 14,250 to 11,250 years ago the regional pollen record shows the transition to a dense closed boreal forest that consisted of spruce and fir blanketing uplands. This was followed by a period (11,250 to 9,700 years ago) where pine became the dominant forest component. These changes in vegetation record the continued warming during the latter part of the Pleistocene and transition from the ice age to a temperate climate. About 9,400 years ago oak became the dominant tree displacing the conifers and marking the transition from a boreal forest to a mixed-hardwoods temperate forest. From about 9000 to 5000 years ago during the early to middle part of the Holocene temperatures in New Jersey were slightly warmer than at present. A slight cooling trend occurred from 5000 to 2000 years ago, which has been followed by an overall increase in temperature over the last 2000 years.

* Preceding article from *Geologic History of New Jersey's Valley and Ridge*, New Jersey Geological Survey Information Circular, 2007, Witte, R.W. and Monteverde, D H.

LATE WISCONSINAN DEGLACIATION AND POSTGLACIAL HISTORY OF MINISINK VALLEY: DELAWARE WATER GAP TO PORT JERVIS, NEW YORK

by

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INTRODUCTION

Several investigations on glaciofluvial terraces in Minisink Valley (fig. 1) have suggested that the late Wisconsinan ice sheet disappeared either by regional stagnation or marginal retreat. This disagreement is a recurring controversy that is not unique to the Minisink Valley area. The earliest researchers (White, 1882 and Salisbury, 1902) favored a marginal retreat model. Their interpretations were largely based on the identification of recessional moraines, and the ice-contact heads of valley-train deposits that represented positions where the retreating glacier margin had halted. Later work by Happ (1938) and Crowl (1971) favored a stagnation model where the uplands were deglaciated first leaving residual masses of ice in the valleys. Large areas of collapsed topography in kames and kame terraces, many kettles, ice-contact slopes, and unpaired terraces were cited as evidence for stagnation. Epstein (1969) and Epstein and Koteff (2002) near Stroudsburg, Pennsylvania, and Ridge (1983), Witte, (1988, 1991, 1997a), and Stone and others (in press) in northwestern New Jersey have returned to the marginal retreat model. Based largely on the morphosequence model of Koteff and Pessl (1981), these researchers have developed a morphostratigraphic framework in northwestern New Jersey and parts of northeastern Pennsylvania that show deglaciation took place largely by the systematic melting back of the margins of the Kittatinny and Minisink Valley ice lobes.

Postglacial alluvial terraces in Minisink Valley have also been the focus of many investigations. Based on their continuity and uniform height above the Delaware River, and occurrence of paleopedologic marker horizons, they are divided into an upper and lower terrace. These morphostratigraphic units are abandoned flood plains and they record the change in the fluvial regime of the Delaware River as it evolved from a braided meltwater-fed stream to its present incised, low-sinuuous meandering form. Archeological studies in the lower terrace have produced cultural artifacts that date from late Pleistocene through to Historic time. Radiocarbon dating, microfaunal analyzes, and paleopedologic studies have produced an extensive amount of data, and have established a baseline for paleoenvironmental change during the Holocene Epoch. The time following deglaciation until the end of the Pleistocene is less understood, because most of the work in Minisink Valley was driven by archaeological interests where efforts were concentrated on Holocene alluvial sequences that contained evidence of Amerind culture. However, nearby palynologic studies on bog-bottom sediments have been used to establish a baseline for paleoclimatic change from a time shortly after deglaciation to the present. Of particular interest is the response of the Delaware River to 1) cessation of its meltwater supply, 2) changes in sediment loads due to floristic evolution in the drainage basin, and 3) effects of delayed isostatic rebound.

PREVIOUS INVESTIGATIONS

Glaciofluvial terraces in Minisink Valley were first discussed by Cook (1880, p. 74-75) in an Annual Report to the State Geologist. He stated "the modified drift bordering the Delaware forms terraces or gravelly and sandy shelves and flats from the hillsides down to the present flood plain." Several outwash terraces were described near Dingmans Ferry, Milford, and Port Jervis, and a lower and abandoned flood plain. Shortly afterwards, White (1882) reported on the glacial geology of Pike and Monroe Counties, Pennsylvania, and described up to five levels of terraces in Minisink Valley. The lowest terrace, which rose as much as 25 feet above the river, was considered a flood plain of postglacial age. The higher terraces were made of reworked drift, and they were laid down by meltwater that accompanied the retreat of the northern ice cap.

A voluminous report by Salisbury (1902) detailed the glacial geology of New Jersey region by region. The Terminal Moraine (fig. 2) and all glacial deposits north of it were interpreted to be products of a single glaciation of Wisconsinan age. Deglaciation largely took place by the glacier melting at its margin, and

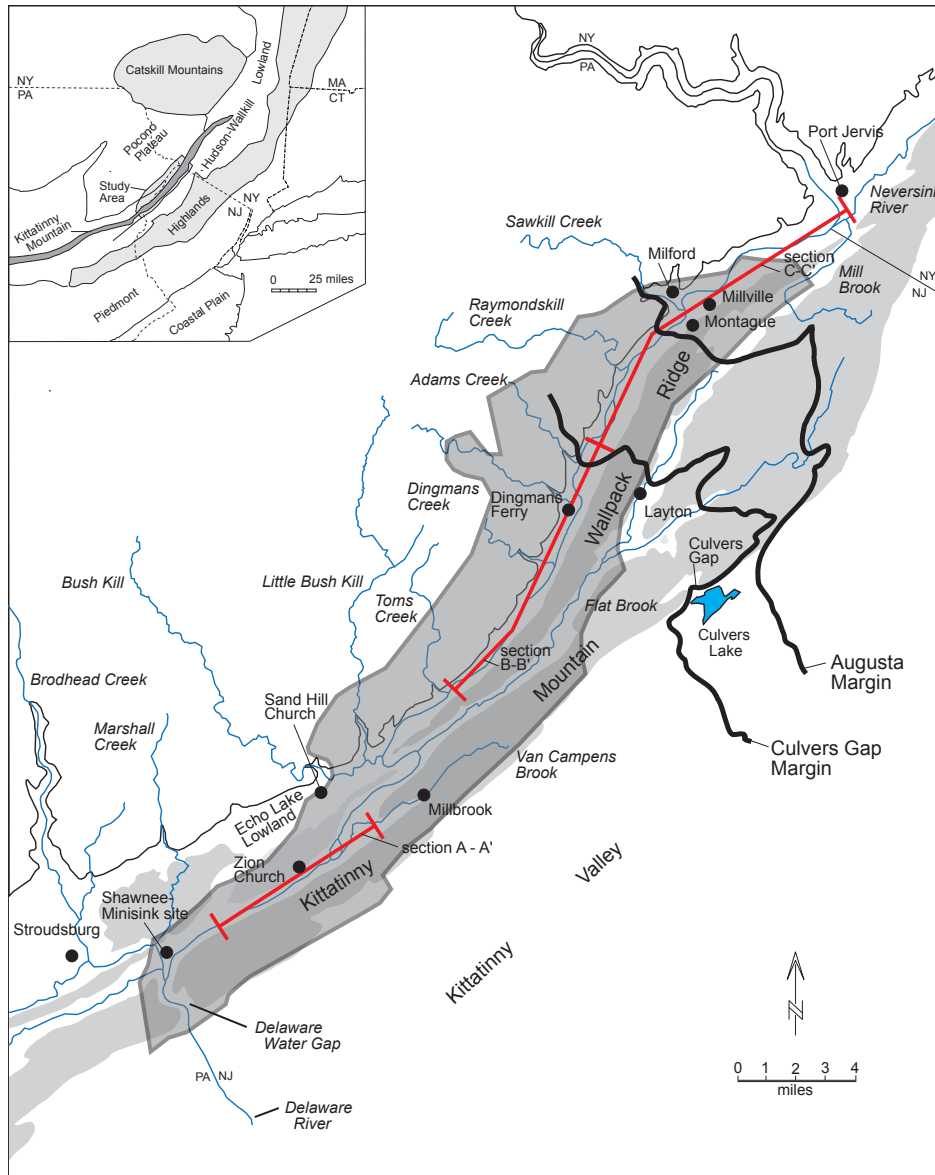


Figure 1. Physiography of Minisink Valley and surrounding area, the location of towns, streams and lakes named in text, and position of longitudinal profiles shown on Figures 7a, b, and c.

retreating unevenly in a northward direction. In Minisink Valley Salisbury (1902, p. 285-286) reported "The stratified drift of the Delaware Valley north of the moraine, is disposed in the form of terraces. The material is of glacial origin and was deposited by the waters arising from the melting of the ice." "The surface of the uppermost terrace at most points represents the depositional surface developed while the valley was being aggraded, during decadence of the ice sheet. Much of the filling has since been removed by the stream, which is still engaged in cleaning out the deposits * * *. The highest terraces represent the remnants of the old aggradation plain. The notably discordant levels of the original aggradation surface, and its failure to decline regularly to the southward, show that the gravel and sand were not deposited continuously from the State line to Belvidere. Rather, they were deposited in sections." Salisbury suggested that the high gravel terraces near Dingmans Ferry, Montague, and the State line, and the recessional moraine near Dingmans Ferry (Fisher School House) represented extended halts in the retreat of the glacier margin. Stagnation was of local extent only, and developed at or near the glacier margin as evidenced by hummocky topography and kettles. The lower terraces along the Delaware were considered a flood plain and remnants of an abandoned flood plain.

Happ (1938) suggested that the stratified deposits in Minisink Valley are kame terraces and delta terraces laid down over and against a small and thin body of ice that covered most of the valley's floor. In places small glacial lakes were formed between the residual ice and reentrants along the valley wall, where tributary streams entered the trunk valley. The larger delta terraces at Milford and Port Jervis may have also been laid down in a larger lake dammed down the valley by a recessional moraine. As to the nature of the ice in Minisink Valley, Happ (1938, p. 438) stated "There does not appear to be any positive evidence to show whether this ice in the valley bottom was an attenuated tongue of live ice, or whether it consisted of isolated and stagnant masses * * *."

Crowl (1971) produced a 1:24,000-scale surficial geologic map of Minisink Valley between Shawnee on Delaware and Matamoras, Pennsylvania, and included detailed observations on its glacial drift and history. Deglaciation, based on nearby palynological studies on bog-bottom deposits, and estimates on the melting rate of ice blocks, started sometime before 15,000 yrs B.P. Kames and kame terraces in the valley show that the ice had disappeared from this area by stagnation and downmelting with ice in uplands melting first. Crowl cited unpaired terraces, collapsed topography, and the position of the highest terraces

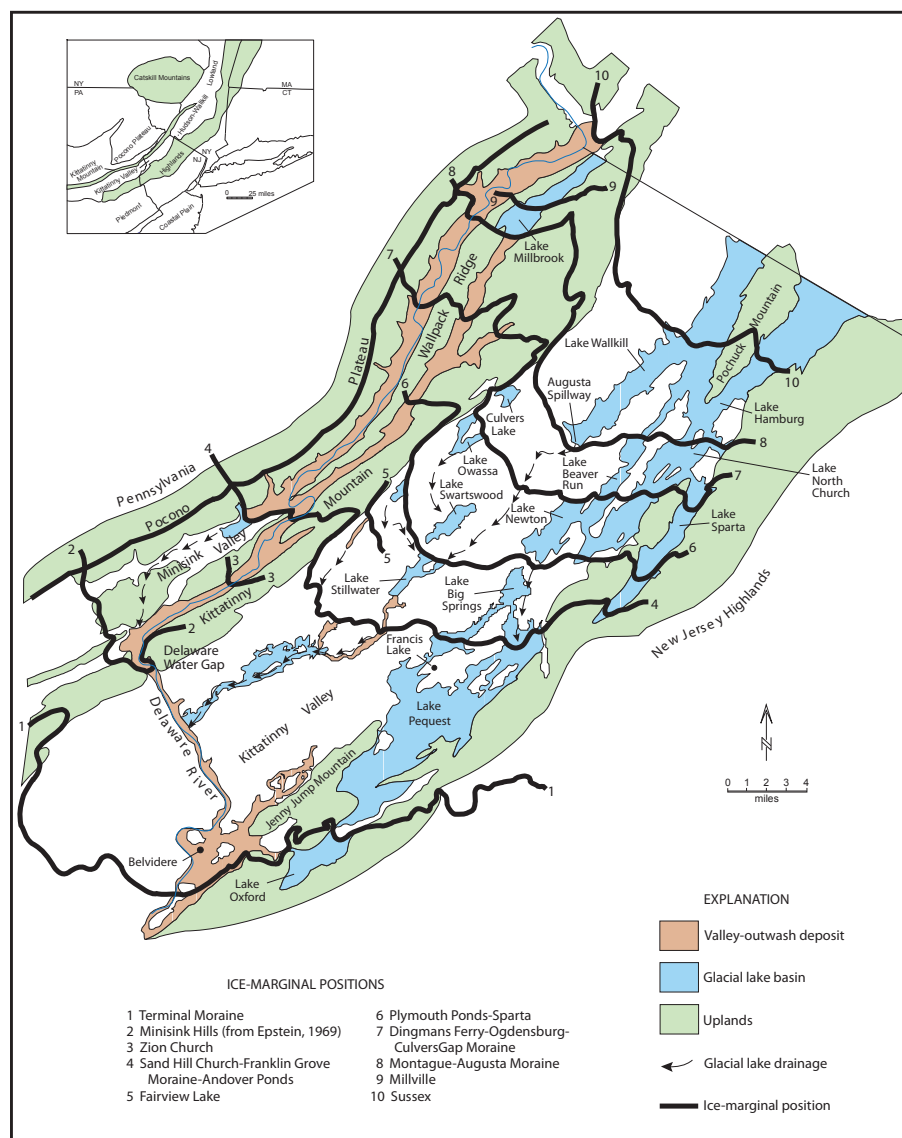


Figure 2. Late Wisconsinan ice-marginal positions in Minisink Valley and the upper part of Kittatinny Valley, location of large glacial lakes, and extensive valley-train deposits in northwestern New Jersey, and northeastern Pennsylvania. Data modified from Epstein (1969), Crowl (1971), Ridge (1983), Witte (1988, 1997a), and Stone and others (2002).

near reentrants along the valley wall, as proof that they appear to be separate entities and not parts of a dissected valley train started at a former glacier margin upstream.”

Epstein (1969), Witte (1997a), Witte and Epstein (2004), Witte and Epstein (in review), and Stone and others (2002) showed that the Minisink Valley lobe retreated in a northeasterly direction by melting at its periphery, and chiefly by a process of stagnation-zone retreat. A similar view was also held by Ridge (1983), and Witte (1988, 1997a) for Kittatinny Valley. Five ice margins, the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex, have been identified, and they delineate major recessional positions of the Kittatinny Valley and Minisink Valley lobes (fig. 2). The Culvers Gap, Augusta, and Sussex margins are traceable across the New Jersey Highlands into the Newark Basin (Stone and others, 2002), and all margins except the Sparta margin are traceable westward across Kittatinny Mountain into Minisink Valley. Additionally, readvances are marked by the Ogdensburg-Culvers Gap and Augusta moraines. The strong evidence of systematic deglaciation, and the presence of at least two readvances, suggests that regional or valley-ice tongue stagnation was not a valid style of deglaciation for Kittatinny Valley and Minisink Valley.

Archaeologic and paleopedologic studies in Minisink Valley have provided an enormous amount of information on its paleoenvironment. Several investigations on deep alluvial sequences have accorded scientists with a nearly complete record of fluvial deposition, land stability, and cultural evolution throughout the Holocene. Important to this study are the thick flood-plain sequences exhumed at the Shawnee-Minisink site (McNett and others, 1977), and upper Shawnee Island site (Stewart, 1989). In addition, nearby palynologic studies of bog- and lake-bottom sediments have provided corollary information on floristic changes in the Minisink Valley area. Together, these studies documented the paleoenvironmental record during the late Quaternary, providing a baseline for future studies.

In the study area Stewart (1989, p. 102) noted that “sedimentary sequences representing habitable landforms generally do not predate 8000/7000 BC implying that the Delaware River did not enter its present channel until after this time.” These terraces are 15 to 25 feet above the river, and they include buried-A horizons, three of which are distinctive chronostratigraphic units, correlated to the Late Terminal/ Archaic, late Middle Woodland, and Late Woodland cultural periods. Based on the degree of pedogenic formation Foss (1989) suggested that these paleosols reflect extended periods of landscape stability of about 500 to 1000 years.

PHYSIOGRAPHY AND BEDROCK GEOLOGY

Minisink Valley (fig.1) is a narrow, deeply trenched lowland underlain by Silurian and Devonian strata. It lies in the glaciated Valley and Ridge province between Pocono Plateau and Kittatinny Mountain, and it runs just north of Port Jervis, New York southwestward toward Delaware Water Gap. Its name does not appear on U.S. Geological Survey topographical maps, but it is defined in Lipponcotts Gazetteer (1931) as “An Indian name for part of the valley of the upper Delaware River, beginning a short distance above Delaware Water Gap, Pa.” The valley was also the former site of a hydroelectric and water storage project by the Army Corps of Engineers. A dam constructed at Tocks Island would have flooded the valley upstream to Port Jervis, New York, and provided a storage capacity of 133.6 billion gallons (Corps of Engineers, 1967). This project has since been de-authorized by the U.S. Congress.

Bedrock in the Minisink Valley area (fig. 3) consists of Silurian and Devonian strata that dip northwest and form a southwest-trending homocline (Drake and others, 1996; Sevon and others, 1989). The Delaware River (fig. 1) enters Minisink Valley at Port Jervis, New York and Matamoras, Pennsylvania where it is joined by the Neversink River. From here, it makes a sweeping right-hand turn, flowing southwestward through Minisink Valley toward Wallpack Bend. Throughout this stretch, Minisink Valley decreases in width from approximately 1.25 to 0.75 miles (2.0 - 1.2 km) as it follows the strike of the weaker limestone and shale formations. The western side of the valley is marked by a high cliff, and the narrow upland of rugged relief that lies above the cliff forms the northwestern border of the Valley and Ridge province. Further to the northwest is the Pocono Plateau, which is part of the Appalachian Plateaus province. At Wallpack Bend, the river follows a large meander through Wallpack Ridge, abandoning the strike valley of its upper part, which continues southwestward into the Echo Lake Lowland. On the east side of Wallpack Ridge, the river turns back to the southwest following the strike of weaker limestone strata toward the Delaware Water Gap. Along this stretch, Minisink Valley is generally less than 0.5 miles (0.8 km) wide and it forms a narrow deep trench within the confines of Wallpack Ridge and Kittatinny Mountain. Wallpack Valley is the northeastward continuation of the lower part of Minisink Valley.

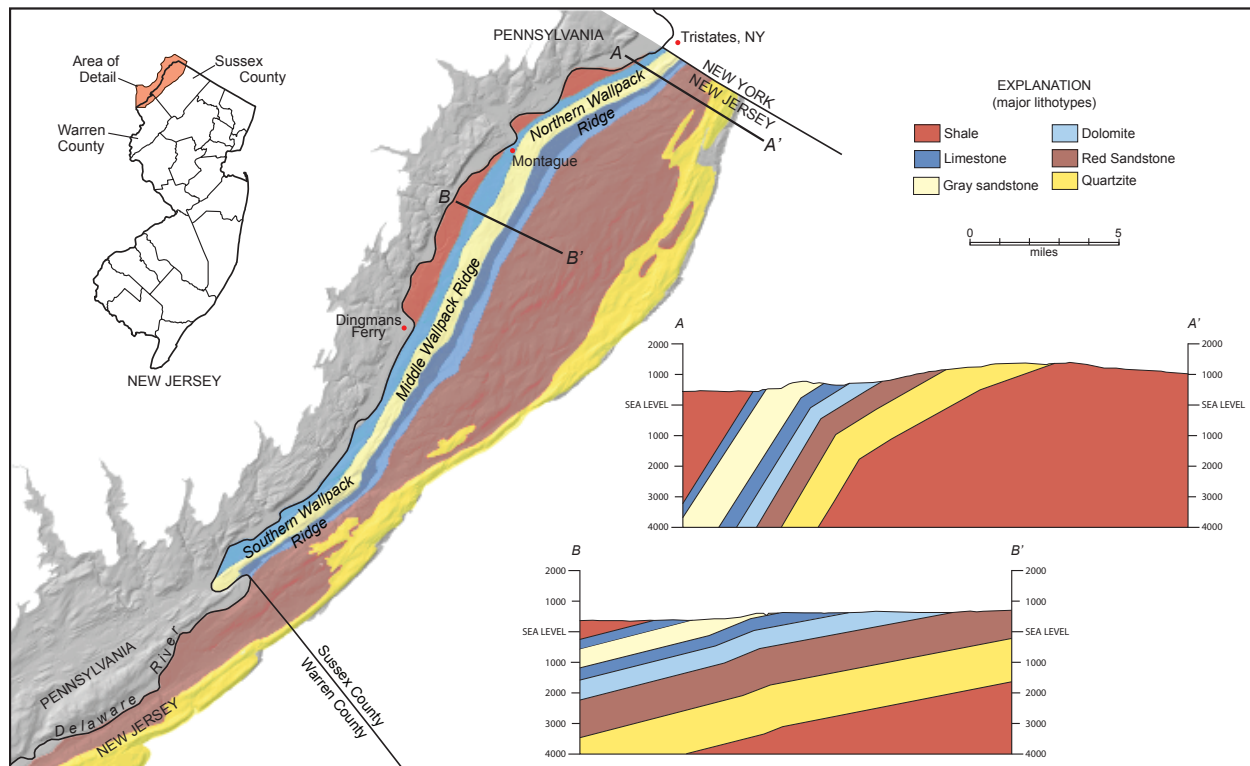


Figure 3. Simplified geologic map of Kittatinny Mountain -- Minisink Valley area (modified from Drake and others, 1996). Cross-sections are not drawn to scale.

Kittatinny Mountain is a prominent ridge that forms the eastern border of the study area. It separates Minisink Valley from Kittatinny Valley, and it runs from the Shawangunk Mountains in New York southwestward through New Jersey into Pennsylvania. It rises as much as 1500 feet (457 m) above the floor of Minisink Valley and it is held up by a very resistant quartzite and quartz-pebble conglomerate. The lower area northwest of the mountain that extends to Wallpack Valley is included with Kittatinny Mountain.

GLACIAL DEPOSITS

Glacial materials in Minisink Valley consist of till and meltwater sediment deposited during the late Wisconsinan glaciation. Collectively they may be as much as 250 feet (76 m) thick and they are correlative with the Olean Drift of northeastern Pennsylvania (Crowl and Sevon, 1980). Meltwater deposits consist of valley-train, outwash-fan, and meltwater-terrace deposits that were laid down at and beyond the margin of the Minisink Valley lobe. The heads of outwash of the valley-train deposits, and the Dingmans Ferry, Montague, and Millville moraines mark retreat positions of the Minisink Valley ice lobe.

Till

Till typically covers the bedrock surface and it is distributed widely throughout the Minisink Valley area. It is generally less than 20 feet (6 m) thick, and its surface expression is mostly controlled by the contour of the underlying bedrock surface. In many places bedrock outcrops, which show evidence of glacial erosion, extend through this cover. Thicker, more continuous till subdues bedrock irregularities, and in places completely masks them, and very thick till makes up drumlins, aprons on north-facing hillslopes, recessional moraine, and ground moraine. It also fills narrow preglacial valleys; especially those oriented transversely to glacier flow.

Till is a compact sandy silt to silty sand containing as much as 20 percent pebbles, cobbles, and boulders. Its provenance is local, and represented by a varying mix of the Silurian and Devonian rock formations. Clasts are typically subrounded, faceted, and striated, and measured clast fabrics show a preferred long axis orientation that is generally parallel to the direction of glacier flow. Presumably, this material is lodgement till. Overlying this lower compact till is a thin, discontinuous, noncompact, poorly

sorted silty sand to sand containing as much as 35 percent pebbles, cobbles, boulders, and interlayered with lenses of sorted sand, gravel, and silt. Overall, clasts are more angular, and clast fabrics lack a preferred orientation or have a weak orientation that is oblique to the direction of glacier flow. This material may be ablation till and flowtill, but it has not been mapped separately due to its scant distribution. In addition, cryoturbation, bioturbation, and mass wasting have altered the upper few feet of till, making it less compact, reorienting stone fabrics, and sorting clasts.

Moraines

End moraines in northwestern New Jersey are prominent, segmented, arcuate belts of hummocky till that cross Kittatinny and Minisink Valleys, Kittatinny Mountain and New Jersey Highlands (Witte, 2001a). They include the Terminal Moraine and several recessional moraines deposited 21,000 to 18,000 years ago during the late Wisconsinan substage of the Wisconsinan glacial stage. They all consist of a complex assemblage of small-scale landforms that collectively define areas of ridge-and-trough and knob-and-kettle topography (fig. 4). Their lobate course, till composition, and preferred development of ridge-and-trough topography along their outer margins show they were initially constructed at the margin of an active glacier.

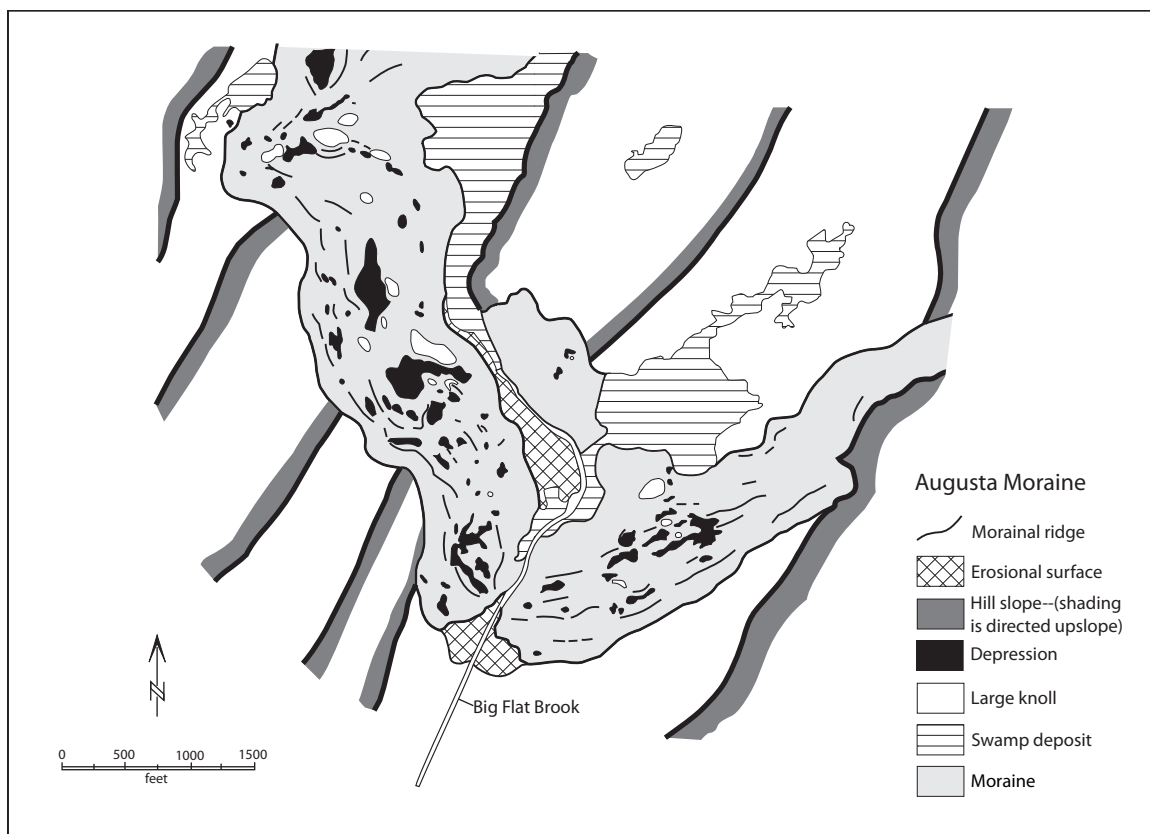


Figure 4. Morphology of the Augusta Moraine where it crosses Big Flat Brook Valley, Kittatinny Mountain, Sussex County, New Jersey. Morainial landform elements collectively define areas of ridge-and-trough, and knob-and-kettle topography.

The Terminal Moraine and some larger end moraines were also laid down following a readvance, further evidence of their association with active ice. Although end moraines were initially constructed at active glacier margins, their final form is largely a product of stagnation. Apparently, the glacier's terminus became buried by its own debris, which resulted in the formation of a narrow zone of dead marginal ice. Except for moraine-parallel ridges, which may be either push ridges or colluvial ramparts, morainial topographic elements were largely formed through topographic inversion after a complex history of collapse, due to melting of buried ice and resedimentation of supraglacial debris by mass wasting.

Morainial deposits in Minisink Valley include the Dingmans Ferry, Montague, and Millville moraines (fig. 2). The Dingmans Ferry moraine, originally called the "Fisher School House" moraine by Salisbury (1902),

traces a lobate course off Kittatinny Mountain, across Wallpack Valley and Wallpack Ridge, and into Minisink Valley where it abruptly ends. The Montague moraine traces a similarly parallel course as the Dingmans Ferry moraine. In Wallpack Valley, it splits into two distinct ridges. From here, it continues across Wallpack Ridge into the Minisink Valley where it ends near the village of Montague. The smaller Millville moraine only lies in Minisink Valley and on Wallpack Ridge. The Dingmans Ferry and Montague moraines mark major ice-retreat positions of the Minisink Valley lobe. They are coeval with the Ogdensburg-Culvers Gap and Augusta moraines that lie to the east and were formed at the margin of the Kittatinny Valley lobe (Witte, 1997a). As previously indicated by Crowl (1971), these recessional moraines have not been observed in Pennsylvania. Although, they may be correlative with ice-contact outwash, mapped by Sevon and others (1989), situated further west on the Pocono Plateau. Presumably, the moraines in Minisink Valley have been partly eroded by meltwater. The scant distribution of till immediately west of Minisink Valley may have also negated moraine formation there. In New Jersey, the recessional moraines are typically larger and more continuous in areas where thick till is near or next to the proximal side of the moraine.

Well-record data in Kittatinny Valley show the Ogdensburg-Culvers Gap and Augusta moraines were laid down following a readvance of the Kittatinny Valley lobe (Witte, 1997a). Although, subsurface data in Minisink Valley is inconclusive, it appears the moraines may have also been laid down following a readvance given their stratigraphic position, thickness of stratified material near the moraine, and correlation with the moraines in Kittatinny Valley.

Deposits of Glacial Meltwater Streams

Sediment carried by glacial meltwater streams in Minisink Valley was chiefly laid down in valley-train deposits and outwash-fan deposits at and beyond the glacier margin. Smaller quantities of sediment were also deposited in meltwater-terrace deposits, and in a few kame terraces and kames. The position of these deposits on the landscape is shown in figure 5.

Most of the higher terraces in Minisink Valley are the remnants of at least four extensive valley trains laid down at the margin of the Minisink Valley lobe (fig. 2). From oldest to youngest they are named Zion Church, Dingmans Ferry, Montague, and Tristates. These outwash remnants form discontinuous, narrow to broad terraces that are typically attached to a valley wall. They have flat surfaces that slope gently down valley, and steep-sided fluvio-erosional escarpments that lie against the younger meltwater-terrace,

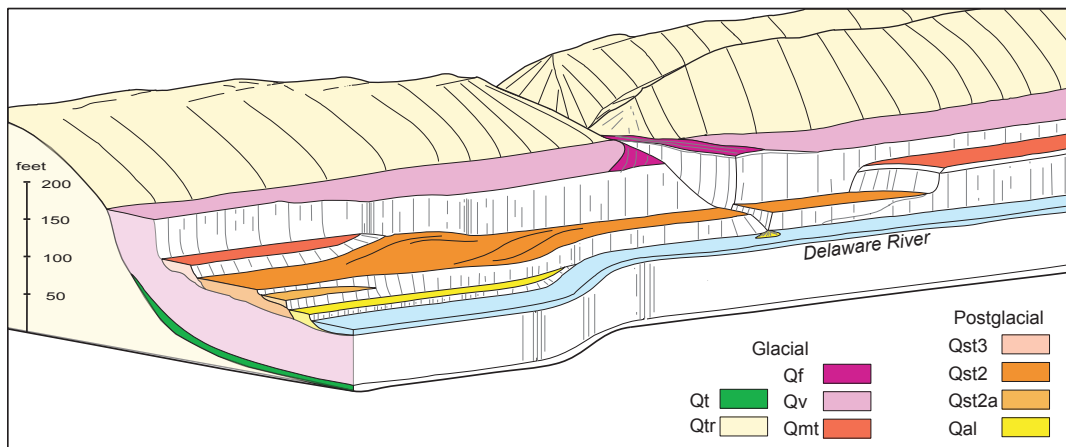


Figure 5. Block diagram illustrating spatial relationships between fluvial landforms in Minisink Valley. Glacial deposits are late Wisconsinan age, laid down in Minisink Valley during retreat of the Minisink Valley ice lobe. Postglacial terraces are of late Wisconsinan and Holocene age. The formation of alluvial terraces was in response to the cessation of meltwater flow due to deglaciation of the Delaware River basin and reduction in sediment supply as vegetation was re-established during postglacial climatic warming. Key to units: Qt - thick till, Qtr - thin till and bedrock, Qf - outwash-fan deposit, Qv - valley-outwash deposit, Qmt - meltwater-terrace deposit, Qst - stream-terrace deposit and Qal - alluvium.

and alluvial-terrace deposits that cover the lower parts of the valley floor. Near their heads of outwash, collapsed topography and kettles indicate deposition over and against small blocks of stagnant ice. Most of the terrace scarps have been modified by fluvial erosion. Therefore, deciding whether the outwash was

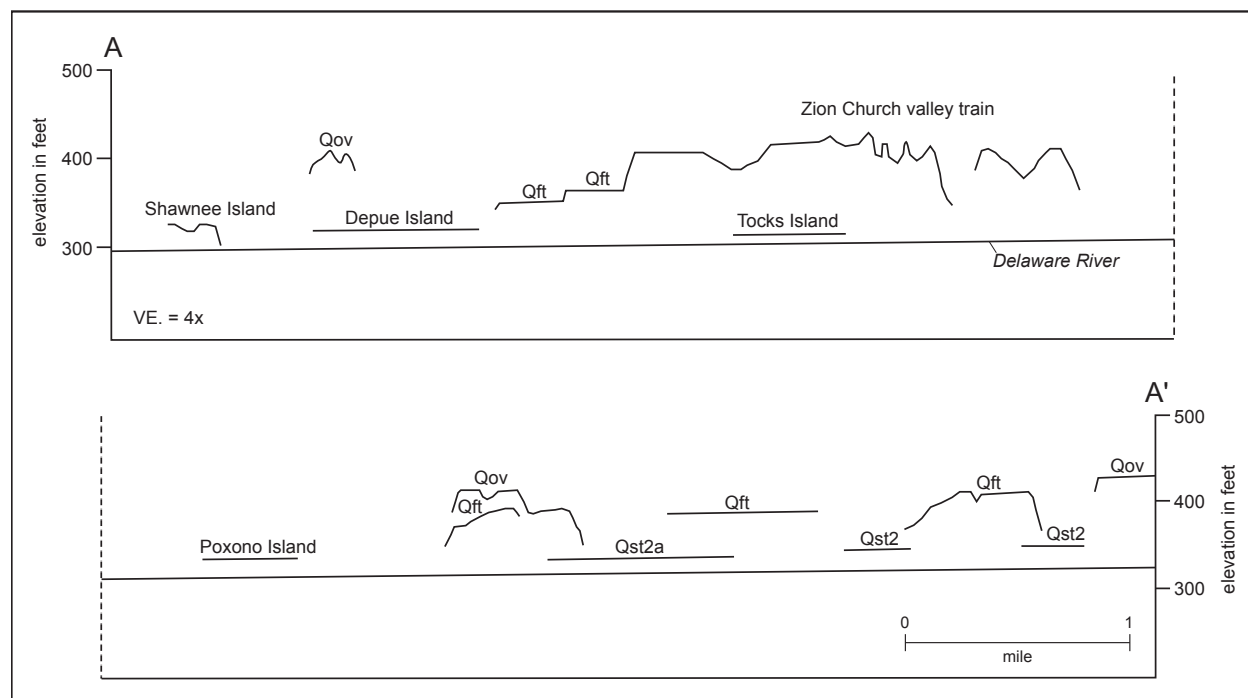
laid down against large blocks of residual ice is problematic. From their upstream to downstream parts the texture of the valley train's decreases from boulder-cobble-gravel to cobble-pebble gravel (fig. 6). In places, they are covered by several feet of wind-blown sand. Outcrops show that the bulk of the gravelly material is planar-bedded and normally graded, and gravel clasts in places exhibit imbrication. The base of many gravel beds is often marked by a thin layer of larger clasts. In places, the gravel is interlayered with thin, lenticular layers of cross-bedded sand. Based on their projected longitudinal profiles (figs. 7a, b, c), and a decrease in grain size downstream, the outwash appears to have been laid down at and beyond the margin of the Minisink Valley lobe, rather than in separate ice-walled depressions between blocks of residual ice.

Records of wells in Minisink Valley (Depman and Parillo, 1969; Witte and Stanford, 1995; Witte, 1997a and 2004) show that silt, very fine sand, and clay lie beneath the coarse gravel and sand of parts of the valley-train deposits. These materials may have been laid down in proglacial lakes that formed between the margin of the Minisink Valley lobe and moraine or outwash down valley, or in large depressions on the deeply scoured valley floor. In other places, as suggested by Happ (1938), this fine-grained material may have been deposited in small ice-dammed lakes that formed in reentrants along the valley wall where tributaries debauched into the trunk valley.

In Pennsylvania, large fan-shaped deposits of sand and gravel lie at the mouths of tributaries that



Figure 6. Cemented cobble-pebble gravel outwash just north of Flatbrookville along the west side of County Route 615 overlying limestone of the Rondout Formation.



Figures 7a, 7b, and 7c. Figure 7a, *above*. Longitudinal profiles (section A-A') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Bushkill and Flatbrookville, PA-NJ, 7.5 minute quadrangles. Figure 7b, *right top*. Longitudinal profiles (section B-B') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Lake Maskenozha and Culvers Gap, PA-NJ, 7.5 minute quadrangles. Figure 7c, *right bottom*. Longitudinal profiles (section C-C') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Milford and Port Jervis South, PA-NJ, 7.5 minute quadrangles. Location of sections are shown in Figure 1. Profiles constructed by projecting elevation and contacts to a centerline drawn up Minisink Valley. Additional elevation data determined from 1:4800 (5 foot contour interval) topographic maps constructed for the Delaware Water Gap National Recreation Area, and measurements using a hand level. List of units where not labeled: Qov - valley train deposit, Qft - meltwater-terrace deposit, Qst3 - abandoned Pleistocene flood plain, Qst2 and Qst2a - abandoned Holocene flood plains, Qk - kame, Qfp - modern flood plain.

feed Minisink Valley. They may be as high as 560 feet (171 m) above sea level, and their apex lies well upstream in the tributary valley and ends at an abrupt increase in slope. These fans were laid down by meltwater streams that drained the adjacent uplands, and they are graded to the surface of the valley-train deposits. The largest fans lie at the mouths of Adams Creek, Toms Creek, and Sawkill Creek, which upon the latter rests the village of Milford. Based on an outcrop on a bluff overlooking the Delaware River, Happ (1938) suggested that the fan at Milford is a delta terrace. Here approximately 28 feet (9 m) of silt is overlain by 22 feet (7 m) of sandy gravel that dips 10 degrees eastward. Capping this sequence is 3 feet (1 m) of bouldery gravel interpreted as topset beds. Happ suggested the delta had been built in a large glacial lake formed between stagnant ice blocks, or dammed by a recessional moraine down valley. Crowl (1971) did not see any evidence to support the former existence of a large lake in the valley, and suggested the deltaic material was a local kettle-hole filling.

Kames and kame terraces consist of a varied mixture of stratified sand, gravel, and silt that lie above local base-level controls. In most places they have terrace forms and appear to have been laid down between the margin of the Minisink lobe and the valley wall. A few form high-standing hummocky hills that suggest they were deposited in a crevasse, ice-walled sink, or moulin within the stagnant glacier margin.

Meltwater-terraces in Minisink Valley (figs. 7a,b,c) are cut terraces eroded in valley-train deposits, outwash-fans, and some meltwater-terrace deposits by meltwater streams emanating from the glacier margin at a distance up valley. These deposits are no more than 15 feet (5 m) thick. They largely consist of material eroded and reworked from adjacent and the upstream parts of higher outwash deposits, and till that covered the lower part of valley slopes. These terraces generally have flat, gently sloping surfaces, which in places are cut by later meltwater channels. In places, the terraces are paired and in other places, they are not. Lateral slip-off slopes show that these later meltwater streams rapidly eroded the more proximal glacial valley fill.

GLACIAL HISTORY

Glacial Erosion

The distribution and differences in weathering characteristics of glacial drift in northwestern, New Jersey (Salisbury, 1902; Stone and others, 2002) show continental ice sheets covered northern New Jersey at least three times during the Pleistocene epoch (fig. 8). The action of each ice sheet modified the landscape by deeply scouring valleys, wearing down and streamlining bedrock ridges, hills, and slopes. Both the floor of the Minisink and Wallpack Valleys, and part of Kittatinny Valley were deeply scoured by glacial erosion. Depressions in the buried-bedrock floor of Minisink Valley indicate that glacial scour exceeds 50 feet (15 m) and may be as much as 150 feet (46 m) (Witte and Stanford, 1995). Due to weathering, only erosional features of the late Wisconsinan glaciation are preserved. These include polished and plucked bedrock, striations, and streamlined bedrock forms called *roche moutonnées*. The many unweathered and lightly weathered bedrock outcrops also show that preglacial saprolite and soil were removed by glacial erosion. However, an outcrop of saprolite observed by the authors on the Poxono Island Formation downvalley in the Bushkill quadrangle shows that at least some preglacial materials were not completely eroded.

Glacial Advance and Changes in Direction of Regional Ice Flow

The late Wisconsinan advance of ice into the upper part of Kittatinny Valley is obscure because glacial drift and striae that record this history have been eroded or were buried. If the ice sheet advanced in lobes as suggested by the lobate course of the Terminal Moraine; then its initial advance was marked by lobes of ice moving down the Kittatinny and Minisink Valleys. Sevon and others (1975) speculated that ice from the Ontario basin first advanced southward into northeastern Pennsylvania and northwestern New

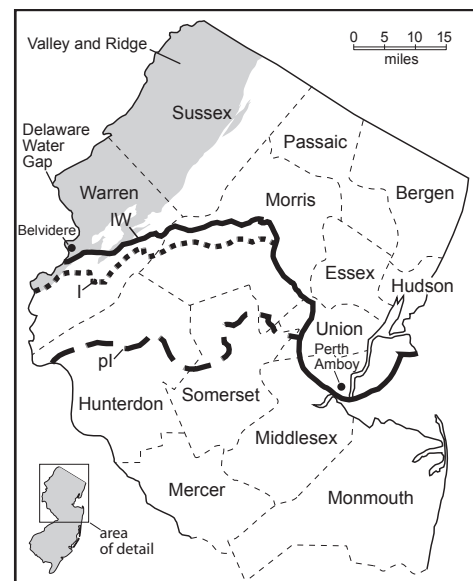


Figure 8. Limits of glaciations in New Jersey and nearby New York. The trace of the IW limit generally marks the position of the Terminal Moraine. Key: IW - late Wisconsinan, I - Illinoian, and pl - pre-Illinoian.

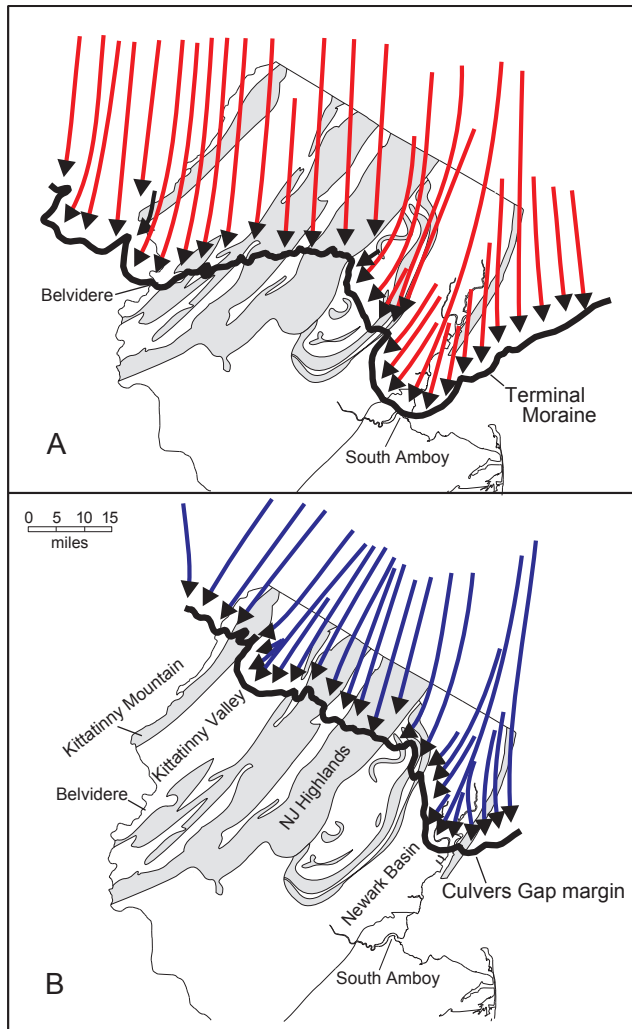


Figure 9, *left*. Generalized direction of ice movement in northern New Jersey during the late Wisconsin. Lines represent regional ice-flow movement at the base of the ice sheet. Flow directions are based on striae, drumlins, dispersal of erratics, and till provenance. Shaded areas represent major uplands. Figure 9a shows direction of ice flow when the glacier margin was at the Terminal Moraine. Field data in the Kittatinny Valley area indicates ice flowed southward across the valley's southwest-trending regional topographic grain. Figure 9b shows direction of ice flow during deglaciation. Flow lines in Kittatinny and Minisink Valleys and surrounding uplands are oriented in a southwest direction with well developed lobate ice flow at the glaciers margin. The change in regional ice flow to a southwest direction appears to be related to thinning of the ice sheet at its margin, and reorganization of ice flow around the Catskill mountains, and in the Hudson-Walkill valley. Data from Ridge (1983), Stanford and Harper (1985), Witte (1988). Sevon and others (1989), Stone and others (2002), and unpublished field maps on file at the New Jersey Geological Survey, Trenton, New Jersey.



Figure 10, *above*. Nepheline syenite erratic enclosed in a cage of rebar on the Glacial Geology Trail, Stokes State Forest, Sussex County, New Jersey.

Jersey. Later, ice from the Hudson-Walkill lowland, which initially had lagged behind, overrode Ontario ice, and ice flow turned to the southwest. In this scenario, the course of the Terminal Moraine in Kittatinny Valley (fig. 9) was controlled by ice flowing from the Hudson-Walkill lowland. Connally and Sirkin (1986) suggested that the Ogdensburg-Culvers Gap moraine represents or nearly represents the terminal late Wisconsin position of the Hudson-Champlain lobe based on changes in ice flow noted by Salisbury (1902) in the vicinity of the moraine. Ridge (1983) proposed that a sublobe of ice from the Ontario basin overrode Kittatinny Mountain and flowed southward into Kittatinny Valley. Southwestward flow occurred only near the glacier margin where ice was thinner, and its flow was constrained by the southwesterly trend of the valley. Analyses of striae, drumlins, and the distribution of erratics in the upper part of Kittatinny Valley (Witte, 1988) and adjacent Kittatinny Mountain (Witte, 2008; Witte and Epstein (2004) partly support Ridge's view. These data further show that by the time the Ogdensburg-Culvers Gap moraine was formed, ice flow in Kittatinny Valley had turned completely to the southwest with extensive lobation at the glacier's margin.

Boulders of nepheline syenite (fig 10), uncovered by a recent landslide in Minisink Valley (Epstein, 2001), provide additional information on the advance history of the Kittatinny and Minisink Valley ice lobes. These boulders, found in the toe-slope deposit of the slide, were apparently derived from thin till that had covered a steeply-dipping bedrock slope or they were part of thicker till at the slope's base that was uncovered by the slide. Previous investigations by Salisbury (1902), Ridge (1983), and Witte (1988, 1991, and 2008) have shown that the boulders are found only in Kittatinny Valley and on Kittatinny Mountain northward of Culvers Gap. Based on their location here, ice would have had to flow S 55° west across Kittatinny Mountain, from the outcrops in Kittatinny Valley. This direction is not consistent with the direction

of ice flow reconstructed from striae, drumlins, and till provenance for this area (Witte, 1997a). Clearly this find represents an anomaly that can not be explained by the direction of regional ice flow during the late Wisconsinan maximum or during deglaciation, both of which has been firmly established by previous investigations. A possible explanation for the location of the syenite-bearing till may be that it represents the initial advance of late Wisconsinan ice into this area, and that ice from the Hudson-Walkill lowland may have been the first to reach the area. Later as the ice thickened and the Kittatinny and Minisink Valleys lobes coalesced, ice flowed turned southward.

TIMING AND STYLE OF DEGLACIATION

The timing of deglaciation is uncertain for the Minisink Valley lobe due to scant radiocarbon dates, errors in dating bog-bottom organic material, and inadequacies of using sedimentation rates to extrapolate bog-bottom dates. Regionally, a few radiocarbon dates bracket the age of the late Wisconsinan terminal moraine and indicate minimum dates of deglaciation. Radiocarbon dating of basal organic material cored from Budd Lake by Harmon (1960) yielded a date of 22,890 \pm 720 yr B.P. (I-2845), and a concretion sampled from sediments of Lake Passaic by Reimer (1984) that yielded a date of 20,180 \pm 500 yr B.P. (QC-1304) suggests that the age of the Terminal Moraine is about 22,000 to 20,000 yr B.P. Basal organic material cored from a bog on the side of Jenny Jump Mountain approximately 3 miles (4.8 km) north of the Terminal Moraine by D. H. Cadwell (written commun., 1996) indicates a minimum age of deglaciation at 19,340 \pm 695 yr B.P. (GX-4279). Similarly, basal-organic material from Francis Lake in Kittatinny Valley, which lies approximately 8 miles (12.9 km) north of the Terminal Moraine indicates a minimum age of deglaciation at 18,570 \pm 250 yr B.P. (SI-5273) (Cotter, 1983). Because the lake lies approximately 3 miles southeast of the Franklin Grove moraine, this age is used here as a minimum date for that feature. Exactly when the ice margin retreated out of the New Jersey part of Kittatinny Valley is also uncertain. A concretion date of 17,950 \pm 620 yr B.P. (I-4935) from sediments of Lake Hudson (Stone and Borns, 1986) and estimated ages of 18,000 yr B.P. for the Ogdensburg-Culvers Gap moraine, and 17,210 yr B.P. for the Walkill moraine by Connally and Sirkin (1973) suggest ice had retreated from New Jersey by 17,500 yr B.P. Correlation of major ice-recessional positions and morphostratigraphic units across Minisink and Kittatinny Valleys is shown in table 1.

The oldest Francis Lake date and the lake's proximity to the Franklin Grove moraine provide the best evidence for the timing of deglaciation in Minisink Valley. The Franklin Grove margin has been correlated to the Sand Hill Church deposits in the Echo Lake Lowland by Witte (1991, 1997a), and its position has been further constrained in Minisink Valley by Witte and Epstein (in review). Based on the tracing of the Franklin Grove margin, a minimum date of deglaciation for the lower part of Minisink Valley is approximately 18,500 yr B.P. Crowl (1980) suggested that deglaciation in northeastern Pennsylvania began at approximately 15,000 yr B.P. based on a suite of basal-organic radiocarbon dates that ranged from 12,500 to 14,170 yr B.P., and estimates on the melting rate of residual ice blocks. The 15,000-year deglaciation date has also been cited in archaeological investigations in Minisink Valley by Stewart (1989), and Dent (1989). Cotter (1983), using radiocarbon dating and pollen stratigraphy, showed that the younger dates in the Minisink Valley area were comparable to other dated pollen sequences in northeastern Pennsylvania and New Jersey that additionally contained lower and older pollen. These basal deposits consistently contained pollen spectra characteristic of the herb pollen zone, an indication of tundra vegetation and a cold and wet climate. Cotter believed that organic sedimentation and Lake Formation had been delayed in Minisink Valley and on the Pocono Plateau. The older date of deglaciation also accords well with younger dates in the upper part of the Susquehanna drainage basin cited in Fleisher (1986) and Ozvath and Coates (1986).

Based on the morphosequence concept of Koteff and Pessl (1981), valley-train deposits show that the Minisink Valley lobe retreated to the northeast in a systematic manner, and chiefly by stagnation-zone retreat. Successive ice-retreat positions are marked by the heads of the Zion Church, Dingmans Ferry, Montague, and Tristates valley trains. Four end moraines, named Millbrook Village, Dingmans Ferry, Montague, and Millville further delineate ice-marginal positions, and both the Dingmans Ferry and Montague moraines are directly traceable eastward where they join the Ogdensburg-Culvers Gap, and Augusta moraines. Outwash-fan deposits provide additional information that supports a marginal retreat model. These deposits lie at the mouths of tributaries that feed the Minisink, and most of them are graded to the surface of adjacent or nearby valley-train terraces. If the fans had been laid down against stagnant ice, they would not be graded to valley-train terraces, and collectively they would lie at greater and varying heights than nearby valley-

train deposits. Meltwater terraces are evidence of glaciofluvial erosion and they show that the slightly older and abandoned valley-train terraces down valley were incised as a new valley train was laid down from an ice-retreat position up valley. This fill and cut model and its resulting assemblage of landforms is illustrated in figure 5.

Most of the terrace scarps have been modified or completely formed by the action of meltwater and postglacial streams. Except in a few places, there is no evidence to suggest that these materials were laid down against remnants glacial ice as envisioned by Happ (1938) and Crowl (1971). The longitudinal profiles of valley-train terraces and their downstream continuation from their heads (figs. 7a, b, c) suggest that large masses of residual ice did not cover the valley floor. Collapsed topography and kettles in the outwash do indicate deposition against or over residual ice. However, these landforms are common components of stagnation-zone retreat, and there is no need to invoke regional stagnation to explain their existence.

Summary of Deglaciation

The Zion Church valley train marks a minor halt in the retreat of the Minisink Valley lobe (fig. 2, table 1). It is tentatively correlated with ice-contact deltaic deposits in the Echo Lake lowland in Pennsylvania. A collapsed area of coarse outwash just south of Zion Church marks the approximate location of the margin of the Minisink Lobe. The uncollapsed parts of its head-of-outwash lie at approximately 450 feet above sea level (fig. 7a). Downstream, most of the valley train has been eroded by later meltwater and postglacial stream action.

The Sand Hill Church ice margin (fig. 2, table 1) marks a major retreat position in the Minisink Valley. In Pennsylvania, it is delineated by ice-contact deltaic deposits laid down at the head of the Echo Lake Lowland, and it is correlated with the Millbrook Village and Franklin Grove moraines in New Jersey.

Retreat of the glacier from the Sand Hill Church margin resulted in a proglacial lake occupying a glacially scoured-bedrock basin in Minisink Valley on the west side of Wallpack Bend. Records of borings (Army Corps of Engineers, on file at the New Jersey Geological Survey, Trenton New Jersey) near Wallpack Bend, show thick deposits of sand and silt lie beneath the floor of Minisink Valley. These fine-grained materials are presumably glaciolacustrine and they suggest a short-lived proglacial lake may have existed in the Minisink Valley. The lake may have formed between the Zion Church valley train and the margin of the Minisink Valley lobe, or the lacustrine materials were laid down in deep ice-scoured depressions eroded in the bedrock floor. Bedrock contours of the valley floor constructed from well records (Witte and Stanford, 1995; Witte, 1997b, and 2004) confirm the existence of these deep depressions.

The next ice-recessional position is marked by the Dingmans Ferry moraine and valley train (fig. 2, table 1, and fig. 7b). Similar deposits lie in Wallpack Valley and the moraine traces eastward on Kittatinny Mountain where it joins the Ogdensburg-Culvers Gap moraine. In Minisink and Wallpack Valleys, the valley-trains extend many miles downstream. Collectively, the moraines and their temporally related outwash deposits, define the Culvers Gap margin (Witte, 1997a) in northwestern New Jersey.

North of Dingmans Ferry, the next retreat position is marked by the Montague moraine and valley train (fig. 2, table 1, and fig. 7c). Again, similar deposits lie in Wallpack Valley and the moraine traces an eastward course on Kittatinny Mountain where it joins the Augusta moraine. Collectively, the moraines and their temporally related deposits, define the Augusta margin (Witte, 1997a). In Minisink Valley, the highest part of the valley train terrace rises up to approximately 520 feet (158 m) at its head near the village of Montague. North of Montague, the highest terrace in the valley drops off to approximately 460 feet (140 m).

The Millville moraine marks the next recessional position in Minisink Valley (fig. 2, table 1). It traces an eastward course from Minisink Valley onto Wallpack Ridge where it terminates. In Wallpack Valley the Millville margin is marked by a large ice-contact delta in the Shimers Brook drainage basin (Witte, 1997b) and it has been tentatively correlated to the Steeny Kill Lake moraine on Kittatinny Mountain. In Minisink Valley, the moraine does not have a large valley train associated with it as do the Dingmans Ferry and Montague moraines down valley.

The youngest recessional position in the study area is delineated by the Tristates valley train. (fig. 2, table 1, and fig. 7c). The terrace from an elevation of 510 feet (155 m) near its head extends down valley to the Montague area where it lies at an elevation of 460 feet (140 m). The distal edge of the outwash fan at Milford also lies at a similar elevation indicating that it is graded to the Tristates valley-train terrace.

CONCLUSIONS

The assemblage and distribution of glacial landforms, stratigraphy, and lobate geometry of the glacier margin show that Minisink Valley was deglaciated by a process of marginal retreat rather than regional stagnation. Valley-train deposits and recessional moraines show that the margin of the Minisink Valley lobe retreated unevenly to the northeast by melting at its periphery. During retreat, a stagnant zone of ice was generally present at the ice lobe's margin, and in several places throughout the valley kettles mark the former site of small detached ice blocks. At times, the retreating glacier margin halted and a large valley train was built up that extended many miles down valley. This style of retreat is called stagnation-zone retreat by Koteff and Pessl (1981). In places the recessional moraines mark readvances, and they show that active ice was also present at the ice lobe's margin. The marginal retreat history of Minisink Valley is further evidenced by a similar history of retreat in Wallpack Valley and Kittatinny Valley (table 1) and the tracing of ice marginal positions. The resulting pattern of systematic deglaciation strongly suggests that regional stagnation did not occur in the Minisink Valley area.

Meltwater deposition in Minisink Valley is best exemplified as a fill and cut model (fig. 5). During halts in glacial retreat, outwash deposits were built up at the margin of the ice lobe while down valley the outwash laid down at older retreat positions was eroded as the meltwater river adjusted to its longer course. In places glacial lakes formed between the ice lobe and older deposits down valley, or in large depressions scoured in the valleys bedrock floor. These lacustrine materials were covered by valley train deposits laid down from retreat positions up valley.

POSTGLACIAL FLUVIAL HISTORY OF THE DELAWARE RIVER IN MINISINK VALLEY

INTRODUCTION

Late Pleistocene glacial outwash and postglacial stream terraces, and Holocene stream terraces and flood plains in Minisink Valley record the change from a braided stream to a meandering stream of low sinuosity. Based on their continuity and uniform height above the Delaware River, alluvial terraces in Minisink Valley consist of at least two stream-terrace deposits (Qst3 and Qst2) that cover large parts of the valley floor and the modern flood plain. Additionally, buried-gravel strath terraces mark the former elevated position of the postglacial Delaware River. Collectively, these deposits define a postglacial history of incision followed by lateral channel migration. Alluvial records show that flood-plain deposition was episodic punctuated by short to long periods of land stability and soil formation. Tentatively, the postglacial stream terraces in Minisink Valley have been assigned these ages (Witte, 2001b) based on stratigraphic position and ¹⁴C dates of their basal layers. Qst3 is late Pleistocene to early Holocene age and Qst2 is early to late Holocene age. Extensive archaeological investigations of Qst2 and the modern flood plain have provided many ¹⁴C dates and have produced cultural artifacts that date from late Pleistocene through to Historic time. Native Americans have utilized Minisink Valley flood plains for short- and long-term occupation with the earliest date of 11,500 yrs BP determined at the Shawnee-Minisink site by McNett and others (1977).

Although a valley-wide compilation of dates, relating to flood plain morphochronology, has not been done, a cursory look at dated alluvial horizons shows that accretion rates were highly variable throughout Minisink Valley. Environmental factors that may have influenced flood plain and terrace formation include 1) stream discharge related to the transition from a meltwater to a groundwater/meteoric source), and changes in precipitation over extended periods, 2) sediment load (coarse and fine), and the effects of vegetation on erosion and sediment supply, 3) nature of river bank materials (grain size, bedrock or surficial, erodability), 4) width of valley (rock walls), may inhibit lateral movement of channel, 5) tributary influence (source of sediment), 6) isostatic rebound (estimated at > 2.5 feet/mile) and 7) anthropogenic influences brought about by deforestation and farming over the last 300 years.

Regionally, the alluvial stratigraphy of Minisink Valley appears to have been largely controlled by climate and its long term influence on discharge (relating to changes in precipitation over extended periods) and sediment load (coarse / fine), (relating to the effects of vegetation on erosion and sediment supply). Because flood plains are built layer by layer they preserve a nearly continuous record of the valley's alluvial history that may be used to document climatic variation since the last ice age. In addition, proxy climate records determined from pollen and other organic materials collected from nearby bogs provide additional data to determine climatic variability. The long-term paleoclimatic record of Minisink Valley and its effects on the alluvial history of the Delaware River will provide a baseline to help us explain future changes in

alluviation in response to climate variability.

PREVIOUS STUDIES

Based on detailed mapping of surficial deposits in Minisink Valley (Stone and others, 2002; Witte (2004), and Witte and Epstein (in review), postglacial-stream terraces in Minisink Valley were divided into an upper (Qst3) and lower (Qst2) terrace based on uniform height above the modern river and continuity throughout the valley.

Archaeologic and paleopedologic studies in Minisink Valley have provided an enormous amount of information on its paleoenvironment. Several investigations on deep alluvial sequences have accorded scientists with a detailed record of alluviation and Amerind culture throughout the Holocene. Important to this study are the thick flood-plain sequences exhumed at the Shawnee-Minisink site (McNett and others, 1977), and upper Shawnee Island site (Stewart, 1989). In the study area Stewart (1989, p. 102) noted that "sedimentary sequences representing habitable landforms generally do not predate 8000/7000 BC implying that the Delaware River did not enter its present channel until after this time." These terraces are 15 to 25 feet above the river, and they include buried-A horizons, three of which are distinctive chronostratigraphic units, correlated to the Late Terminal/Archaic, late Middle Woodland, and Late Woodland cultural periods. Based on the degree of pedogenic formation Foss (1989) suggested that these paleosols reflect extended periods of landscape stability of about 500 to 1000 years. In addition, nearby palynologic studies of bog- and lake-bottom sediments have provided corollary information on floristic changes in the Minisink Valley area. Together, these studies have documented the paleoenvironmental record during the late Quaternary, providing a baseline for future work.

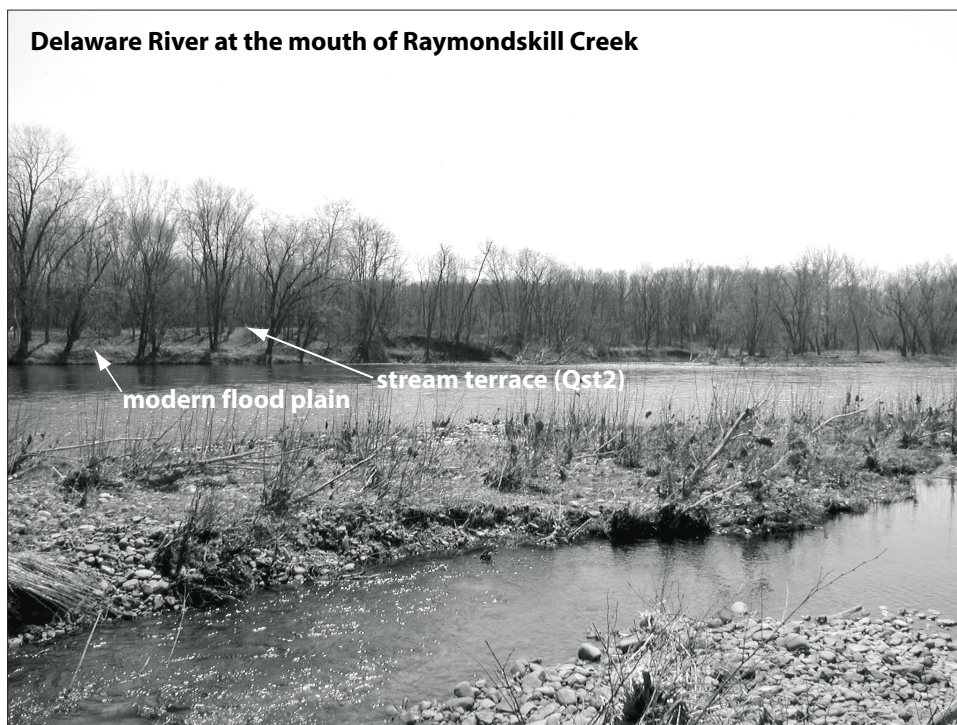


Figure 11. The modern flood plain in Minisink Valley forms a narrow strip of riparian land along the course of the Delaware River. Higher terraces (Qst2 shown here) were built up during the Holocene. They are inundated only during the greatest of floods and are considered abandoned flood plains.

GEOLOGY OF ALLUVIAL DEPOSITS

The modern flood plain, as defined by Leopold and others (1964) has a recurrence of flooding on the order of one and one half to two years. Based on this definition the modern flood plain in Minisink Valley lies as much as 12 feet (4 m) above the mean-annual elevation of the Delaware River (fig. 11). It consists of thinly-bedded, vertically and laterally accreted silt and very fine sand, and it forms a narrow, discontinuous strip of land that lies along the modern course of the river. The lower islands in the river channel are made of sand and gravel, and they principally grow in an upstream direction by accretion of coarse bed load and downstream by the accumulation of sand in slack-water settings formed on the lee-side of the island. These islands in places are covered by thin deposits of overbank sediment.

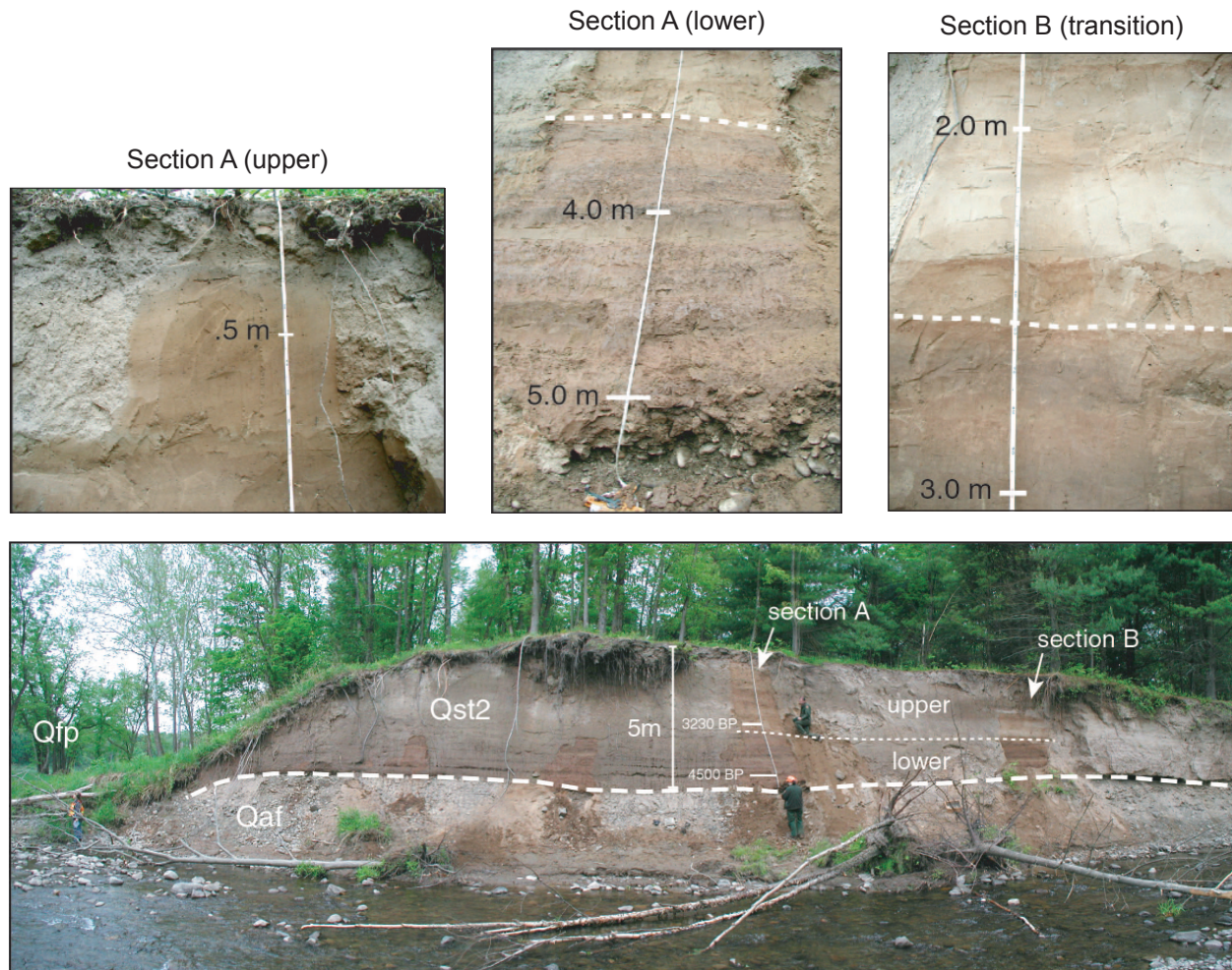


Figure 12. Exposed bluff of Qst2 overlying alluvial-fan gravel at the mouth of Raymondskill Creek. the low terrace to the left of the bluff is the modern flood plain. Figure from Witte and Wright (2001).

Stream-terrace deposits (fig. 12) consist of overbank and minor channel sediment that lie as much as 35 feet (11 m) above the modern flood plain and below meltwater-terrace deposits. Their position in the valley is illustrated in Figures 7a, 7b, and 7c. Based on Leopold and others (1964) definition of the modern flood plain, stream terraces in Minisink Valley are abandoned because they are inundated only during the greatest of floods and annually there is a net loss of land. The latter statement is based on visual estimates. Most land loss is largely brought about by slumping along terrace cut banks, which is especially active during the late winter and early spring.

Based on their continuity and uniform height above the Delaware they have been grouped in two distinct sets. The youngest postglacial terrace (Qst2) lies between 20 and 35 feet (6 and 11 m) above the mean-annual elevation of the Delaware River, and it consists of as much as 20 feet (6 m) of overbank fine sand and silt overlying coarse to fine gravel and sand. In places the underlying coarse material appears to be channel deposits of a postglacial river. Overbank materials consist of thin planar-bedded very fine sand, silt, and fine sand that makes up the levee or near-levee facies, and thinly-bedded clay and silt that defines the back-channel or slack-water facies. Buried paleosols are common and they mark extended periods of little or no overbank deposition and overall land stability. Stewart (1989) showed that there are three buried-A horizons in Minisink Valley that form basin-wide chronostratigraphic units, and these correlate with the Late Terminal/Archaic, late Middle Woodland, and Late Woodland cultural periods. Foss (1989) showed that most of the more mature paleosols he studied in Minisink Valley have cambic B horizons that require at least 500 years to form. Argillic B horizons that require more than 4000 years to develop are absent.

Qst2 covers large portions of the valley floor. Its higher parts lie next to the Delaware River typically

forming a low levee. In a few places the levee is well developed and it forms a prominent ridge that is as much as 6 feet (2 m) high. In most other areas the levee is the highest part on a gently-inclined surface that slopes away from the river. At the base of the valley wall the terrace is marked by a back-channel that in places contains slack-water deposits and organic materials. In several areas on the terrace surface channel scrolls are preserved, especially where the terrace lies on a large inside bend of the river. The range in the height of the terrace above the Delaware River throughout the valley is partly explained by erosion, and differences in local riparian conditions and channel morphometry of the postglacial Delaware River. It is also possible that Qst2 may include several levels as suggested by Wagner (1994) and Witte (2001b). However, without better elevation control, and chronostratigraphic control afforded by radiocarbon dating, correlating these terrace subsets on a regional scale is difficult. Radiocarbon dating of the Qst2 strata by Stewart (1989) and McNett and others (1977) showed that the base of the terrace in places is older than 11,000 yrs B.P. Its upper foot (< 1 m) has been dated to historic times. Basal dates of 4500 B.P. +/- 40 years (GX-28162) at the Bushkill Access site and 4105 B.P. +/- 90 years (GX-22942) at the mouth of Raymondskill Creek (Witte and Wright, 2001) show that in places the Qst2 is much younger (mid-Holocene) indicating that the Delaware River has had an active component of lateral channel migration throughout much of the Holocene.

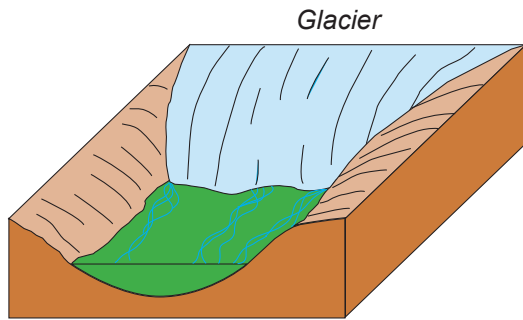
The oldest stream-terrace deposits (Qst3) lie 40 to 50 feet (12 to 16 m) above the mean-annual elevation of the river, and as much as 15 feet above the higher parts of Qst2. Longitudinal profiles of the postglacial terraces (figs. 7a, 7b, and 7c) parallel each other, although there is a slight suggestion of divergence going upstream. Downstream below the Delaware Water Gap they are absent. This may be a relict of delayed postglacial rebound, but without better control on their elevation this position is highly speculative. Qst3 consists of as much as 10 feet (3 m) of thin planar-bedded overbank very fine sand and fine sand, and minor pebbly sand, and like the younger Qst2 they are an abandoned flood plain. This material is similar to the levee and near levee facies described for Qst2, but it has a slightly coarser texture. Paleosols are noticeably scarce, and in the few places they were found they were represented by a truncated B horizon. In many places Qst3 overbank materials lie on coarse gravel and sand with the contact marked by a bouldery lag. In other places they overlie as much as 12 feet of thinly bedded, planar to cross-stratified very fine sand and sand with lesser amounts of pebbly sand, pebble gravel and silt. The base of this material also rests on coarse gravel with the contact marked by a bouldery lag. These deposits of intermediate texture may represent deposition in the river channel or deposition near the channel on a very low flood plain. They suggest the height of the river that laid down Qst3 had to be well above the modern Delaware River and the late Pleistocene and Holocene rivers that laid down Qst2.

Qst3 terraces are typically smaller than, and flank the younger Qst2 terraces, and in some places they lie surrounded by younger deposits. In several areas throughout Minisink Valley, gravel terraces lie next to the sandier parts of Qst3 and Qst2. Apparently, the overlying materials were removed as the postglacial river cut down to its Qst2 level. No dates are available for the Qst3, but based on the age of its younger sibling, it is late Wisconsinan age and it may represent a transition from glaciofluvial to postglacial fluvial environments.

FLUVIAL EVOLUTION

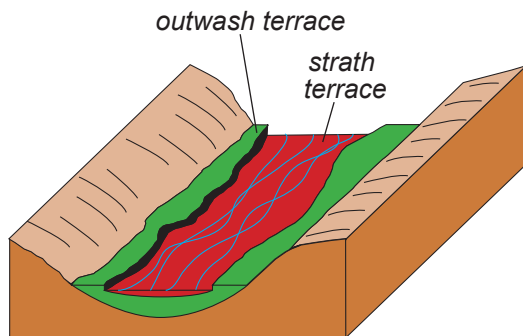
In Minisink Valley the late Pleistocene is marked by the transition of the Delaware River from a braided glacial meltwater stream to a postglacial meandering stream of low sinuosity (fig. 13). The late glacial Delaware River is assumed to be a braided stream, given the large volume of meltwater that flowed through the valley and the readily available source of sediment. Most of the sediment, both suspended and bedload materials, were largely derived from local sources, both eroded along the main reach of the valley and carried in by the many tributaries that drained adjacent uplands. Meltwater derived from ice-marginal positions in the upper parts of the Delaware River drainage basin would have also supplied a steady influx of fine sediment to the Minisink. Coarser materials (cobbles and boulders), given the large distance to Minisink Valley, would have been deposited nearer the glacier margin in valley trains, and outwash fans. Based on the fill and cut model illustrated in Figure 13, a braided river occupied broad parts of the valley floor, lying well below the local valley-train and outwash-fan terraces. In a few places the river's course was constrained to a single channel, lying between high-standing remnants of valley-train deposits and/or the bedrock valley wall. The stratigraphy of the Qst3 and Qst2 terraces, and radiocarbon dating of the Qst2 alluvial sequence shows that the late glacial Delaware River was at least 30 feet above the modern river.

Glacial Fluvial and Postglacial Fluvial History of Minisink Valley



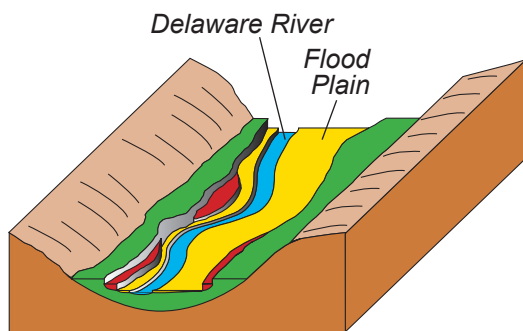
Deposition of glacial outwash (valley train) from local ice-retreat position.

18,000 years ago



Downcutting of "older" valley train deposit and formation of meltwater-terrace (strath) deposits from ice-retreat position far upstream. About 15,000 years ago, glacial meltwater no longer flowed into the Delaware River drainage basin.

16,000 years ago



Erosion of glacial outwash deposits chiefly by incision in response to glacial rebound and lateral migration of the river channel across valley floor. Due to the narrow width of the valley the Holocene flood plain built up to a height that it was only inundated by the greatest of floods. Over time the flood plain became abandoned forming a steam terrace.

8,000 years ago

Figure 13. Late glacial to postglacial fluvial history of Minisink Valley illustrating the development of the valley floor by cut and fill.

In contrast to the late glacial river, the modern Delaware is a nonmeandering stream of low sinuosity that is flanked by 2 abandoned flood plains (Qst3 and Qst2 terraces). The Qst2 alluvial sequences at the Shawnee-Minisink (McNett and others, 1977), and Upper Shawnee Island (Stewart, 1989) sites show that the form of the Delaware River at the end of the Pleistocene was also a non-braided one with a well-established flood plain, and the river was at or slightly above its present elevation.

The transition, from a braided glacial stream with a very distant meltwater source to a non-meandering stream, resulted in significant hydraulic changes in Minisink Valley during the close of the Pleistocene. Most obvious was a substantial decrease in discharge due to the retreat of the Laurentide ice sheet from the Delaware River drainage basin. The minimum date for this event is estimated at 14,000 yrs B.P., based on

the mapping and correlation of ice-marginal positions by Ozvath and Coates (1986) in the Western Catskill Mountains, and Fleisher (1986) in the upper part of the Susquehanna drainage basin. The dramatic decrease in discharge was accompanied by a change in channel form from braided to a non-meandering channel of low sinuosity. This transition may have already been underway during the later stages of deglaciation of the drainage basin when meltwater found new flow paths into the Susquehanna River and to a lesser extent the Hudson River drainage basins.

At some point during the latter part of the late Wisconsinan, the Delaware River underwent a period of incision to a level at or near its present elevation. The timing and possible causes of this event will be examined. Previous investigations by Crowl (1971) and Dent (1989) suggested the coarse gravel beneath Qst2 is glacial outwash, laid down by meltwater during the latter stages of deglaciation of the Delaware River drainage basin. Based on the oldest dates at the Shawnee-Minisink and Upper Shawnee Island sites (McNett and others, 1977), the basal gravel is older than 11,000 yrs B.P. The period represented by the sequence of sediments below the older dates and above the coarse gravel is unknown. Because the rate of sedimentation for the late Pleistocene alluvium has not been constrained by radiocarbon dating and is too variable throughout the valley, an accurate estimate of its age cannot be determined. However, ancillary evidence (chiefly stratigraphic) suggests the basal gravel beneath Qst2 is not glacial outwash, but outwash reworked and incised by the postglacial Delaware River. The position and stratigraphy of the Qst3 terrace show that it is older than Qst2 and that it was laid down by a river that was much higher than the river that deposited Qst3. Qst3 represents the oldest flood-plain deposits in Minisink Valley that were probably laid down by a non-meltwater fed stream that had a non-meandering channel form. This river, apparently operating under a condition of equilibrium, deposited a thin flood plain. It is assumed that this flood plain could not have been formed if the stream had a braided channel form largely fed by meltwater. Because the Qst3 deposits appear to have been laid down by a non meltwater or largely non meltwater-sourced stream, they may date to a period about 14,000 to 15,000 years ago. Incision to the Qst2 level appears to have been initiated by the onset of delayed isostatic rebound, and possibly a reduction in sediment supply due to the floristic transition from tundra to a closed boreal forest. The 14,000 yr B.P. maximum date for the start of rebound (Koteff, 1989) and the 14,250 yr B.P. date marking the transition from herb to spruce pollen zones (Cotter, 1983), seems to be in accordance with the estimated age of Qst3.

A large area of wind blown sand on the T3 terrace just south of Minisink Island (fig. 14) provides additional evidence that the Qst3 terrace is of Pre-Boreal age. Here small sand dunes cover part of the Qst3 surface and extend eastward over the surface of the Montague valley train and up the lower part of the eastern valley slope. The wind-blown materials are not found on the surface of Qst2 next to and westward of the dune field. The position of the dune field shows that it was deposited after the Qst3 flood plain was abandoned, but before the growth of an extensive cover of vegetation. The eolian sand may be reworked Qst3 material blown off a formerly larger and more extensive Qst3 flood plain.

Based on the estimated age of deglaciation for the Delaware River drainage basin, and a maximum age of 11,000 yrs B.P. for the Qst2 alluvium, it is estimated that the Qst3 to Qst2 incision of the Delaware River lasted only a few thousand years. The timing of this event seems to correspond with the onset of delayed rebound and the floral transition from tundra to a closed boreal forest, which may have lowered

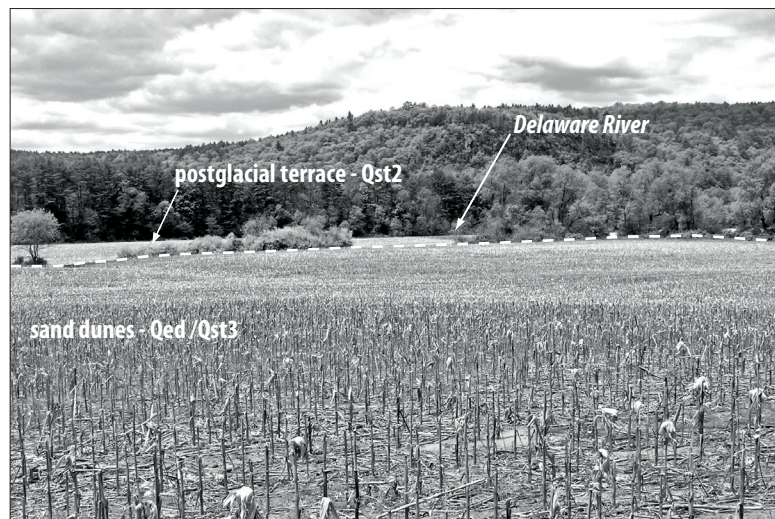


Figure 14. Low sand dunes overlying postglacial terrace Qst3 in Minisink Valley, Namanock Island area. The lower Qst2 postglacial terrace is not blanketed by eolian sand suggesting that the dunes were deposited in late glacial to early postglacial time (~16 to 13 ka). Eolian deposits (low dunes and sheet sands) were locally derived from deflated glacial outwash terraces in Minisink Valley.

Years B.P.	Ecological Period	Cultural Period	Pollen Zone	Climate	Vegetation	Fluvial Activity	Radiocarbon Dates	Soil Strat. Unit	Morphostratigraphic Unit.
2000	Sub-Atlantic	Historic	Oak-mixed Hardwoods	Similar to modern conditions	Oak and chestnut forest	Flood plain stability Minor alluviation Formation of the modern flood plain		Ab1	T1
		Woodland		Warmer and dryer,(very low ppt. rates (altithermal))	Oak and hickory forest	Flood plain stability Minor alluviation		Ab2	
4000	Subboreal	Archaic	Oak-Hemlock	Warmer with near-modern conditions	Oak and hemlock forest; gradual declineof hemlock	Increased alluviation during the latter part of the Atlantic Period	(4) 9330 (3) 10,750 (2) 12,160 Onset of rebound Deglaciation of the upper Delaware	Ab3	T2
6000	Atlantic					Flood plain stability			
8000			Lateral channel migration and accretion of the T2 flood plain						
10,000	Boreal		Pine	Warm and dry	Pine and birch to pine and oak forest	Channel of the Delaware River at or slightly above modern level.			
12,000	Preboreal	PaleoIndian	Spruce	Cool and wet	Spruce and fir forest	Incision, abandonment of the T3 flood plain. Initial formation of the T2 flood plain			Qes
14,000	Postglacial			Cold and wet	Open tundra to spruce parkland	Cut off of meltwater, shift from a braided to a nonmeandering stream. Lateral channel migration and flood plain formation.			T3
16,000		Herb	Coldest				Open tundra	Deposition of outwash Aggradation near the glacier margin, erosion down stream	(1) 18,570
18,000	Glacial								
									Qd

Table 2. Summary of paleo-environmental data for Minisink Valley from Delaware Water Gap to Port Jervis, New York, and correlation of soil stratigraphic and morphostratigraphic units. Pollen zone data modified from Cotter (1983), climate data modified from Dent (1979) and Vento and Rollins (1989), vegetation data modified from Dent (1979), interpretation of fluvial activity modified from Vento and Rollins (1989) and Witte (1997a, 1997b, and unpublished data), soil stratigraphic units modified from Vento and Rollins (1989) and Stewart (1989), and morphostratigraphic units from Witte (1997a, 1997b, and unpublished data). Radiocarbon dates from (1) Cotter (1989), (2 and 3) McNett and others (1977), (4) Stewart (1989). Delayed rebound from Korteff and Larsen (1989).

sediment yield in the drainage basin (Table 1). Qst2 represents episodic periods of alluviation throughout the Holocene. Leopold and others (1964, p. 326) noted that the "progressive lateral migration of the river channel removes portions of the flood plain and hence limits the elevation of its surface." Due to the narrow width of Minisink Valley and low sinuosity of the Delaware River channel, the Qst2 flood plain outgrew its fluvial setting and eventually became abandoned, receiving sediment only during the greatest of floods.

RECORD OF PALEO-ENVIRONMENTAL CHANGE

The paleoenvironmental record of Minisink Valley, from late glacial to modern time is summarized in table 2. It provides a baseline to interpret the paleofluvial responses of the Delaware River within the context of climatic change in the river's drainage basin. The earlier postglacial climate in the Minisink Valley area has been resolved chiefly by palynologic investigations of bog- and lake-bottom sediments. Results from these studies have provided a detailed record of floristic evolution based on the identification and changes in the percentage of arboreal and nonarboreal pollen. In the Minisink Valley area, sequences of similar pollen spectra, dated by radiocarbon, define zones that are regionally and chronologically consistent (Sirkin and Minard, 1972; Cotter, 1983), and have given scientists a valuable tool to investigate the migration and distribution of vegetation during the Late Quaternary and interpret the paleoclimate on both a local and regional scale.

The late Pleistocene pollen record marks the transition from a cold to a temperate environment. Major divisions of dominant pollen taxa are represented by the herb, spruce, and pine zones (table 2). The herb pollen zone indicates the presence of tundra vegetation and the presence of many wetlands in the Minisink

Valley area (Cotter, 1983). The younger part of the zone typically shows a rise in the percentage of spruce, pine, and birch and this reflects a warming of the climate and a corresponding increase in the variety and percentage of arboreal taxa. The paleoenvironment interpreted from the pollen record shows that the study area was initially a tundra with sparse vegetal cover. As the climate warmed the open area of the tundra was replaced by an open parkland of sedge and grass with scattered arboreal stands that largely consisted of spruce. The spruce pollen zone defines a period from about 14,250 to 11,250 yrs B.P., and its pollen sequence records: 1) the rise in the percentage of spruce and pine, 2) secondary increases in fir, oak and birch, and 3) decreases in percentage of nonarboreal plants. This change in pollen spectra and percentages records the continued amelioration of the climate and the transition to a dense closed boreal forest that consisted of spruce and fir blanketing the uplands and hillslopes with pine and birch covering the floor of Minisink Valley (Dent, 1989). Dent (1989) further suggested that a true boreal forest did not become established in Minisink Valley until 10,680 yrs B.P.

The pine pollen zone represents the period between 11,250 and 9,700 yrs. B.P. and it marks the emergence of white pine as the dominant arboreal component. Dent (1989) suggested that the uplands and hillslopes at this time were populated by a dense pine forest, and the valley floor was covered by a mix of pine, birch, and cedar. Open areas of land were rare only forming small meadows on the river's flood plain or on adjacent terraces that were subject to infrequent floods. The climate during this period, although warmer than the past, was still cooler than the modern-day climate. The transition from coniferous (pine) to deciduous (oak) forest is estimated at 9700 yrs B.P. by Cotter (1983). However, Dent (1989) placed the break at 9250 yrs B.P. which marks the transition between the Pre-Boreal and Boreal periods. The Boreal period was divided by Dent based on the shift from a boreal forest dominated by pine and birch to a forest dominated by oak and birch.

The earliest record of man in Minisink Valley is recorded at the Shawnee - Minisink site (McNett and others, 1977), where charcoal associated with a hearth and Paleoindian artifacts yielded a date of 10,590 +/- 300 yrs B.P. (W-2994), and 10,750 +/- 600 yrs B.P. (W-3134), and the nearby upper Shawnee Island site (Stewart, 1989) which yielded a date of an Early Archaic date 9380 +/- 545 yrs B.P. from a similar setting. Based on the cultural and biotic components at the occupation sites, Paleoindians were small bands of hunter-foragers that visited the Minisink Valley during the warmer months of the year to fish, hunt (?), and gather seeds (Dent, 1989). Their arrival in Minisink Valley shows that climate, and ecology of the Preboreal to Boreal flood plain was suitable for occupation.

CONCLUSIONS

Based on the estimated age of deglaciation for the Delaware River drainage basin, and a maximum age of 10,750 yr. B.P. for the Qst2 alluvium, it is estimated that the Qst3 to Qst2 incision of the Delaware River lasted only a few thousand years. The timing of this event seems to correspond with the onset of delayed rebound and the change from tundra to a closed boreal forest, which may have lowered sediment yield in the drainage basin (Table 1). The Qst2 terrace represents episodic periods of alluviation throughout the Holocene. Due to the narrow width of Minisink Valley and low sinuosity of the Delaware River channel, the Qst2 flood plain outgrew its fluvial setting and eventually became largely abandoned (at least in the context of the modern floodplain), receiving sediment only during the greatest of floods. Presently most of the Qst2 terraces are losing material, largely due to slumping along their riparian borders.

Original article *Late Wisconsinan deglaciation and postglacial history of Minisink Valley, Delaware Water Gap to Port Jervis, New York*, printed in Inners, J.D., and Fleeger, G.M., eds., 66th Annual Field Conference of Pennsylvania geologists, Delaware Water Gap, p. 99-118.

REFERENCES

- Connally, G. G., and Sirkin, L. A., 1973, Wisconsinan history of the Hudson-Champlain lobe, in Black, R. F., Goldthwait, R. P. and William, H. B. (eds.), *The Wisconsinan stage*: Geol. Soc. Amer. Memoir 136, p. 47-69.
- _____, 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, Bull. no. 455, p. 50-69.
- Cook, G.H., 1880, Glacial drift: N.J. Geological Survey Ann. Rept. of 1880, p. 16-97.

- Corps of Engineers, U. S. Army (1967), Tocks Island Reservoir, Site Geology, Design Memorandum no. 6: USA Engineer District, Corps of Engineers, Philadelphia, Pa.
- Cotter, J. F. P., 1983, The timing of the deglaciation of northeastern Pennsylvania and northwestern New Jersey: Doctoral Dissertation, Lehigh University, Bethlehem, Pa., 159 p.
- Cotter, J. F. P., Ridge, J. C., Evenson, E. B., Sevon, W. D., Sirkin, Les, and Stuckenrath, Robert, 1986, The Wisconsin history of the Great Valley, Pennsylvania and New Jersey, and the age of the "Terminal Moraine": in Cadwell, D. H. (ed.), The Wisconsin stage of the First Geological District, eastern New York: N.Y. State Museum Bulletin 455, p. 22-50.
- Crowl, G.H., 1971, Pleistocene geology and unconsolidated deposits of the Delaware Valley, Matamoras to Shawnee on Delaware, Pennsylvania, Pennsylvania Geological Survey, 4th., ser., General Geology Report 71, 68 p.
- _____, 1980, Woodfordian age of the Wisconsin glacial border in northeastern Pennsylvania: *Geology*, v. 8, p. 51-55.
- Crowl, G.H., and Sevon, W.D., 1980, Glacial border deposits of late Wisconsin age in northeastern Pennsylvania, Pennsylvania Geological Survey, 4th ser., General Geology Report 71, 68 p.
- Dent, R. J., 1989, Archaeology in the upper Delaware Valley: the earliest populations: in Orr, D. G., and Campana, D. V. (Eds.) *The People of Minisink, Papers from the 1989 Delaware Water Gap Symposium*: National Park Service, U.S. Department of the Interior, p. 117-143.
- Depman, A. J., and Parillo, D. G., 1969, Geology of the Tocks Island area and its engineering significance: in Subitsky (ed.) *Geology of selected areas in New Jersey and Pennsylvania*: Rutgers Univ. Press, New Brunswick, New Jersey, p. 354-362.
- Drake, A. A., Jr., Volkert, R. A., Monteverde, D. H., Herman, G. C., Houghton, H. H., Parker, R. A., and Dalton, R. F., 1996, Bedrock geologic map of northern New Jersey: U.S. Geological Survey Misc. Geol. Inv. Map I-2540-A, scale 1:100,000.
- Epstein, J. B., 1969, Surficial Geology of the Stroudsburg Quadrangle, Pennsylvania-New Jersey: Pennsylvania Geological Survey, 4th series, Bulletin G57, 67p., scale 1:24,000.
- _____, 2001, Geologic controls of landslides in the Delaware Water Gap National Recreation Area, New Jersey – Pennsylvania, and Lehigh Gap, Pennsylvania: in Inners, J.D., and Fleeger, G.M., eds., 66th Annual Field Conference of Pennsylvania Geologists, Delaware Water Gap, p. 119-135.
- Epstein, J.B., and Koteff, Carl, in press, Surficial Geology of the Saylorsburg Quadrangle, Pennsylvania: U.S. Geological Open-file Report.
- Fleisher, P. J., 1986, Glacial geology and late Wisconsin stratigraphy, upper Susquehanna drainage basin, New York: in Cadwell, D.H., (ed.), *The Wisconsin Stage of the First Geological District, Eastern New York*: New York State Museum, Bull. no. 455, p. 121-142.
- Happ, S. C., 1938, Significance of Pleistocene deltas in the Minisink Valley: *Amer. Jour. Sci. Series V*, v. XXXVI, p. 417-439.
- Harmon, K. P., 1968, Late Pleistocene forest succession in northern New Jersey: unpublished M.S. thesis, Rutgers Univ., 164 p.
- Hoff, D., 1969, Mastadon at Marshall's Creek, Pennsylvania Game News, 40: p. 3-7.
- Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.
- Koteff, Carl, and Larsen, F. D., 1989, Postglacial uplift in western New England: Geologic evidence for delayed rebound in Gregersen, S., and Basham, P. W., (eds.), *Earthquakes at North Atlantic passive margins: Neotectonics and Postglacial Rebound*, p. 105-123.
- Leopold, L. B., Wolman, M. G., and Miller, J. P., 1964, *Fluvial Processes in Geomorphology*: W. H. Freeman and Company, San Francisco, 522 p.
- Leverett, Frank, 1934, Glacial deposits outside the Wisconsin terminal moraine in Pennsylvania: Pennsylvania Geol. Survey, 4th ser., Bulletin, G 7.
- Lippincott's Gazetteer, 1931, a Complete Pronouncing Gazetteer or Geographical Dictionary of the World: Heilprin, Angelo and Heilprin, Louis (eds.), J. B. Lippincott Company, Philadelphia and London.
- McNett, C. W. Jr., McMillian, B. W., and Marshall, S. B., 1977, The Shawnee-Minisink site in Newman, W. S., and Salwen, Bert (eds.) *Amerinds and Their Paleoenvironments in Northeastern North America*, *Annals of the New York Academy of Sciences*, v. 288, p. 282-298.
- Minard, J. P., 1961, End moraines on Kittatinny Mountain, Sussex County, New Jersey: U.S. Geological

- Survey Professional Paper 424-C, p. 61-64.
- Munsell Color Company, 1975, Munsell soil color charts: a division of Kollmorgen Corp., (unnumbered text and illustrations).
- Ozsvath, D.L., and Coates, D.R., 1986, Woodfordian stratigraphy in the western Catskill Mountains, in Cadwell, D.H., (ed.), *The Wisconsin Stage of the First Geological District, Eastern New York*: New York State Museum, Bull. no. 455, p. 109-120.
- Reimer, G. E., 1984, *The sedimentology and stratigraphy of the southern basin of glacial Lake Passaic, New Jersey*: unpublished M.S. thesis, Rutgers University, New Brunswick, New Jersey, 205 p.
- Ridge, J.C., 1983, *The surficial geology of the Great Valley section of the Valley and Ridge province in eastern Northampton County, Pennsylvania, and Warren County, New Jersey*: Unpublished M.S. thesis, Lehigh University, Bethlehem, Pa., 234 p.
- Salisbury, R. D., 1902, *The glacial geology of New Jersey*: N.J. Geological Survey Final Report, v. 5, 802 p.
- Sevon, W.D. Crowl, G.H., and Berg, T.M., 1975, *The Late Wisconsin drift border in northeastern Pennsylvania: Guidebook for the 40th Annual Field Conference of Pennsylvania Geologists*, 108 p.
- Sevon, W.D., Berg, T.M., Schultz, L.D., and Crowl, G.H., 1989, *Geology and mineral resources of Pike County, Pennsylvania*: Pennsylvania Geological Survey, County Report 52, 141 p. 2 plates, scale, 1:50,000.
- Sirkin, L. A., and Minard, J. P., 1972, *Late Pleistocene glaciation and pollen stratigraphy in northwestern New Jersey*: U.S. Geological Survey Prof. Paper 800-D, p. D51 - D56.
- Stanford, S. D., and Harper, D. P., 1985, *Reconnaissance map of the glacial geology of the Hamburg quadrangle, New Jersey*: New Jersey Geological Survey, Geol. Map Series 85-1, map scale 1:24,000.
- Stewart, Michael, 1989, *Archaeology and environment in the upper Delaware*: in Orr, D. G., and Campana, D. V. (Eds.) *The People of Minisink, Papers from the 1989 Delaware Water Gap Symposium*: National Park Service, U.S. Department of the Interior, p. 79-115.
- 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*: in Sibrava, V., Bowen, D. Q., and Richmond, G. M. (eds.), *Quaternary glaciations in the northern hemisphere: 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 Map Series, scale 1:100,000.
- Vento, Frank and Rollins, H. B., 1989, *Development of a Late Pleistocene-Holocene Genetic Stratigraphic Framework as it Relates to Atmospheric Circulation and Climate in the Upper and Central Susquehanna River Drainage Basin*. Submitted to the Bureau for Historic Preservation, Pennsylvania Historical and Museum Commission.
- Wagner, D. P., 1994, *Pedology and geomorphology of the Depew Recreation Area*, in Inashima, P. Y., *Geomorphology, Remote Sensing, and Archaeological Monitoring at Depew Recreation Area*, Dept of the Interior, National Park Service, p. 1-36.
- White, I.C., 1882, *The geology of Pike and Monroe Counties*: Pennsylvania Geological Survey, 2d, Report G-6, 333 p.
- Witte, R.W., 1988, *The surficial geology and Woodfordian glaciation of a portion of the Kittatinny Valley and the New Jersey Highlands in Sussex County, New Jersey*: Unpublished M.S. thesis, Lehigh University, Bethlehem, Pa., 276 p.
- _____, 1991, *Deglaciation of the Kittatinny and Minisink Valley area of northwestern New Jersey: Stagnant and active ice at the margin of the Kittatinny and Minisink Valley lobes*: in Geological Society of America, *Abstracts with Programs*, v. 23, p. 151.
- _____, 1997a, *Late Wisconsin glacial history of the upper part of Kittatinny Valley, Sussex and Warren Counties, New Jersey*: *Northeastern Geology and Environmental Sciences*, v. 19, no. 3, p. 155-169.
- _____, 1997b, *Surficial geology of the New Jersey portion of the Milford and Port Jervis South quadrangles*: N.J. Geological Survey Open-file Map, scale 1:24,000.
- _____, 2001a, *Late Wisconsin end moraines in northwestern New Jersey: observations on their distribution, morphology, and composition*: in Inners, J.D., and Fleeger, G.M., eds., *66th Annual Field Conference of Pennsylvania Geologists, Delaware Water Gap*, p. 81-98.
- _____, 2001b, *Late Wisconsin deglaciation and postglacial history of Minisink Valley, Delaware Water*

Gap to Port Jervis, New York: in Inners, J.D., and Fleeger, G.M., eds., 66th Annual Field Conference of Pennsylvania Geologists, Delaware Water Gap, p. 99-118.

_____, 2008 GMS 08-2, Surficial Geologic Map of the Branchville Quadrangle, Sussex County, New Jersey, New Jersey Geological Survey Map Series, GMS 08-2., scale: 1 to 24,000, 2 plates size 40x52; 39x51, 3 cross-sections, 3 tables and 5 figures.

Witte, R. W., and Stanford, S. D., 1995, Surficial geology and earth-material resources of Warren County, New Jersey. N.J. Geological Survey Open-file Map no. 15c, scale 1:48,000, 3 plates.

Witte, Ron W. and Epstein, Jack B., 2004, , Surficial Geologic Map of the Culvers Gap Quadrangle, Sussex County, New Jersey, New Jersey Geological Survey Map Series., GMS 04-1, scale: 1 to 24,000, 2 plates, sizes 31x34 and 34x36, 1 cross-section, 1 figure, 1 table, 20-page pamphlet.

Witte, R.W., and Epstein, J.B., in review, Surficial geology of the Flatbrookville quadrangle, New Jersey - Pennsylvania: U.S. Geological Survey Open-file Map.

Pahaquarry Copper Mine

by

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Pahaquarry Copper Mine lies along the western slopes of Kittatinny Mountain in the Delaware Water Gap National Recreation Area (DEWA, figure 1). DEWA's Kaiser trail climbs up the Mine Brook drainage amongst the Pahaquarry mine workings. This mine has a long history of exploration that became entwined with local lore. Local history suggests that it is the oldest mine worked by European settlers in New Jersey and possibly the United States, dating back to Dutch explorers as early as the 1650's and native Americans even earlier (Garber, 1917; Woodward, 1944). This history is still commonly recorded for Pahaquarry. Researchers traced this declaration back to an 1828 publication of several letters in Hazard's Register (Burns Chavez and Clemensen, 1995; Kraft, 1996). This planted the Dutch colonial connection and allowed it to flourish up into the late 20th century. More recent historical research has shown the mine's true origination as 100 years younger, dating back to 1750 (Burns Chavez and Clemensen, 1995; Kraft, 1996).

Even though Pahaquarry is not the first copper mine in New Jersey it lived a long and varied history. It underwent 5 separate periods of exploration and attempted exploitation beginning in 1750's, and continuing from 1829-30, 1847-48, 1861-62 and ending in 1901-12. The following historical description of the Pahaquarry Copper Mine is freely based on the excellent research found in Burns Chavez and Clemensen (1995) and Kraft (1996). The description of the ore mineralization and genesis will be based entirely on Woodward (1944). The reader is referred to their work for all original sources.

The Bloomsburg Red Beds is the Pahaquarry mine's host rock. This Upper Silurian clastic unit overlies the Shawangunk Formation, which holds up the main ridge of Kittatinny Mountain. The Bloomsburg covers the western subsidiary ridges and the rock's northwest dip produces the ridge's western slope. Copper in the Bloomsburg is an uncommon occurrence. When hiking across the Bloomsburg into the mine workings malachite and chrysocolla supply the first evidence of the copper mineralization. This is probably the indicator that first led to developing the Pahaquarry mine (Weed, 1911).

HISTORY OF MINING AT PAHAQUARRY 1750s

Historical records state that copper exploration in the Pahaquarry region began in the 1750's. Between 1753-55, three men, John Reading Jr., Anthony Maxwell, and Martin Ryerson bought and

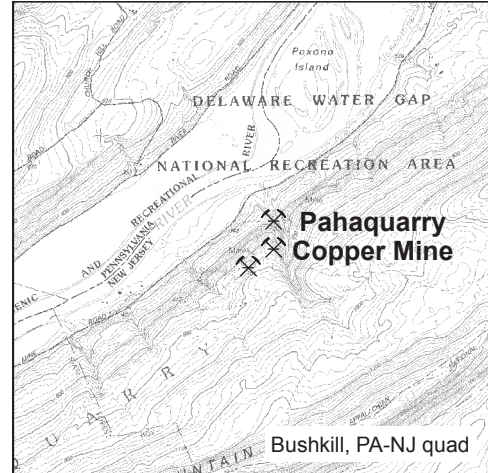


Figure 1. Pahaquarry Copper Mine located within the Delaware Water Gap National Recreation Area, Hardwick Township, Warren County, New Jersey.

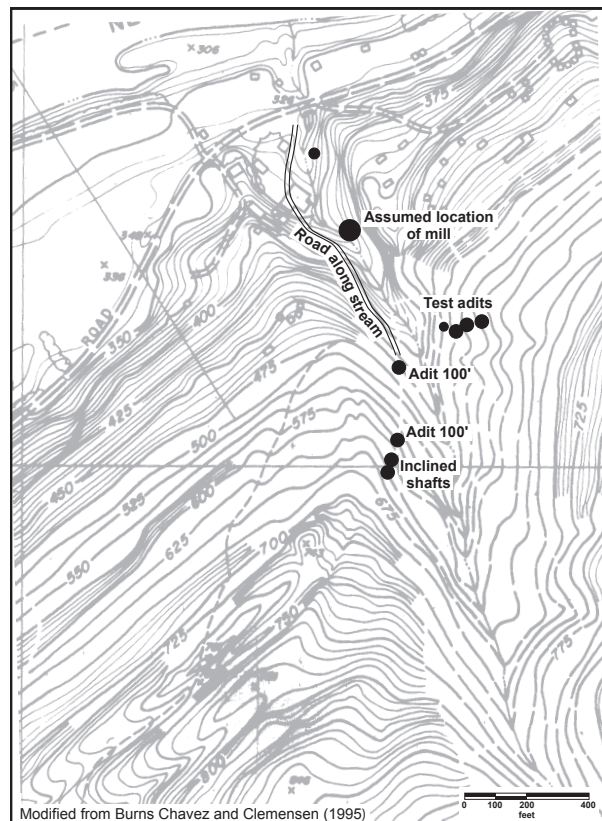


Figure 2. Pahaquarry Mine workings from 1753 to 1760. Several prospects were dug on the northeast side of Mine Brook before work concentrated on the southwestern side. The workings include strike parallel adits and dip parallel shafts. Modified from Burns Chavez and Clemensen (1995).

began developing their copper find. They investigated both sides of Mine Brook. Five prospect pits northeast of the brook can be found up on the hillside (figure 2). Southwest of the brook seemed more promising and underwent the greatest development at this time. They tried to follow the copper mineralization by excavated two adits between 50 to 100 feet each and two 45° inclined shafts that tracked the bedding dip. Several structures consisting of a stamp mill, several dams and water races and buildings for lodging and equipment storage were built on the Delaware River terrace to facilitate the exploration and processing of the copper.

Ore excavation during colonial times was labor intensive. Miners used a single jack (four pound sledgehammer) to drill the rock. Workers held metal drill with one hand that they drove into the rock with the single jack. After each hit, the drill was turned one half revolution before the next strike. Three foot depth proved to be about the maximum attainable before blasting proceeded. Black powder wrapped in paper was pushed in the drill holes by wooden or copper poles. Any other type of pole might cause sparks and ignite the powder. Blasted rock was removed by hand and material high graded at the excavation. Ore was then transported to the mill where it underwent further sledgehammer-induced size reduction and sorting before insertion into the stamp mill where final pulverization and waste washing was completed. The residue of high-grade ore consisting of copper sulfide and carbonates was then transported to a smelter for final copper extraction. The initial phase of copper mining at Pahaquarry ended in 1760 when the work proved unprofitable due to the low grade of the deposit and the difficulty of extraction.

1829-30

Little actual excavation work accompanied a renewed copper mining interest at Pahaquarry in 1829-1830. Historical records show that a group of interested men signed a contracted grant to last 999 years with the Pahaquarry landowners to reestablish copper mining. The grant required 10% of all copper and other minerals extracted to go to the landowners. Further, the contact would become void if no mineral production were completed in the grant's first five years. After assay results from Pahaquarry ore samples proved uneconomical no more work was done. The grant became null and void and the mine went dormant.

1847-48

Interest in the Pahaquarry copper mineralization again increased partly due to its elevated need to supply increased brass manufacturing. Land acquisition occurred in 1845-46 followed by the formation of the Alleghany Mining Company in 1847. Mine development began with lengthening of the old adits, digging of new adits and other prospect openings. 10 total sites were excavated including lengthening an original 1750s adit and digging 4 additions shallow prospect shafts on the higher topography southwest of Mine Brook (figure 3). Additional diggings included 2 adits, one of unknown length, and a second running 10 feet into the rock before digging a 75 foot cross cut. On the higher topography they excavated a 100 foot long by 15 foot deep cut that led to another new opening. The last 2 shafts consisted of one 15 foot deep and the last was 20 feet deep with a diameter of 7 by 15 feet.

Mine development progressed by methods similar to those employed in the 1750s. Drills were hard-

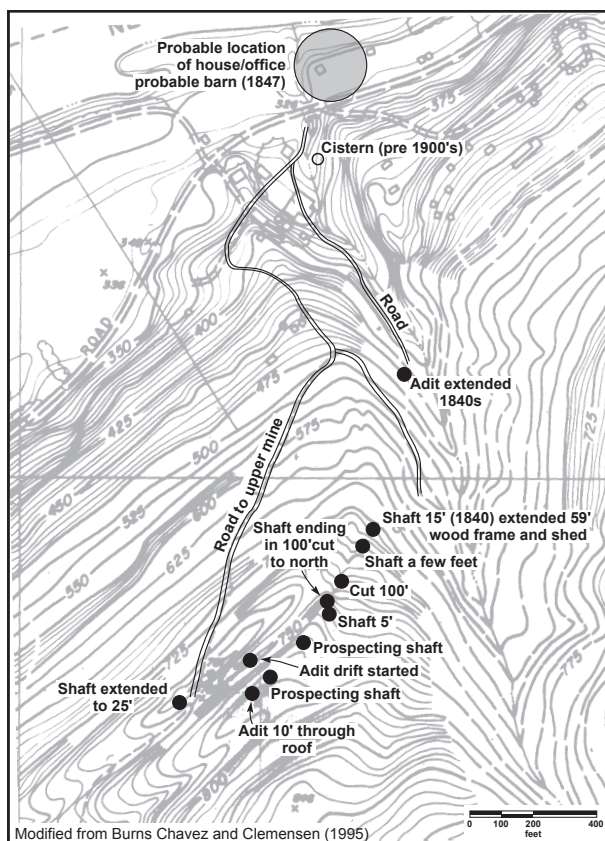


Figure 3. Pahaquarry Mine workings from 1847 to 1862. Workings extended the colonial adit as well as prospected the southwest ridgeline. Modified from Burns Chavez and Clemensen (1995).

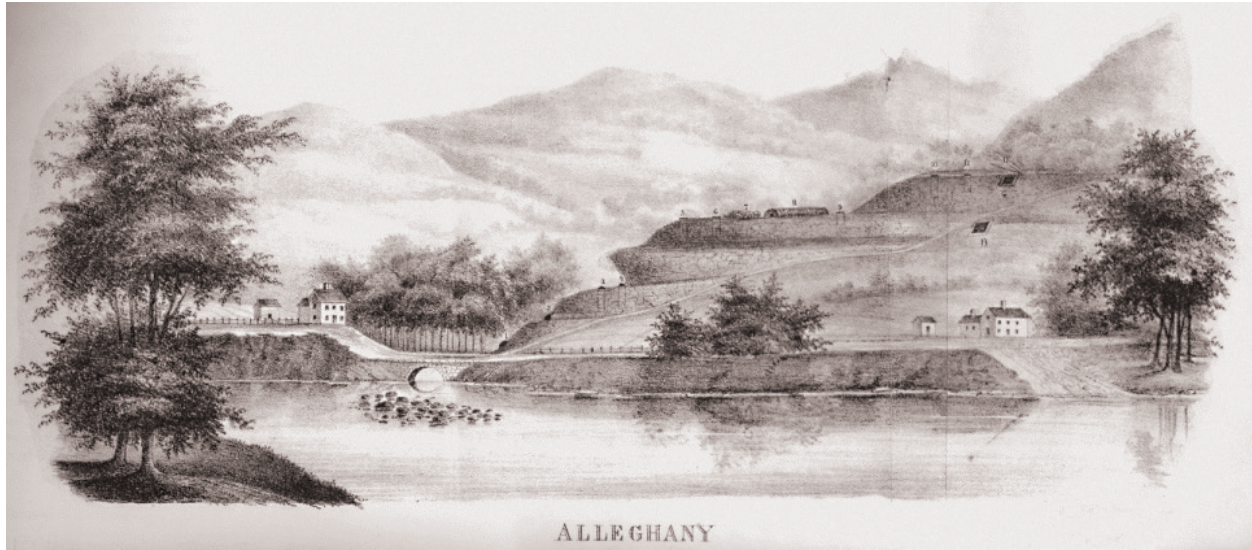


Figure 4. Landscape drawing of the Pahaquarry Copper Mine looking southeast from the Pennsylvania side of the Delaware River. Numbers locate both the adit along Mine Brook and vertical shafts above and to the southwest (right side of drawing). Drawing from Dickeson (1862; reprinted in Burns Chavez and Clemensen, 1995).

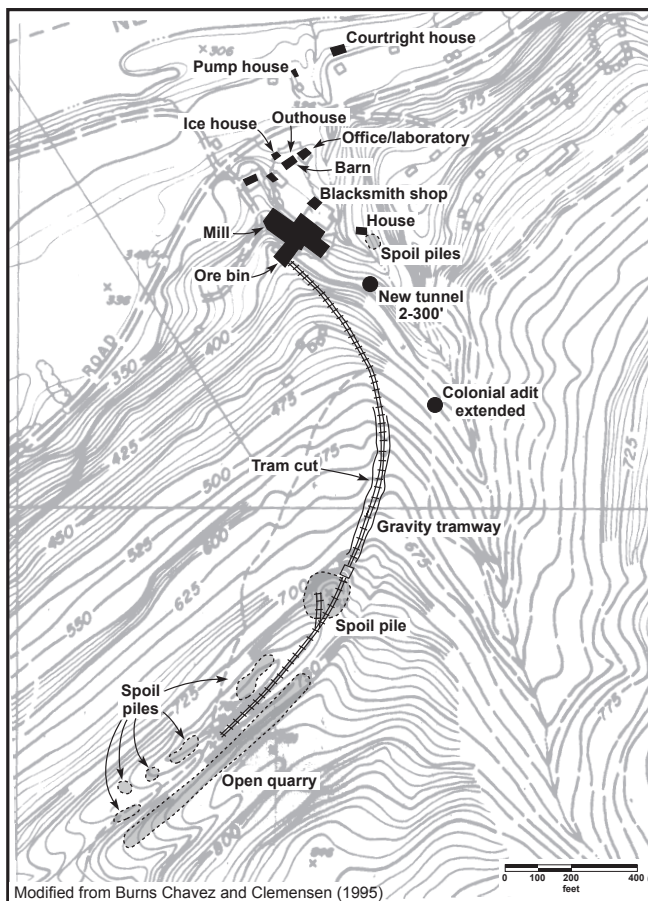


Figure 5. Pahaquarry Mine workings from 1901 to 1912. Building construction including the large mill was a major part of Pahaquarry development as well as the start of open pit mining. Modified from Burns Chavez and Clemensen (1995).

ened steel now, which increased their usage but still employed a single jack. Blasting still used black powder. Advancement in black powder fuses made blasting less dangerous. All the ore was transported by wagon to Flemington, New Jersey where one mine owner had a mill and smelter near a second copper mine. Alleghany constructed additional buildings to support the copper mining labor force. Mining ended again in 1848 due to lack of profitability.

1861-62

The Civil War induced the resurrection of the Alleghany Mining Company in 1861 under a new directorship. An economic geologist worked on site and planned its further development. A new wood frame covered an 1847 shaft, which was extended from 20 to 59 feet deep. A second shaft was deepened from 15 down to 25 feet. Figure 4 shows an artistic depiction of the Alleghany Mining Company's Pahaquarry mine landscape and workings at this time (Dickeson, 1862 from Burns Chavez and Clemensen, 1995). Some amounts of ore were collected. Even with an overly optimistic interpretation of the mineral potential (Dickeson, 1862 as described by Zdepski, 2002) at Pahaquarry its lean grade defeated this latest attempt to develop the copper deposit.

1901-12

The last stage of the Pahaquarry Copper Mine began at the turn of the 20th century. The increased demand for copper-based electrical

wiring renewed interest in the Pahaquarry copper deposit. In 1901, the Montgomery Gold Leaf Mining Company, which owned a gold mine in Pennsylvania, pursued further development in the Pahaquarry mine. They entered into a 1-year lease agreement with the Alleghany Mining Company. They excavated a new adit into the hillside and removed 100 tons of ore (figure 5). Ore value was insufficient to even cover the wagon hauling costs over the seven miles to the railroad depot. Even with this low grade ore, Montgomery wanted to further develop the mine. So in 1902, Montgomery bought the Alleghany Mining Company lands at sheriff sale. They constructed numerous structures including a powder house, blacksmith, barn, oil and icehouses. By 1904, the adit reached 300 feet in length (figure 6).

Mining technological advances allowed easier ore extraction at this time. Gas driven generators supplied power for mechanical drills. Advances in blasting methods also moved forward to facilitate excavation. Dynamite was now used instead of unstable black powder and its poor quality fuses. Small hand detonators could ignite the dynamite with a small electrical current.

Now as before the mining problem rested in the low grade of the Pahaquarry ore. Montgomery decided that concentrating the ore was the answer. They decided to build a concentrating mill designed by one of Thomas Edison's former advisors Dr. Nathaniel Shepard Keith. Montgomery's gold mine was no longer operational so they reorganized as the Pahaquarry Copper Company. The following two years saw increased construction, most of all, the new mill.

Meanwhile exploration continued above the mine along both the northeastern and southwestern ridges atop Mine Brook. A long exposure of the mineralized host rock was discovered along the southwestern ridge. This find allowed an open pit mining stage at Pahaquarry. To accommodate the ore removal from the new open pit a tramway was constructed that brought the ore down hill to the mill in self-emptying cars. Tramway construction required remodeling of the mill. All the construction finished in 1911. The quarry then operated for three months and the mill for two. The quarry grew to 2,000 feet long and between 30 and 40 feet wide (figure 7 and 8). Unfortunately the new mill did not sufficiently concentrate the ore to make a profit. A floatation concentration process used in the mill could not capture the finer size fraction of ore minerals that dropped out in the waste material. Concentrated copper ore was sent to the smelter and produced only 3 copper ingots for a total value of \$15.00, not nearly enough to maintain a mining concession. More recent research suggests the possibility of more ingots in existence (Zdepski, 2002). Another new concentrating process was attempted which also failed. This ended copper mining at Pahaquarry in 1912.



Figure 6. View looking southwest of the colonial adit entrance that was extended during 1901-1912 mining. The adit entrance is gated and usually locked. Bloomsburg Red Beds exposed consist of gray medium bedded, quartz sandstone and less common thinner bedded siltstone and shale. Bedding dips towards the northwest and cleavage towards the southeast.

GEOLOGY

Regional

Kittatinny Mountain geology is dominated by the Silurian aged Shawangunk Formation and Bloomsburg Red Beds (Epstein, 1992; Epstein and Epstein, 1972). The Middle Silurian Shawangunk consists of light gray quartz pebble conglomerate, quartz sandstone and minor shale. Cuts on both sides of the Delaware Water Gap beautifully expose the Shawangunk's three members. These consist of an upper and lower conglomerate, sandstone and quartzite facies (Tammany and Minsi members, respectively) separated by an intervening gray shale-dominated Lizard Creek Member (Epstein and Epstein, 1972). Bedding generally dips northwest though there are many small folds and faults cutting the Shawangunk. Regionally the Shawangunk averages 1,400 feet thick. These coarse clastic sediments are resistant to weathering and

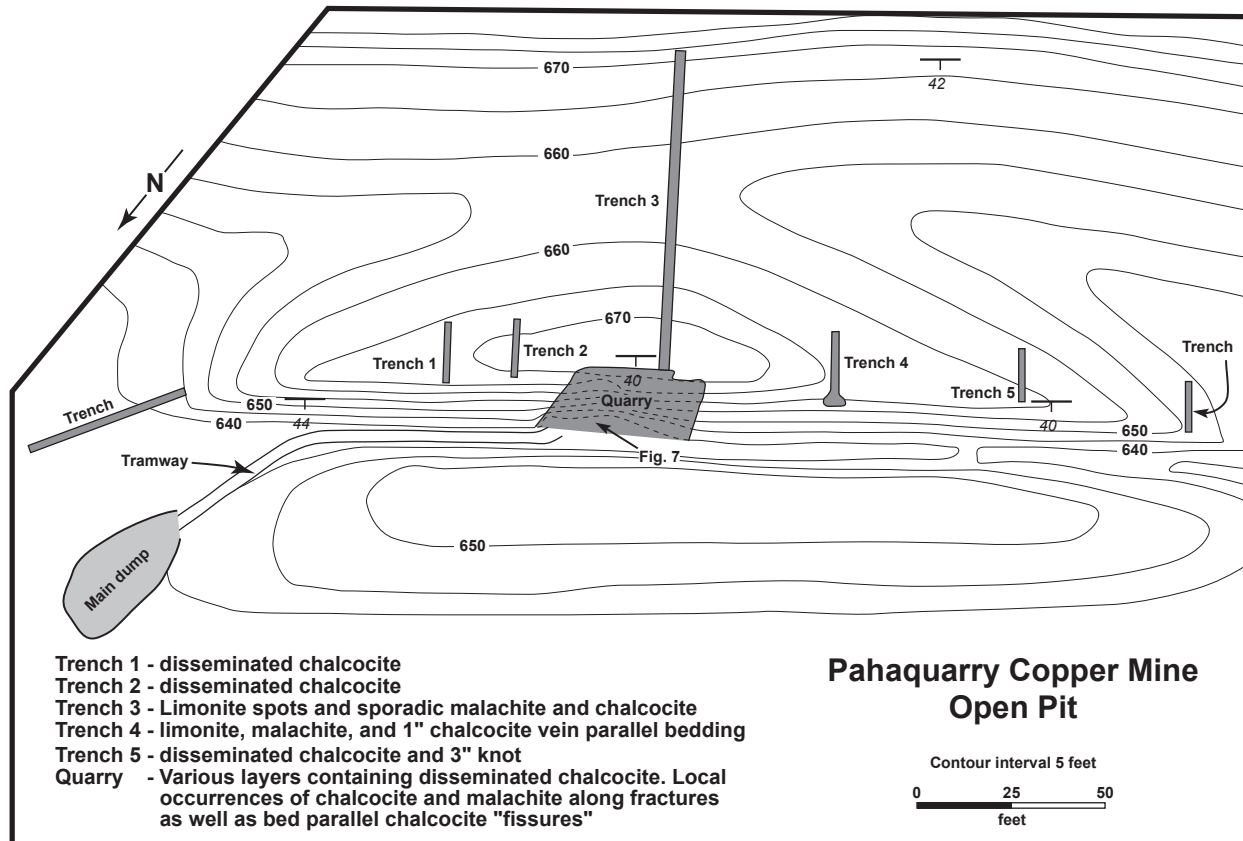


Figure 7. Detail of the open pit quarry and trench above and southwest of the adit as outlined by the USGS for a strategic mineral assessment. The open pit dates to the 1901-1912 period while the trenches probably are part of the 1847-1862 development period. Modified from Cornwell (1945).

support the main Blue-Kittatinny-Shawangunk mountains' ridgeline that traverse Pennsylvania, New Jersey and New York, respectively. Even though the Shawangunk holds up the highest regions of New Jersey it was not deposited in such lofty regions. Supplied by clastic debris from a weathering eastern mountain source, Shawangunk deposition occurred under braided stream and marginal marine paleoenvironmental conditions along a northwest facing coastline (Epstein and Epstein, 1972).

The Middle to Upper Silurian-aged Bloomsburg conformably overlies the Shawangunk throughout the Blue-Kittatinny-Shawangunk mountain chain's length. Red and less common gray and green, sandstone, siltstone, shale, and minor conglomeratic sandstone, arranged in repetitive fining-upwards cycles characterize the Bloomsburg. Sandstones commonly have an erosive base that upwards contains cross bedded to laminated sandstones. Siltstones overlie the sands and gradually grade upward into shale that may be mudcracked. Bloomsburg deposition occurred in a shallow to marginal marine paleo-environment. Their fining-upwards cycles represent the effects of rising and lowering relative sea level during deposition. The sediments can be burrowed mottled and locally fish scales occur. A fish scale location exists at mile 15.4 on the alternate road log, south of the Pahaquarry trail parking lot. Parris and others (1987) found dorsal and ventral plates of the ostracoderms *Vernonaspis* and *Americaspis* there. Bloomsburg thickness is approximately 1,500 feet thick. Bloomsburg sandstones may be clean, quartz-rich sands equally resistant to weathering as the Shawangunk or clayey sandstones that erode and underlie the lower topographic locales on Kittatinny Mountain.

The Bloomsburg and Shawangunk formations have similar deformational histories. Due to their relatively similar rock strengths and combined thickness they reacted to the northwest-directed progressive strain of the Alleghenian orogeny in a uniform way. The weaker siltstone and shales below and limestone units above more readily folded and faulted than the stronger more resistant Shawangunk-Bloomsburg rock package. This allowed Epstein and others (1967) and Epstein (this volume) to divide these rocks into different lithotectonic units according to their combined rock strength and subsequent reaction to the

Alleghanian applied stresses. Lithotectonic unit 2 containing the Shawangunk and Bloomsburg displays broad, open to locally overturned folds that commonly exhibits flexural slip. Wedge and thrust faulting are common as seen on both sides of the northern section of Delaware Water Gap. Cleavage is better developed in the Bloomsburg due to its higher clay content than the Shawangunk. Regional cleavage dips to the southeast with a moderate to steep dip.

ORE GEOLOGY

The host Bloomsburg beds dip northwest forming a dip slope (average bedding orientation is 053/42NW) that has been bisected by the Mine Brook drainage. A cross sectional view of the layers was exposed when this brook downcut to its present level. Interbedded layers of gray clean sandstone, red and gray clayey sandstone and reddish siltstone and shale are well exposed along the brook. Sandstone beds dominate the area. Excellent glacially polished examples of sandstone can be seen in the quarry on the upper ridge southwest of Mine Brook. There, the gray sandstone is medium to thick bedded, cross-bedded to massive. Joints cut the sandstone and parallel the well developed cleavage in the finer grained siltstones and shales. Fining-upwards sedimentary cycles repeat throughout the mine region. Sandstones dominate the overall cycles, as shale layers tend to be thin.

Early workers identified chalcocite (Cu_2S , copper sulfide) as the main copper bearing mineral in the mine. Secondary copper minerals, malachite ($\text{Cu}_2(\text{CO}_3)(\text{OH})_2$ copper carbonate hydroxide) and chrysocolla ($\text{CuSiO}_3 \cdot n\text{H}_2\text{O}$, hydrated copper silicate), probably first caught the eye of early prospectors. Examples can still be found in the rock exposures and along old dump piles of these secondary minerals. They coated the bedrock along bedding planes and certain joint surfaces and were restricted to exposed rock surfaces, penetrating only slightly into the rock. The chalcocite is disseminated in select gray sandstone beds and very difficult to discern without magnification. Chalcocite may also form either as thin seams paralleling bedding or joints, or as irregular shaped patches and/or nodules several inches or up to a foot long that has partially replaced the host Bloomsburg bed. Thickness of the copper bearing Bloomsburg may be as much as 200 feet thick. The highest reported natural samples contained 3.25% copper, but the entire copper-bearing horizon was a much lower percentage.

Pahaquarry is thought to be an epigenetic deposit (Woodward, 1944). The copper was believed disseminated throughout the original beds as a primary detrital grain. Waters high in salt (sodium chloride) and gypsum (calcium sulfate) remobilized the copper into solution. The copper was reprecipitated in nearby sandy horizons supporting a suspected more highly acidic condition. Temperature levels are thought not to have exceeded 91° C. Later, meteoric water interaction aided the growth of the secondary minerals, malachite and chrysocolla (Woodward, 1944) that can still be found near the workings.

CONCLUSION

Pahaquarry Copper Mine has gone through an extensive history. Local lore describing the native American and early Dutch exploration of the mineral deposit still persists. More recent historical investigations do not substantiate this early description but instead suggest that initial exploration only goes back to the 1750's (Burns Chavez and Clemensen, 1995; Kraft, 1996). The ore grade never proved high enough to maintain as an economic deposit. Today the area is preserved within the Delaware Water Gap National Recreation Area. The forests have reclaimed the workings from the more open landscape depicted by Dickeson (1862, figure 4).

REFERENCES



Figure 8. Pahaquarry open pit photographed towards the northeast on ridge above and southwest of Mine Brook and the adit.

- Burns Chavez, S.R., and Clemensen, A.B., 1995, Pahaquarry Copper Mine Delaware Water Gap, final cultural landscape report, v. 1, U.S. Department of the Interior, National Park Service, Denver Service Center, 373 p.
- Dickeson, M.W., 1862, Report of the geological survey and condition of the Alleghany Mining Company's property, Warren County, New Jersey, with map and drawing, Philadelphia, PA, 30p.
- Epstein, J.B., 1993, Stratigraphy of Silurian rocks in Shawangunk Mountain, southeastern New York, including a historical review of nomenclature: U.S. Geological Survey Bulletin 1839L, 40p.
- 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, A.G., Epstein, J.B., Spink, W.J., and Jennings, D.S., 1967, Upper Silurian and Lower Devonian stratigraphy of the northeastern Pennsylvania, New Jersey and southeasternmost New York, U.S. Geological Survey Bulletin 1243, 74p.
- Garber, J.P., 1917, The Delaware River prior to the coming of Penn, *in*, Philadelphia History consisting of papers read before the City History Society of Philadelphia, Philadelphia, PA, p.127-161.
- Kraft, H.C., 1996, The Dutch, the Indians and the quest for copper, Seton Hall University Museum, South Orange, NJ, 183p.
- Monteverde, D.H., 2001, Pahaquarry copper mine, *in*, Inners, J.D., and Fleeger, G.M., eds., 2001 – A Delaware River odyssey, Guidebook, 66th Annual field conference of Pennsylvania Geologists, Shawnee-on-Delaware, PA, p.150-155.
- Weed, W.H., 1911, Copper deposits of the Appalachian States, U.S. Geological Survey, Bulletin 455, p.181-186.
- Woodward, H.P., 1944, Copper mines and mining in New Jersey, New Jersey Department of Conservation and Development, Geologic Series Bulletin 57, 156 p.
- Zdepski, M., 2002, Pahaquarry Copper Mine History relating to existing surface features, *in*, D'Amato, D., editor, Geology of the Delaware Water Gap Area, field guide and proceedings, 19th Annual Meeting of the Geological Association of New Jersey, Shawnee-on-Delaware, NJ, p 68-74.

THE TOCKS ISLAND DAM AND OTHER DAMS PROPOSED FOR THE DELAWARE

by

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The Delaware River (figure 1) has been described as the only major free flowing river in the United States east of the Mississippi. This is not from lack of dams on tributary streams, dam proposals for the main stem, or adequate dam sites. Also, it does not imply an absence of water management within the basin. Proposals for dams along the Delaware have been under discussion from early in the 19th century, dams have been proposed at many locations, and there are numerous dams on tributary valleys for water supply, flood control, and low flow augmentation. The main stem of the Delaware, however, originates along the New York-Pennsylvania state line at the confluence of its East and West Branches, and follows state lines all the way to Delaware Bay. Proposals were seldom seen as equally beneficial to states on both sides of the line, and were repeatedly blocked by political interests in the seemingly disadvantaged state. Many of the projects were blocked using a 1783 treaty between New Jersey and Pennsylvania which prohibited dams by declaring the Delaware to be "... a common highway, equally free and open for use, benefit, and advantage of each state". Only wing dams (which did not extend across state boundaries), and the Lackawaxen Dam (part of the Delaware and Hudson Canal) were completed. The Lackawaxen Dam was built in 1825, increased in height in 1848, and maintained until the canal was abandoned in 1898.

The most recent dam proposal for the main stem, authorized by Congress in 1962, was for a large multipurpose dam at Tocks Island, a short distance upriver from the Delaware Water Gap. Geologic conditions at Tocks Island were not ideal, but were amenable to engineering solutions. The project initially enjoyed wide public support, and the 1783 New Jersey – Pennsylvania treaty had been repealed by both states. Construction appeared certain. Repeated delays, however, mostly due to budget shortfalls, pushed the scheduled groundbreaking to 1975. By this time, the project had become politically untenable. Within weeks of the scheduled groundbreaking, the governors of New Jersey, New York, and Delaware went on record as opposing the dam. Subsequent to rejection of the Tocks Island Dam, Delaware Basin water resources have been managed based on a "Good Faith" agreement administered by the Delaware River Basin Commission.

This history is presented in "Damming the Delaware," by Richard C. Albert. The unattributed historical material in this article is summarized from that thorough treatment. The reader is referred to Mr. Albert's book for further detail.

WING DAMS AND CANAL DIVERSIONS - 1800-1840

The 1783 treaty was intended to protect navigation, primarily lumber rafting, but remained in effect long after the last lumber raft traveled the Delaware in 1923. In the early 1800s, lumber rafting grew rapidly at the same time as numerous wing dams and canals were built. Although the treaty applied to wing dams and diversions as well as larger dams, both New Jersey and Pennsylvania violated the treaty by unilaterally permitting the construction of numerous wing dams to serve mills and divert water to canals. The dams and canals were built despite vigorous legal and physical protest.

One of the higher profile legal disputes ensued in 1815 from New Jersey's approval of a wing dam. The dispute appeared to be headed for the U. S. Supreme Court until a survey of wing dams between Trenton and Belvidere found that most were on the Pennsylvania side of the river and most interfered with navigation.

In 1823, the 72-mile long Lehigh Canal opened to transport anthracite to the Delaware at Easton, and work had begun on Delaware Canal, paralleling the Delaware River downstream from Easton. The Delaware Canal would require diversion of Delaware River water at Easton, and this diversion was strenuously contested by New Jersey. Despite the unilateral Pennsylvania diversion, the 1824 New Jersey charter for the Delaware and Raritan Canal included a provision that, because of the treaty, Pennsylvania's approval was required for water diversion. Pennsylvania set unreasonable requirements for consent, and

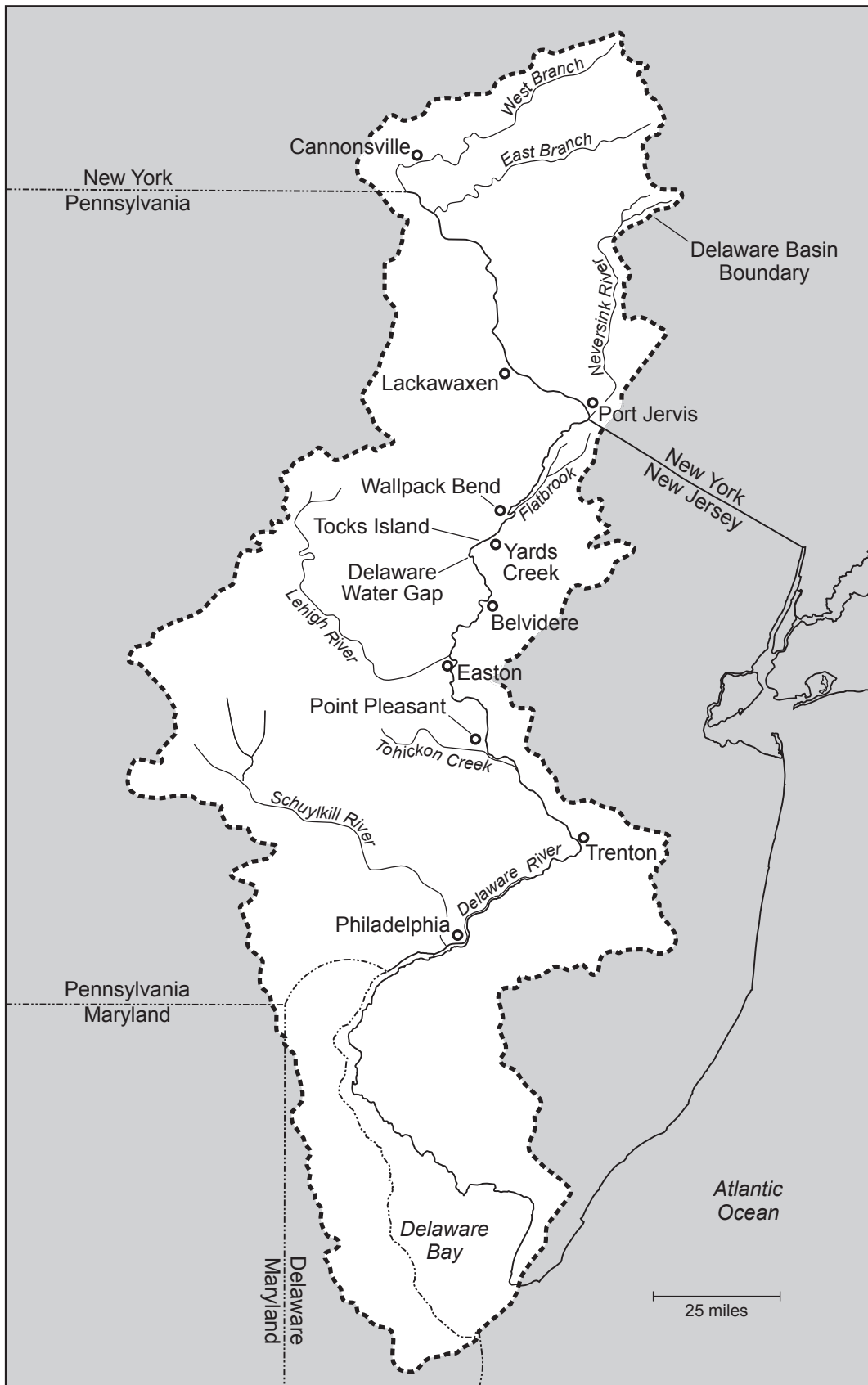


Figure 1. Map of Delaware River drainage basin showing features mentioned in text.

the originally chartered company did not build the canal. In 1828, the Circuit Court for New Jersey found that Pennsylvania's consent for the water diversion was not required so long as navigation was not affected. By this time a different company had already completed the canal.

Opposition was more physical at the Lackawaxen Dam, built to allow boats on the Delaware and Hudson Canal to cross the Delaware. The dam was a major hazard for lumber rafters, and the rafters made their annoyance known by repeatedly ramming canal boats, and by fighting, beating, and sometimes shooting the boat operators. In 1829, explosives were used to destroy an 80-foot section of the dam (Dale, 1996). This opposition and persistent delays at the crossing led to the 1849 construction of the Roebling Aqueduct, safely above the lumber rafters. The dam was not decommissioned, but was instead raised from a 7-foot-high slack-water barrier to 16 feet, and equipped with a chute for the rafts (Dale, 1996). It remained in place until the abandonment of the canal in 1898.

Following completion of the canals and the ascendance of steam over water power, there was limited interest in Delaware River dams and diversion for several decades. The formerly controversial topic became dormant until hydroelectric power generation became feasible.

HYDROELECTRIC GENERATION PROPOSALS - 1894 – 1920

New Jersey was the first state to become involved in evaluation of the Delaware for hydroelectric generation with the publication in 1894 of a survey of the water supply and water power potential of New Jersey rivers (Vermeule, 1894). In 1897, New Jersey's legislature enacted a law allowing companies to be formed for the construction of dams for power generation. The law clearly included the Delaware as dams were limited to a height of 10 feet and were required to have chutes for lumber rafts and shad.

The first survey of the Tocks Island site for dam construction may have been for a 1902 proposal for a "Delaware Water Gap Power Company". The company appears, however, to have been the summer project of wealthy, bored vacationers at a local resort. The proposal was not pursued further.

The New York State Water Supply Commission put forth the first serious hydropower proposal in 1907. The proposal included four dams: three hydroelectric dams across the Delaware and a larger water regulation dam and lake near Cannonsville. An even larger concurrent proposal included forty power supply dams upstream from Port Jervis. A major problem for both proposals was the lack of enough local customers. Construction was never begun on either project.

A more realistic hydropower proposal involved a dam near Belvidere. This raised the issue of the 1783 treaty. In the opinion of Pennsylvania's Attorney General, "While this agreement remains in force, the only method by which permission could be legally secured to dam the Delaware River, would be by concurrent legislative action by both states." Despite efforts over an 8-year period, legislative action was not forthcoming, and the dam was not built.

The most ambitious power generation proposal to involve the Delaware was the "super-power system" proposed in 1918 by the president of the New York, New Haven, and Hartford Railroad. This was a system of hydroelectric dams and steam generators stretching from Boston to Washington. The primary customers were to be railroads, and the power would have allowed electrification of 19 thousand miles of tracks in the region. Six flow regulating reservoirs and sixteen power supply dams were proposed for the Delaware basin. Probably because of costs exceeding one billion 1920 dollars, the system was never built.

While hydropower continued to be a consideration in dam proposals up to and including Tocks Island, it became clear through a series of proposals and controversies in the 1920s that the water supply needs of the region were more important than hydroelectric power needs. Water supply and stream flow maintenance became the primary focus of water management, with hydropower, flood control, recreation, irrigation, and navigation assuming lesser importance.

PUBLIC WATER SUPPLY

The major public water supplies depending in whole or part on Delaware River water are Philadelphia, New York City, and northern New Jersey. Public water supply came to these areas beginning about 1800 in response to devastating fires and water born diseases that killed thousands each year. One fire in New York, for example, in 1835, burned 700 buildings and bankrupted every insurance company operating in the city. The oldest water supply system in the area is Philadelphia's, funded and built according to Benjamin Franklin's will, and opened in 1801. The system initially used water from the Schuylkill River, and later the Delaware. Eventually about half of the water was from the Delaware. Both rivers were polluted

to begin with, and pollution worsened as Philadelphia and upstream cities grew, industrialized, and built sewers. Alternatives to the grossly polluted local supplies began to be investigated by the 1850's with reservoirs proposed in neighboring Bucks and Montgomery Counties. One of the Bucks County proposals, in an 1883-85 series of studies, envisioned, initially, a water powered pump station at Point Pleasant on the Delaware and, eventually, reservoirs upstream from the Delaware Water Gap. The water was to be pumped from Point Pleasant, and eventually from the Poconos, to a reservoir on Tohickon Creek, then delivered to Philadelphia. While Philadelphia's engineering staff recommended this solution and continued to recommend upstream water sources in numerous reports at least until the 1970s, the recommendations were not supported until the 1950s in part because there continued to be influential citizens of the opinion that cheaper local water could be made fit for drinking by filtration. Sand filters were eventually installed between 1899 and 1911. They removed the threat of water-borne diseases, but did little to help the taste and odor. The water supply remained offensive and the butt of jokes until decreased contamination and improved treatment became effective in the 1950s.

Public water supply was a later development in New York City, with the first water delivered from the Croton Reservoir in 1842. The Croton system was repeatedly enlarged from then until 1906, when the massive New Croton Dam was completed. Further enlargement was prohibited after 1906 by the Smith-Dutchess County Act, which forbid additional water diversions from counties in the vicinity of New York City. This focused New York's attention first on the Hudson basin portion of the Catskills, then, by 1920 when the Catskill reservoirs proved inadequate, on the Delaware Basin.

Like New York City, northern New Jersey began looking to the Delaware Basin for water in the 1920s. A 1921 study proposed four alternatives, two of which depended on Delaware River water. The first of these proposed a reservoir in the Raritan Basin above Somerville, which would meet needs for some time. When increased needs required additional water, the system was to be augmented by pumping Delaware Basin water into the Raritan basin near Clinton. The other proposal, known as the Long Hill project, called for eight reservoirs to be built over many decades as need required. The final installation was to be a pump station on the Delaware near Wallpack Bend capable of delivering 1,500 mgd. This was twice the projected need for 1970.

A 1925 study recommended, instead, a reservoir at Chimney Rock, eventually to be fed by New Jersey's Delaware tributaries. A pump station was projected to be necessary eventually on the Delaware at Belvidere.

Following abandonment in 1934, the Delaware and Raritan Canal was proposed in 1938 for diversion of Delaware River water to the Raritan Basin. Of New Jersey's Delaware River diversion proposals, only the Delaware and Raritan Canal diversion was carried out. That diversion was not begun until the mid 1950s.

By the mid-1920s, in order to overcome numerous interstate disputes, a tri-state commission had proposed a water sharing scheme to: 1) allow each state a proportional share of Delaware River water, and 2) maintain minimum downstream flows. Objections were raised, however, that New York City would be the immediate beneficiary of approval as its plans for diversion (involving reservoirs on the East and West Branches of the Delaware and on the Neversink) were furthest along and that Pennsylvania was short-changed because the formula for water division did not recognize that a large percentage of the Delaware basin is in Pennsylvania. Additional objections to this and other plans were raised by the Lehigh Coal and Navigation Company, which had been granted diversion rights to the Lehigh River during building of the Pennsylvania Canal one hundred years earlier. This company had influential Pennsylvanians as friends and a corporate interest in seeing that potentially competing entities did not develop water systems along the Delaware. While New York approved the proposal, New Jersey and Pennsylvania did not.

A revision which overcame some of the objections was strongly opposed by the City of Trenton, which contended that the Delaware would become a "mere brook" during the summer months. More than quantity, water quality was at issue. The Trenton water supply was experiencing severe quality problems when thunderstorms flushed industrial wastes and coal debris from the Lehigh watershed, and city officials worried that water quality under the additional diversions would force weeks of shutdown or even abandonment of the Delaware intakes. Again the compact did not pass the three legislatures as required. At this point, New York City sought and obtained legal opinions that it was entitled to a share of Delaware Basin water and, at the same time, notified New Jersey and Pennsylvania that it intended to build dams in the Delaware Basin. New Jersey initiated a lawsuit to bar the New York diversions. Pennsylvania soon entered the case, not to bar the diversions but to ensure an equitable allocation for itself.

In 1934, the Supreme Court recognized New York's right to divert water from the basin, with Justice Oliver Wendell Holmes declaring that:

"A river is more than an amenity, it is a treasure. It offers a necessity of life that must be apportioned among all those who have power over it. New York has the physical power to cut off all water within its jurisdiction. But clearly the exercise of such power to the destruction of the lower states could not be tolerated. And on the other hand, equally little could New Jersey be permitted to require New York to give up its power altogether in order that the River might come down undiminished. Both States have real and substantial interests in the River that must be reconciled as best they may be. The effort is always to secure an equitable apportionment without quibbling over formulas."

The decision went on to outline rules that would govern Delaware River diversions for decades, but New York was unable to complete the proposed reservoirs initially because of budgetary constraints during the depression, then because of World War II.

Water supply plans for New York, New Jersey, and Philadelphia continued to be proposed and debated in the following decades, particularly after 1934, when the first of a series for "308 Reports" covering the Delaware Basin was published. The 308 Reports were prepared pursuant to House Document 308, authorizing the U.S. Army Corps of Engineers to evaluate rivers where power development might be feasible in conjunction with navigation, flood control, and irrigation needs. The 308 Reports for the Delaware Basin included water supply, even though this was not specifically authorized, because this was "...the most important consideration in connection with any large-scale development of the river." (U.S. Congress, 1934). The federal government was not significantly involved in water supply projects at this time, and, due to limited navigation and lack of serious flooding prior to 1955, involvement of the Corps on the main stem above Trenton was minimal following publication of the 308 reports. The Corps had been, and continued to be, involved in navigation projects in the lower Delaware Basin (including the channel dredging, the Chesapeake and Delaware Canal, and harbor improvements) and flood control in tributary basins (including the Lehigh and Lackawaxen basins).

While proposals continued to be debated, actual work on major water supply projects in the basin was not resumed until the mid-1950s. At this time, New York was prepared to begin work on upper Delaware reservoirs redesigned from its 1920s proposal, New Jersey was prepared to begin diversions to the Raritan Basin by way of the Delaware and Raritan Canal, and Pennsylvania and New Jersey were moving towards cooperation on a dam at Wallpack Bend. Neither the New York project nor the canal diversion were in accord with the allocations approved in the 1931 Supreme Court decree, and work was delayed until Court approval in 1954 of an agreement negotiated by New York, New Jersey, and Pennsylvania.

THE TOCKS ISLAND DAM PROPOSAL

Most major water supply proposals subsequent to 1885 for Philadelphia and 1921 for northern New Jersey called for dams or pumping stations to eventually be constructed on the upper Delaware. Tocks Island was not generally the proposed location. In a number of the studies, dams were proposed at Wallpack Bend, about 10 miles to the north. At Wallpack Bend the structure could be smaller and anchored into rock. By 1931, it was generally assumed that a dam would eventually be built at Wallpack Bend, and this assumption was embodied in provisions of the 1931 Supreme Court decision. The first serious evaluation of Tocks Island as a dam site was in the Corps of Engineers 308 Report released in 1934. The report found it to be promising. The Wallpack Bend site was more favorable for construction, but did not maximize reservoir capacity. Reservoir capacity along this section of the Delaware is limited by the expense and political impracticality of flooding Port Jervis to the north and the Water Gap to the south. Within these constraints, moving the dam downstream from Wallpack Bend to Tocks Island gained a relatively small amount of storage by increasing the length of the reservoir along the main stem Delaware valley and a substantially greater volume of storage by capturing tributary storage in the Flatbrook valley.

Further federal involvement was limited, until 1955, to test borings completed in 1942 for a navigation project. A dam at Tocks Island was under consideration as a source of fresh water to forestall contamination of aquifers by salt water along a proposed sea level canal from Bordentown to Raritan Bay. Borings to 140 feet did not reach rock, and the site was rejected as unfeasible or excessively costly.

By the early 1950s, preparatory work towards a dam at Wallpack Bend was proceeding quickly. In

Pennsylvania, population and water use were expanding rapidly. In New Jersey, the Delaware and Raritan Canal diversion had not addressed northern New Jersey's water needs. By 1955, both states had repealed the anti-dam provisions of the 1783 treaty, an extensive study of the benefits of a dam at Wallpack Bend had been completed, and it had become evident that New Jersey might work in partnership with Pennsylvania on such a dam.

The project changed dramatically in 1955. The year started out very dry, and by mid-August a severe drought appeared imminent. This changed quickly. On August 12, Hurricane Connie dumped up to 12.5 inches of rain in the Delaware watershed. On August 18 Hurricane Diane dumped an additional 11 inches of rain on the already saturated soils. The ensuing floods claimed one hundred lives along Delaware River tributaries and caused many millions of dollars in property damage. The flooding did not cause planning for a Delaware River dam. That planning was already far along. It did cause a widespread call for flood control dams and, by placing the Delaware within the Corps of Engineers flood control mandate, it paved the way for massive Federal involvement.

Responding to the calls for flood control and a thorough re-evaluation of water management options, the Corps began a large and uniquely comprehensive survey of the Delaware Basin in 1956. A Tocks Island portion of the study was released in 1957 indicating that an earth fill dam was feasible. In comparison with the Wallpack Bend site, capacity would be doubled and costs would only be increased by 50 to 60 percent. From this point, the Wallpack Bend site was no longer seriously considered.

Even though the report addressed only the dam's practicality in comparison with the Wallpack Bend proposal, and did not evaluate its economic feasibility or relationship to other water management projects in the basin, there was widespread call for immediate construction. Additional evaluations were completed by January 1959, and a report was released recommending the construction of 5 dams (including Tocks Island) before 1980 and the construction of sixteen dams as needed afterward. Tocks Island was by far the largest of the dams, and was a multipurpose structure providing water supply, low-flow augmentation, flood control, hydroelectric power, and recreational benefits. By November 1961, the four Delaware River Basin Commission states (New Jersey, New York, Pennsylvania, and Delaware) and the federal government had agreed to the plan.

Intensive geologic investigation of the site began following Congressional authorization in 1962 and continued through several years. A few borings were available from the 1930s, 1942, and the 1950s. Much more work was needed prior to construction. Ultimately 20 or more miles of seismic line and a total of 7 miles of cores were collected. As summarized by Depman and Parrillo (1969) the axis of the dam was to cross the valley above a southwest plunging asymmetric syncline (figure 2). The New Jersey abutment of the dam was to be excavated into the Bloomsburg Red Beds (table 1, recognized as the High Falls Formation at the time of original publication). The Pennsylvania abutment was to overly Silurian and Devonian limestone, shale, and sandstone from the Bossardville Formation through the New Scotland Formation. Engineering significance of the bedrock formations is summarized in table 1. Most of the valley floor is underlain by roughly 200 feet of unconsolidated material. Within the unconsolidated materials, the usual sequence is late Wisconsinan till overlying bedrock, a complex assemblage of late Wisconsinan stratified sands and gravels together with fine-grained lake sediments, then post-glacial alluvial sands and gravels. Significant geologic obstacles to be overcome were:

Instability of the Lacustrine Sediments

Instability of lake sediments underlying the valley affected both the location of the dam and its design. The originally proposed location crossed the valley at the northern tip of Tocks Island. In 1964, it was concluded based on geophysical evidence and a few borings that the dam could not be constructed at the proposed location or at most other locations in this part of the valley because of the predominance of lake sediment. The lake sediments had little strength and were prone to liquefaction. Some cores, in fact, liquefied before reaching the lab. An earthquake or rapid water-level fluctuations could lead to liquefaction and dam failure. The only area in which a dam might be feasible was a 3,000-foot stretch of the valley between the southern tip of Tocks Island and Labar Island.

Further work within this area identified a section of the valley centered about 100 feet south of Tocks Island in which sand and gravel were more abundant than elsewhere. Borings on a 200-foot grid

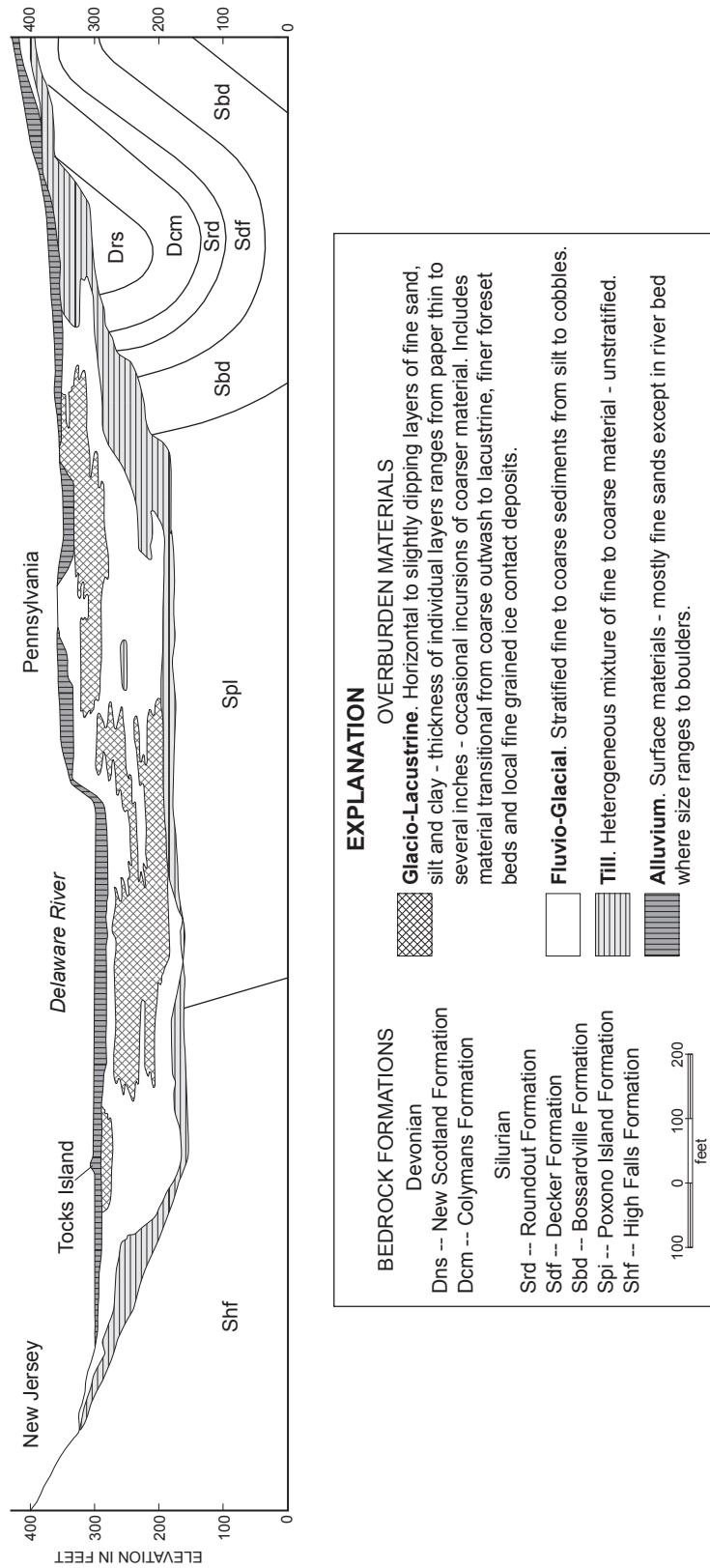


Figure 2. Typical traverse section of dam (after a map from Depman and parrillo, 1969).

within this area generally showed a predominance of till, sand, and gravel and a lesser volume of lake sediment. Till formed a hummocky surface above bedrock and fluvioglacial sediments rose, in some places, completely through the lake sediment. Depman and Parrillo attributed the hummocky till and the abundant gravel to deposition along a stagnant ice margin and identified this same margin to the east and west of the valley fill based on moraines extending diagonally up the valley walls. Between the stagnant margins, lake sediment was predominant over till and gravel and rendered the valley unsuitable for dam construction. Witte (2001) confirmed an ice margin, the Zion Church margin, at this

Age	Formation	Member	Thickness of Unit of Sam Site	Description	Engineering Significance & General Remarks	
Devonian	Marcellus		Not Present at Site	Black, fissile carbonaceous SHALE.	<i>Marcellus</i> -- forms steep walls and extensive talus slopes along right rim of reservoir from Wallpack Bend to Matamoras. Intense wave action and fluctuating water levels may cause local slides & slumping.	
	Onondaga			Dark, thin-bedded SHALE and med.-bedded cherty LIMESTONE.		
	Oriskany			White, f. to c. conglomeratic SANDSTONE.		
	Port Ewen			Gray, calcareous & siliceous silty SHALE.		
	Minisink			Gray, argillaceous LIMESTONE & SHALE.		
	New Scotland	Flatbrookville	35'±	Med. dk. gy. calcareous fossiliferous SHALE & argillaceous LIMESTONE with nodules and layers of chert; cleavage.	<i>Onondaga through Bossardville</i> -- used mainly for stratigraphic & structural correlation. These formations form the left rim of the reservoir from Wallpack Bend to Port Jervis, NY, and the right rim from Wallpack Bend to Tocks Island. All limestones are susceptible to solution activity & potential reservoir leakage; however, no cavernous conditions or areas of large scale solution activity are known to exist. At the dam site these formations will not be involved in any major excavation.	
	Coeymans	Total 82' - 89'		Med. gy. quartzose conglomeratic SANDSTONE w. calcareous cement. Local limestone beds & cherty zones.		
		Stormville Disconformity				
		Shawnee I.	42'	Massive fossiliferous LIMESTONE with reef structure 25' thick overlain by sandy LIMESTONE & BLACK shale; mud cracks.		
		Peters Valley	5' - 7'	Lt. gy. f.-c. calcareous SANDSTONE.		
Depue I.		20'	Med.-thin bedded clayey LIMESTONE.			
Silurian	Rondout	Total 31' - 34'		Thin bedded dk. gy. calc. SHALE & sandy LIMESTONE; bedding undulatory; mud cracks common. Buff, platy weathering.		
		Mashipicong	11' - 12'			
		Whiteport	5' - 7'	Med. bedded to massive DOLOMITE.		
		Duttonville	15'	Interbedded calc. SHALE & LIMESTONE.		
	Decker Ferry	Wallpack Center	55'	Med.-thin bedded sandy LIMESTONE & DOLOMITE, calc. SILTSTONE, SANDSTONE & SHALE. Fossiliferous, reef structures.	<i>Poxono</i> -- underlies valley fill in the reservoir area & at the dam site. Several zones exhibit extensive solution activity & much rehealing.	
	Bossardville		112'	Laminated to thin bedded clayey, v.f. crystalline LIMESTONE; calc. SHALE & SILTSTONE, some beds massive & dolomitic.		
	Poxono Island		675'	Green, red & mottled calc. SHALE, MUDSTONE, dolomitic SHALE, DOLOMITE & LIMESTONE.		<i>High Falls</i> -- forms left rim or reservoir from Wallpack Bend to the dam site. All structure in left abutment will be founded in this formation.
	High Falls (Bloomburg)		1,500'	Red & green mottled SHALE, MUDSTONE, SILTSTONE, SANDSTONE & QUARTZITE; cross bedding common in coarser grained rock.		
	Shawangunk		1,500'	Cross-bedded QUARTZITE & CONGLOMERATE, with interbedded black ARGILLITE.		
	Unconformity or fault					Not of significance.
Ordovician	Martinsburg		Not present at site	SLATE & GRAYWACKE		

Table 1. Stratigraphic column at Tocks Island Dam site and engineering significance of rock types (from Depman and Parrillo, 1969).

location, but interpreted it as left by an active glacier rather than remaining from widespread stagnation. The initial design called for a steep-sloped structure, with run-to-rise of 2.5 to 1 and 3 to 1. This would have required removal of the bulk of the lake deposits. This was considered feasible, but expensive. Also, it could have led to unforeseen problems. Instead, the dam was re-designed as a flat-topped structure with sides at 10 to 1. Instead of being 400 to 900 feet wide, the re-designed dam was to be over 3,000 feet wide. The potential instability of lenses of lake sediment was considered less of a problem under a dam of this width. While a considerable volume of lake sediment would have to be excavated, much could remain in place.

Permeability of Coarse-grained Materials

Significant problems were foreseen from seepage through sand and gravel and, to a lesser extent, solution cavities and bedrock fractures. If not controlled, seepage could have affected stability of the dam, and a costly seepage control program was anticipated (Albert, 1987). Removal of an extensive blanket of permeable glacial deposits on the Pennsylvania bluff was not considered necessary (Depman and Parrillo, 1969). Weathering to a depth of 10 feet had created a naturally impervious blanket of decomposed shale. The costly aspects of seepage control were to include emplacement of an impervious sediment blanket where necessary upstream from the dam, drainage beneath the dam and downstream through strategically placed beds of coarse materials, installation of pressure relief wells, grouting of upslope permeable deposits where they constituted recharge areas, and grouting of bedrock fractures and solution cavities (Corps of Engineers, 1965-1972).

Suitability of Bedrock for Foundations

According to Depman and Parrillo (1969), the spillway location was governed by rock type. In the original design, the spillway and generators were over limestone adjacent to the bluff on the Pennsylvania side. Cores and downhole cameras revealed solution activity in the dam foundation area. Permeable unconsolidated sediments in upslope areas would be a possible source of recharge (Corps of Engineers, 1965-1972). In the final design, the spillway was moved to an excavation in sandstone, siltstone, and shale on the New Jersey side.

Slope Instability

The primary slope failure possibilities recognized during planning for the Tocks Island Dam were: 1) Local sliding and slumping due to wave action or water level fluctuations along steep valley walls and extensive talus slopes where the Marcellus formation borders the Pennsylvania side of the reservoir. 2) Similar sliding or slumping along bluffs consisting of varved lake sediments, also on the Pennsylvania side. And, more significant, 3) Failure along bedding-plane faults in the Bloomsburg Red Beds in excavation walls over 300 feet high at the spillway and intake structures (Corps of Engineers, 1965-1972, Depman and Parrillo, 1969). Epstein (2001) describes, in addition, failures in the valley from soil slips on glacially polished bedrock surfaces, rockfalls along fractures that parallel road cuts, and debris flows in glacial till. While the failures described by Epstein have led to an estimated \$150,000 in repair costs to roads and trails within the park area, they were either not recognized or not considered substantial problems during the Tocks Island planning process.

Rock conditions at the 300-foot high cuts above the intakes and spillway were investigated by driving a 5 by 7 foot adit 600 feet into the Bloomsburg at the location and elevation of the spillway base (figure 3). Numerous bedding plane slips were identified dipping towards the excavation at angles between 17° and 35° (Depman and Parrillo, 1969). Many of these are zones of abundant ground water flow (Epstein, 2001). Similar faults were identified in the Bloomsburg throughout eastern Pennsylvania by Epstein and Epstein (1967, 1969). Regardless of position on a fold, the overriding beds consistently moved from east to west. Two of the faults identified in Depman and Parrillo (1969) were at or near the base of laminated shales and sandstones with deep desiccation cracks. Both showed weathering and decomposition (Depman and

Parrillo, 1969). In a detailed investigation of these two zones, 1) cross cuts were extended 100 feet from the adit along the deeper of the two decomposed zones and at the end of the adit, 2) three 36-inch diameter core holes were drilled from the land surface to the faults to allow in situ observation and strength evaluation, and 3) 4-inch cores were drilled behind the proposed excavation to investigate fault geometry and to install piezometers. The faults were found to be continuous through the area. Stress measurements above and below the deeper of the faults found there to be little strength across the fault, and it was predicted that the

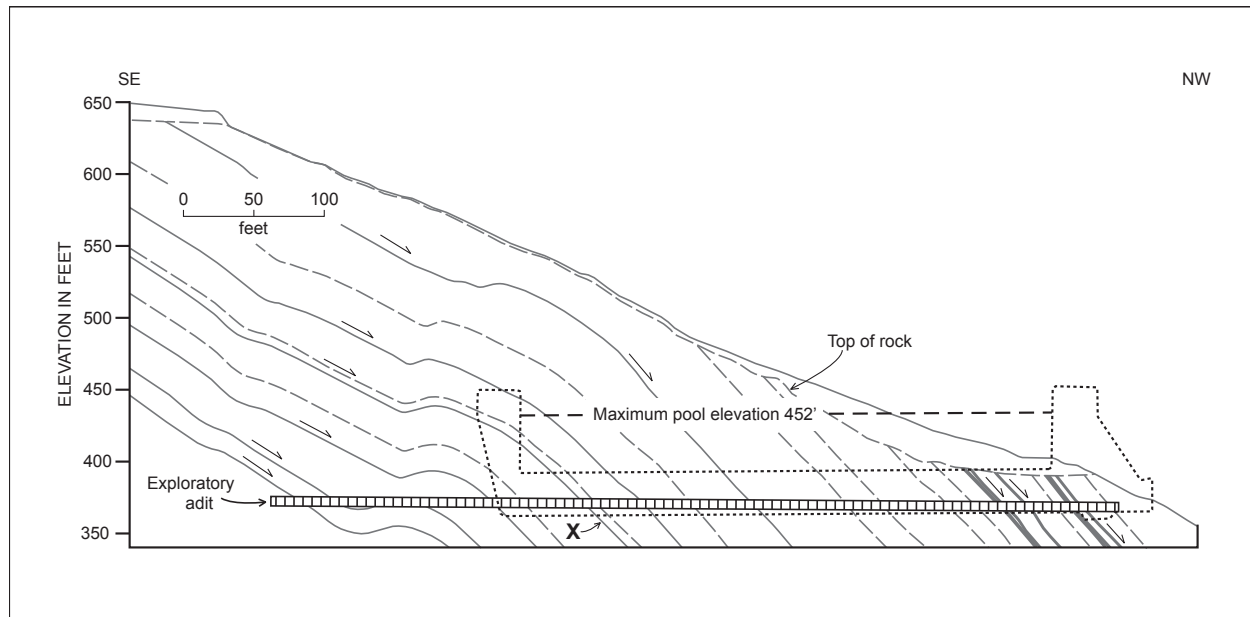


Figure 3. Cross section at site of spillway showing location of spillway (dotted line) and exploratory adit (from Epstein, 2001, as modified from Depman and Parrillo, 1969). Solid heavy lines are bedding-plane faults. Arrows show direction of movement. Dashed lines are normal bedding surfaces. X is lower of the two weathered bedding-plane faults discussed in the text.

entire mass would move downhill if the toe of the fault were daylighted (Dan Parrillo, U.S. Army Corps of Engineers, personal communication to Epstein, 1970, reported in Epstein 2001).

Further investigation was planned in a specially designed quarry to be dug in the early stages of reservoir construction. The slopes above the intakes and spillway were to be designed in a safe manner based on the results of findings at the quarry.

Other major changes to the dam not directly related to geologic conditions were as follows:

Increase in Dam Height.

As a result of a severe drought, which lasted for several years through the mid-1960s, safe yield estimates for the basin were revised downward. To compensate for the decreased safe yield, the height of the Tocks Island pool level was increased to the maximum practical height of levees at Port Jervis. This was to store water for low-flow maintenance and to repel salt water intrusion in the Delaware estuary. The proposed dikes caused considerable anger in the Port Jervis area.

Pumped Storage

A pumped storage facility built in the early 1960s at Yards Creek, close by the Tocks Island reservoir, uses two reservoirs, upper and lower, connected by pipelines and turbines. Water is pumped from the lower reservoir on Yards Creek, east of Kittatinny Mountain to an upper reservoir atop the mountain during off-peak hours (nights and weekends, for example). The water is then dropped to the lower reservoir for power generation during peak need periods. About 70 percent of the energy used in pumping is recovered, but the difference in value between peak power and off-peak power more than makes up for this loss (Smith, 1969).

Kittatinny Mountain was first considered for pumped storage in 1947, but the idea was not feasible until large, reversing turbines were developed in the 1950s. Interest in pumped storage atop Kittatinny Mountain revived in 1956. Not by coincidence, this was the year in which the Corps of Engineers released their report favoring a dam at Tocks Island. With the Tocks Island reservoir to the west of Kittatinny Ridge and Yards Creek to the east, pumped storage appeared to be an ideal way to transport water across Kittatinny Ridge at a profit to supply northern New Jersey. In an initial design, Delaware River water was to be pumped in off-peak hours to a series of three reservoirs atop the mountain and, for peak need generation, dropped to the Yards Creek reservoir in the Paulins Kill watershed. The water was then to be transferred as needed to the Round Valley Reservoir.

The existing portion of the system did not depend on the Tocks Island Dam and was completed in 1965. Numerous issues, particularly involving funding, ownership, and revenue distribution were unresolved at the time of the 1962 Tocks Island authorization, and pumped storage was not included the Congressionally authorized project. Over the next several years, these and other issues (particularly preservation issues at Sunfish Pond) were addressed, and, in a 1970 re-authorization, pumped storage was included as a purpose of the dam. Generators were eliminated from the dam plans, and conventional power generation was de-authorize.

FAILURE OF THE TOCKS ISLAND PROJECT

From the time of its authorization in 1962, the Tocks Island project experienced financial difficulties. Many claimed that the original cost estimates had been understated to assure approval. From a 1962 estimate of \$90 million, costs rose to \$400 million by 1975, and were still climbing. While soaring costs and budget shortfalls were clearly important in the demise of the project, their main effect may have been to cause delays which allowed time for other questionable aspects to come forward. Aspects which presented the greatest concern as the project moved forward were the cost benefit ratio, eutrophication, questionable recreation benefits, infrastructure needs which had not been addresses in the original proposal, and a near absence of evaluation of alternatives in the original proposal. Perhaps equally important in leading to strong grass roots opposition to the project were: 1) seemingly unending mismanagement and public relations disasters in land acquisition and the management of acquired land and 2) changing public attitudes towards large dams in general and Corps of Engineers in particular through this stage in the development of the environmental movement.

Although no specific issue stands out as the one that killed the Tocks Island Dam, and although the failure of the project was seen by many water resource managers as a prelude to continuing public supply shortages and low-flow related water quality issues in streams, the project had become politically untenable before the groundbreaking. The governors of New Jersey, New York, and Delaware rejected its continuation in 1975. Only the governor of Pennsylvania, citing water supply needs, continued to support the project. The project was formally deauthorized in 1992.

REFERENCES CITED

- Albert, Richard C., 1987, *Damming the Delaware*, Pennsylvania State Univ. Press, University Park, Pennsylvania, 202 p.
- Bell, J.C., 1910, Official opinion of the [Pennsylvania] Attorney General, p. 3.
- Dale, Frank T., 1996, *Delaware diary, episodes in the life of a river*: Rutgers Univ. Press, 203 p.
- Depman, A.J., and Parrillo, D.G., 1969, Geology of Tocks Island area and its engineering significance: *in* Subitsky, S., ed., *Geology of selected areas in New Jersey and eastern Pennsylvania*, Rutgers University Press, New Brunswick, NJ, p. 354-362.
- Epstein, J.B., 2001, Geologic controls of landslides in the Delaware Water Gap National Recreation Area, New Jersey-Pennsylvania, and Lehigh Gap, Pennsylvania: *in* Inners, J.D., and Fleeger, G.M., eds., 2001 – *a Delaware River odyssey*, Guidebook, 66th Annual Field Conference of Pennsylvania Geologists, Shawnee-on-Delaware, PA, p. 119-135.
- Mark, Eric, and Pierce, David, 2001, The legacy of Tocks Island Dam, A three-part retrospective: *Pocono Record*, August 12, 13, 14, 2001 [10 articles].
- Smith, Bennett L., 1969, Engineering geology of the Yards Creek hydro-electric pumped storage project: *in* Subitsky, S., ed., *Geology of selected areas in New Jersey and eastern Pennsylvania*, Rutgers Univ.

- Press, New Brunswick, NJ, p. 348-353.
- U.S. Army Corps of Engineers, 1965-1972, Tocks Island Lake, Pennsylvania, New Jersey, New York:
Design Memorandums 1 – 10, variously paginated.
- U.S. Congress, 1934, House Document 522 (Delaware River 308 Report).
- Vermeule, Cornelius C., 1894, Report on water-supply, water-power, the flow of streams, and attendant phenomena: Final Report of the State Geologist, v. 3, 352 p., New Jersey Geological Survey, Trenton, NJ.
- Witte, Ron W., 2001, Sand Hill Delta: Wisconsin glacial deposits in the Echo Lake lowland and manner of deglaciation: *in* Inners, J.D., and Fleeger, G.M., eds., 2001 – a Delaware River odyssey, Guidebook, 66th Annual Field Conference of Pennsylvania Geologists, Shawnee-on-Delaware, PA, p. 244 – 250.

Mesoproterozoic Rocks of the New Jersey Highlands and Overview of Associated Ore Deposits

by

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ABSTRACT

The New Jersey Highlands preserve a diverse assemblage of Mesoproterozoic-age rocks that correlate in part to other Grenville terranes in eastern North America mainly north of the Blue Ridge. Voluminous calc-alkaline metaplutonic and metavolcanic rocks of the Losee Suite were formed in a continental-margin magmatic arc at 1.28 to 1.25 Ga, and they represent the southern continuation of magmatic arcs formed along eastern Laurentia prior to 1.2 Ga.

Spatially associated with the Losee Suite is a widespread succession of supracrustal metasedimentary gneisses, locally stromatolitic marble, and cogenetic rhyolitic gneiss and locally pillowed mafic metavolcanic rocks that were deposited at 1.29 to 1.25 Ga in a back-arc basin inboard of the Losee arc. Location of the back-arc basin inboard (west by present coordinates) of the Losee arc implies a northwest-dipping subduction zone beneath the eastern Laurentian margin.

Emplacement of large volumes of A-type granite of the Byram and Lake Hopatcong Intrusive Suites at ca. 1.18 Ga followed the termination of arc and back-arc magmatism. These suites correlate in part to anorthosite, mangerite, charnockite, granite (AMCG) magmatism in the Adirondack Mountains and Grenville terranes in southeastern Canada.

Rocks in the New Jersey Highlands older than 1.04 Ga were metamorphosed to granulite-facies conditions during the Grenvillian (Ottawan) orogeny. The postorogenic, undeformed Mount Eve Granite, dated at 1.02 Ga, fixes the close of this tectonothermal event in the Highlands. Undeformed, discordant pegmatites and small granite bodies were emplaced between 1004 and 986 Ma, coeval with post-metamorphic hydrothermal activity associated with local U-Th-rare earth element (REE) mineralization.

INTRODUCTION

Mesoproterozoic rocks of the New Jersey Highlands have been a subject of study for more than a century due largely to past economic interest in the numerous low-Ti magnetite mines that are distributed throughout the region and the marble-hosted Zn-Fe-Mn deposits at the Franklin and Sterling Hill mines. The four-unit subdivision of Mesoproterozoic rocks developed by early workers (e.g., Spencer et al., 1908) was expanded by geologists from the New Jersey Zinc Company (Hague et al., 1956) and the U.S. Geological Survey (Hotz, 1952; Sims, 1958; Offield, 1967; Drake, 1969;

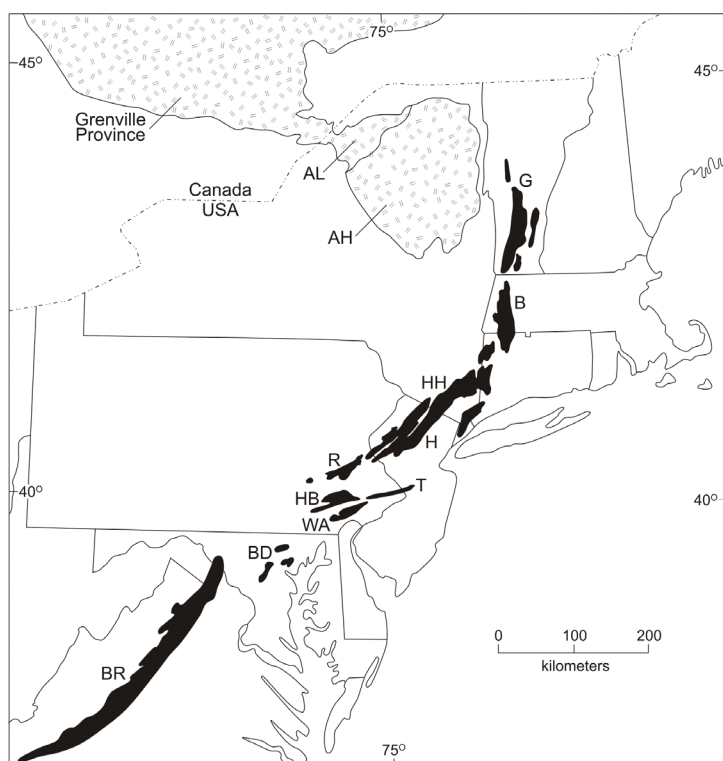


Figure 1. Grenville-age rocks in the Appalachians (solid), Adirondack Highlands (AH) and Lowlands (AL), and the Grenville Province of southeastern Canada (patterned). BR, Blue Ridge; BD, Baltimore Gneiss domes; WA, West Chester and Avondale Massifs; HB, Honey Brook Upland; T, Trenton Prong; R, Reading Prong; H, New Jersey Highlands; HH, Hudson Highlands; B, Berkshire Mountains; and G, Green Mountains. Modified from Rankin et al. (1993).

Baker and Buddington, 1970).

From 1984 until 1990, reconnaissance-scale mapping of the Highlands was undertaken jointly by the New Jersey Geological Survey and the U.S. Geological Survey, leading to publication of a 1:100,000-scale bedrock geologic map of New Jersey (Drake et al., 1996). From 1991 until the present, the New Jersey Geological Survey has conducted detailed 1:24,000-scale bedrock mapping of the 32 7.5-minute quadrangles that make up the Highlands. This work has built significantly upon previous studies and regional syntheses (e.g., Rankin et al., 1993), permitting a comprehensive comparison of the Mesoproterozoic rocks of the western and eastern Highlands that are separated by Paleozoic-age cover rocks. These results, in combination with recent U-Pb zircon geochronology (Volkert et al., 2005; Aleinikoff et al., 2007; Volkert et al., in press), provide important constraints on the timing of major tectonic, magmatic and metamorphic events, and they permit a comparison of the Mesoproterozoic rocks of the Highlands and other eastern North American Grenvillian terranes. The latter has proven critical in understanding the geologic evolution of Mesoproterozoic rocks in the north-central Appalachians within the context of the broader Grenville Orogenic Cycle.

The New Jersey Highlands constitute one of the numerous Appalachian Grenvillian inliers that extend along eastern North America (Fig. 1). Rocks of Mesoproterozoic age underlie most of the approximately 1,000 km² area of the Highlands region that is separated into western and eastern parts by downfaulted, Paleozoic-age rocks of the Green Pond Mountain region (Fig. 2). Mesoproterozoic rocks in the western Highlands are nonconformably overlain by the Chestnut Hill Formation, a locally preserved sequence of Neoproterozoic siliciclastic and felsic volcanic rocks (Drake, 1984; Gates and Volkert, 2004). Throughout the Highlands, Mesoproterozoic rocks are intruded by a northeast-trending swarm of Neoproterozoic alkalic to tholeiitic diabase dikes emplaced during breakup of the supercontinent Rodinia (Volkert and Puffer, 1995). The Early Cambrian Hardyston Formation rests nonconformably on Mesoproterozoic rocks and on the Neoproterozoic rocks, where present.

Along the eastern border of the Highlands, Mesoproterozoic rocks are in fault contact with Late Triassic to Early Jurassic terrestrial fluvial and lacustrine sedimentary rocks and Early Jurassic tholeiitic basalts of the Piedmont (Newark basin), and locally with sedimentary rocks of Cambrian and Ordovician age (Fig. 2). Along the western border, Mesoproterozoic rocks are nonconformably overlain by, or in fault contact with, a passive margin clastic and carbonate shelf sequence of Cambrian and Ordovician age.

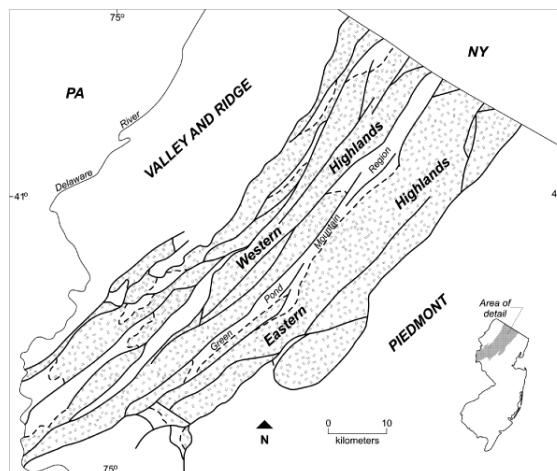


Figure 2. Simplified geologic map of Mesoproterozoic rocks (patterned) in the western and eastern New Jersey Highlands. Unpatterned areas within the Highlands are underlain by Paleozoic rocks. Solid lines are faults; dashed lines are unconformities.

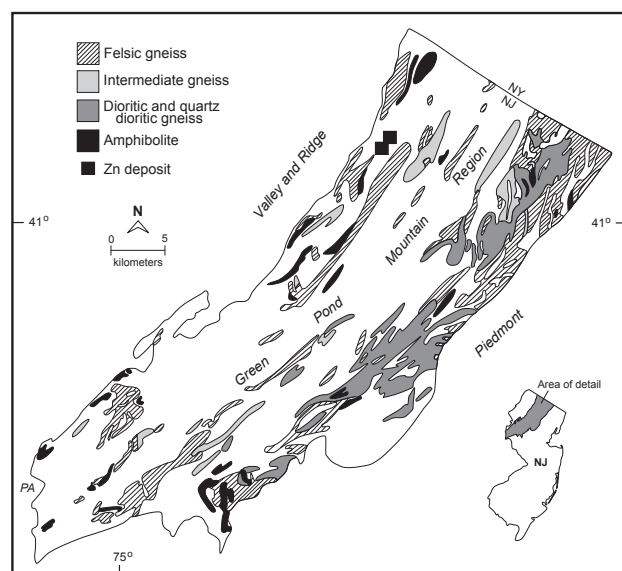


Figure 3. Simplified geologic map of metavolcanic and metaplutonic rocks of the Losee Suite. Modified from Drake et al. (1996).

MAGMATIC-ARC ROCKS

The Losee Suite (Drake, 1984; Volkert and Drake, 1999) in the New Jersey Highlands contains voluminous calc-alkaline and tholeiitic, metaplutonic and metavolcanic, arc-related rocks (Puffer and Volkert, 1991; Volkert, 2004) that represent the southern continuation of magmatic arcs formed along the eastern margin of Laurentia prior to 1.2 Ga. Similar arc complexes that are mineralogically and geochemically similar to the Losee Suite are

widespread in Grenvillian terranes to the north, and these include the Green Mountains (Ratcliffe et al., 1991), Adirondack Highlands (McLelland and Chiarenzelli, 1990), and Central Metasedimentary Belt in southeastern Canada (e.g., Corfu and Easton, 1995). U-Pb zircon and monazite sensitive high-resolution ion microprobe (SHRIMP) geochronology of representative rocks of the Losee Suite yielded ages of 1.28 to 1.25 Ga (Aleinikoff et al., 2007; Volkert et al., in press).

Rocks of the Losee Suite are nearly equal in abundance throughout the Highlands (Fig. 3) where they underlie about 30% of the region. Based on major- and trace-element geochemical compositions they are subdivided into felsic rocks which include: dacitic gneiss, low-K rhyolitic gneiss, and tonalitic gneiss; rocks of intermediate composition which locally contain orthopyroxene and include dacitic gneiss, andesitic gneiss, and dioritic to quartz dioritic gneiss; and mafic rocks mapped as amphibolite (Volkert, 2004).

Felsic and Intermediate Rocks

Felsic rocks weather white, are light greenish gray on fresh surfaces, and are composed of quartz and plagioclase (oligoclase to andesine), variable amounts of amphibole, biotite, clinopyroxene, and Fe-Ti oxides. Garnet and rutile are locally present. Felsic rocks have 64-77 wt. % SiO_2 , are divisible into high-Al (>15 wt. %) and low-Al (<15 wt. %) types, and are predominantly dacitic or tonalitic in composition (Puffer and Volkert, 1991; Volkert and Drake, 1999; Blake, 2003).

Intermediate rocks weather light gray or tan, are greenish gray to brownish gray on fresh surfaces, and display greasy luster. They are composed of plagioclase (oligoclase to andesine), clinopyroxene, amphibole, orthopyroxene, and variable amounts of quartz, biotite, and Fe-Ti oxides. Quartz-rich, intermediate, metavolcanic gneisses have 60-73 wt. % SiO_2 and 15-17 wt. % Al_2O_3 (Volkert and Drake, 1999; Volkert, 2004). Based on their geochemical compositions, dacitic gneisses of intermediate composition are geochemically similar to felsic rocks of the Losee Suite, whereas andesitic gneisses reflect a slightly more mafic composition (Volkert and Drake, 1999). Intermediate, dominantly quartz-poor metaplutonic rocks are dioritic gneiss that contain 51-58 wt. % SiO_2 and 15-21 wt. % Al_2O_3 . Compared to quartz-rich intermediate gneisses, dioritic gneiss has higher Al_2O_3 , FeO, MgO, CaO and Cr (Volkert and Drake, 1999).

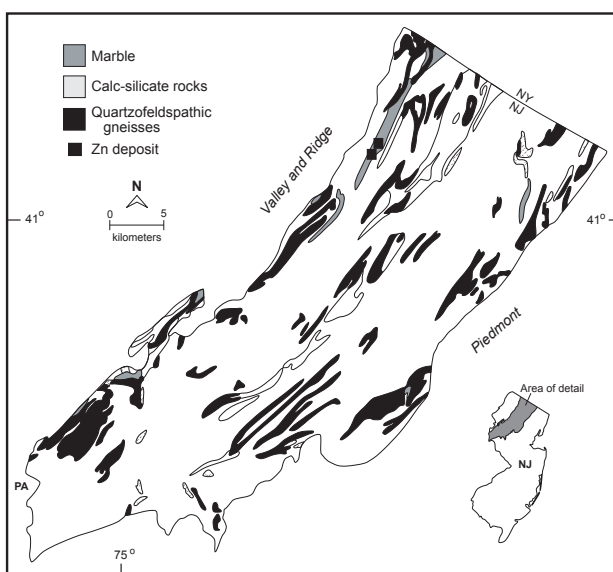


Figure 4. Simplified geologic map of metasedimentary gneisses and marble. Modified from Drake et al. (1996).

Mafic Rocks

Medium-grained, well-foliated amphibolites that contain hornblende + plagioclase + augite \pm hypersthene \pm biotite are commonly intercalated with felsic and intermediate variants of the Losee Suite, although they are volumetrically a relatively minor component. Amphibolites of the Losee Suite have predominantly calc-alkaline compositions indicative of basalt to basaltic andesite protoliths containing 49-56 wt. % SiO_2 (Maxey, 1971; Volkert and Drake, 1999; Kula and Gorrington, 2000). They plot mainly in the field of island-arc tholeiites or destructive margin basalts on tectonomagmatic diagrams of TiO_2 -MnO- P_2O_5 and Hf/3-Th-Ta and Mg# versus TiO_2 (Volkert, 2004).

BACK-ARC SUPRACRUSTAL ROCKS

Metamorphosed supracrustal rocks in the Highlands include quartzofeldspathic gneisses, calc-silicate gneisses, quartzite, marble, and metavolcanic rocks that underlie about 20% of the region. Ultramafic rocks are absent in the succession. Supracrustal rocks predominate in the western part of the region where they reach an estimated thickness of ~2,100 m (Hague et al., 1956) (Fig. 5). U-Pb SHRIMP geochronology of supracrustal felsic metavolcanic rocks yielded ages of 1.29 to 1.25 Ga (Aleinikoff et al., 2007; Volkert et al.,

in press) that overlap the age of the Losee Suite. Thus, these ages provide bounding limits for the age of the supracrustal succession, as well as an important constraint for the timing of zinc mineralization in the carbonate protolith of the Franklin Marble.

Metasedimentary rocks

Quartzofeldspathic rocks occur nearly equally in both areas of the Highlands (Fig. 4). These rocks include a range of gneisses whose protoliths were predominantly sedimentary as indicated by their heterogeneous textures, intercalation with quartzites, and their geochemical characteristics, particularly their siliceous compositions and variable Zr/Nb and Zr/Y ratios (Volkert and Drake, 1999). Quartzofeldspathic gneisses are typically pinkish white to pinkish gray, locally rusty, and range in texture from well layered to massive. They are composed primarily of quartz, K-feldspar, and oligoclase. Ferromagnesian minerals, where present, include biotite, amphibole, and (or) clinopyroxene. Garnet, sillimanite, zircon, apatite, and Fe-Ti oxides occur as common accessory phases. Graphite and sulfides are confined mainly to rusty biotite-quartz-feldspar gneiss. Some metasedimentary gneisses locally preserve relic graded beds, thin layers of pebble conglomerate with deformed and flattened quartz pebbles, and trough cross-laminated beds despite the effects of high-grade metamorphism and recrystallization.

Quartzite is sparsely exposed throughout the Highlands where it occurs as lenses and layers intercalated with quartzofeldspathic gneisses, calc-silicate gneisses, and marble. Quartzites can locally be traced along strike for distances of more than 1 km. Quartzite is characteristically light gray, vitreous, medium grained, and well layered to massive. Where associated with quartzofeldspathic gneisses it commonly contains biotite, garnet, graphite, and locally abundant tourmaline, and where associated with calc-silicate gneisses it contains mainly diopside. Protoliths for most quartzite are interpreted as impure quartzose sandstone, although some thin discontinuous lenses and layers in marble and calc-silicate rocks may have originally been chert. Quartzites lack textural evidence suggestive of an origin by metamorphic segregation or anatexis, or as hydrothermal deposits.

Calc-silicate rocks consist of diopside-bearing gneisses that encompass a range of compositions whose protoliths likely were calcareous sandstone, quartzose and argillaceous carbonate rocks, cherty dolomite, and carbonate-bearing volcanoclastic sandstone (Volkert and Drake, 1999). Calc-silicate rocks are light gray, green, or greenish gray, locally rusty, and range texturally from well layered to massive. Ferromagnesian minerals consist of diopside \pm amphibole and biotite. Quartz, oligoclase, K-feldspar, carbonate, epidote, titanite, scapolite, sulfides, and Fe-Ti oxides occur in varying proportions in different rock types.

Marble

Marble is a volumetrically minor rock that underlies about 5% of the Highlands and is most abundant in the western part of the region (Fig. 4). Marbles reflect carbonate deposition in a marine environment, as suggested by the carbon isotope data (Johnson et al., 1990; Peck et al., 2006) and the occurrence of stromatolites (Volkert, 2004). Marbles are white to light gray, fine to coarse crystalline, well layered to massive, and mainly calcitic to very locally dolomitic. Common accessory minerals include graphite, phlogopite, and variable amounts of chondrodite, calcic amphibole and clinopyroxene.

New Jersey Zinc Company geologists (Hague et al., 1956) subdivided marble in the northwestern Highlands into the 335- to 457-m-thick lower Franklin Marble layer, which hosts the Zn deposits at Franklin and Sterling Hill, and the approximately 91-m-thick upper Wildcat Marble layer (Fig. 5).

Marble in the eastern Highlands (Fig. 4) occurs as disconnected layers and small bodies typically

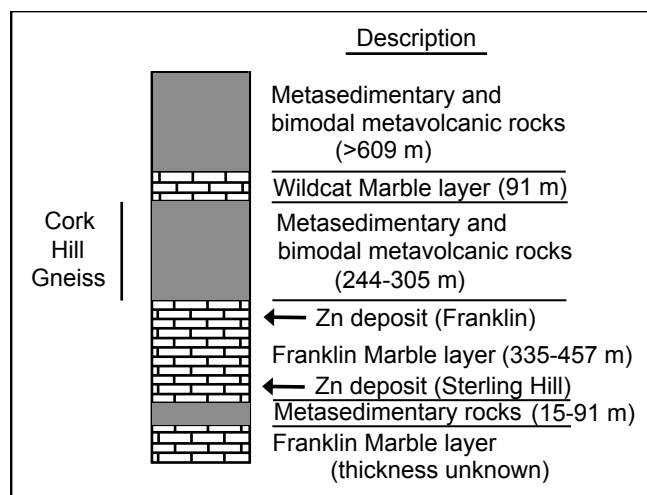


Figure 5. Generalized stratigraphic column of supracrustal rocks from the western New Jersey Highlands. Modified from Hague et al. (1956) and Volkert (2004).

less than 30 m thick that grade along strike into a thick succession of metasedimentary gneisses that are interpreted to reflect deposition of both protoliths in a deeper water marine environment (Volkert et al., 2000a; Johnson and Skinner, 2003; Volkert, 2004).

Metavolcanic rocks

Supracrustal felsic metavolcanic rocks of rhyolitic composition are pinkish white to pinkish gray, foliated, and composed primarily of quartz, K-feldspar, oligoclase, biotite, garnet, zircon, apatite, and Fe-Ti oxides. They are thickest in the northwest Highlands where they form part of a cogenetic bimodal volcanic sequence with mafic metavolcanic rocks mapped as amphibolite. Felsic metavolcanic rocks occur at several different stratigraphic levels and are conformably intercalated with marble (Fig. 5) and with layered gneisses whose variations in texture and modal composition, particularly their locally high contents of quartz and biotite, suggest a sedimentary protolith (Hague et al., 1956; Volkert and Drake, 1999). Felsic metavolcanic rocks have 67-76 wt % SiO_2 , are peraluminous (Al saturation index = 1.05-1.34), and have moderately high Li concentrations (14- 40 ppm) that contrast with younger metaluminous granites of the Byram and Lake Hopatcong Suites.

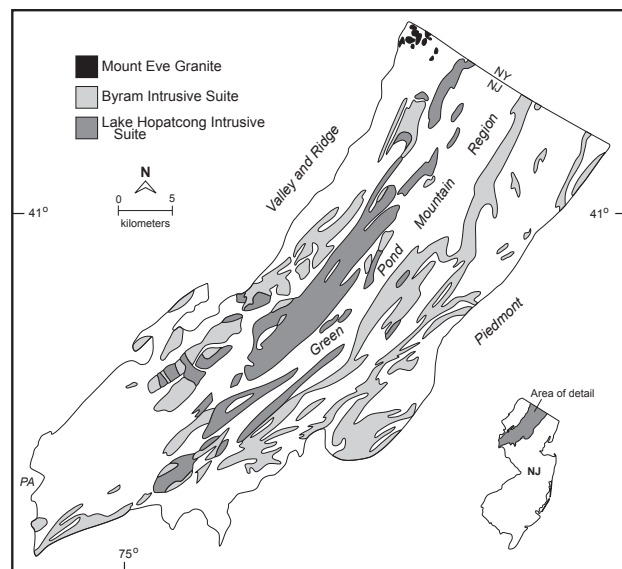


Figure 6. Simplified geologic map of granitic rocks of the Byram and Lake Hopatcong Intrusive Suites (Vernon Supersuite) and the Mount Eve Granite. Modified from Drake et al. (1996).

These characteristics, along with lower Zr/Y ratios, also help to distinguish supracrustal rhyolitic gneiss from low-K rhyolitic gneiss of the Losee Suite. Mafic metavolcanic rocks mapped as amphibolite are medium-grained, foliated gneiss composed of hornblende + plagioclase \pm augite \pm biotite \pm Fe-Ti oxides. They occur at several different stratigraphic intervals in both areas of the Highlands. Amphibolites locally contain pillow structures (Hague et al., 1956; Volkert, 2004), and they have geochemical compositions similar to metabasalt, indicating that protoliths were most likely mafic volcanics (Maxey, 1971; Volkert and Drake, 1999). However, very locally some amphibolites interlayered with calc-silicate gneiss of known sedimentary parentage commonly weather rusty, are sulfidic, and may represent metamorphosed shale (Volkert and Drake, 1999). Most supracrustal mafic metavolcanic rocks are tholeiitic and have geochemical compositions similar to mid-ocean ridge basalt (MORB), although some mafic metavolcanic rocks from the eastern

Highlands are calc-alkaline and have characteristics that are transitional between MORB and island-arc basalts.

GRANITE SUITES

Voluminous granite and related rocks of the Byram and Lake Hopatcong Intrusive Suites, postorogenic Mount Eve Granite, and other postorogenic felsic intrusive rocks underlie approximately 50% of the Highlands (Fig. 6). The Byram and Lake Hopatcong suites, which together constitute the Vernon Supersuite (Volkert and Drake, 1998), form linear belts tens of km in length. Rocks of both suites display similar fold geometries, conformable contacts, and locally display evidence of magma commingling (Volkert, 1995). Contacts between the granite suites and other Mesoproterozoic rocks generally are conformable. However, near their margins the granites very locally contain xenoliths of other Mesoproterozoic rocks.

Byram and Lake Hopatcong Intrusive Suites

The Byram Intrusive Suite includes monzonite, quartz monzonite, granite, alaskite, and pegmatite, a lithologic assemblage that is interpreted to represent a magmatic differentiation series (Volkert, 1995). These rocks are foliated and contain hastingsite \pm biotite (annite) as determined by microprobe analysis (Volkert et al., 2000b). The Lake Hopatcong Intrusive Suite includes monzonite, quartz monzonite, granite, and alaskite that are also interpreted to represent a product of magmatic differentiation (Volkert, 1995). Pegmatitic

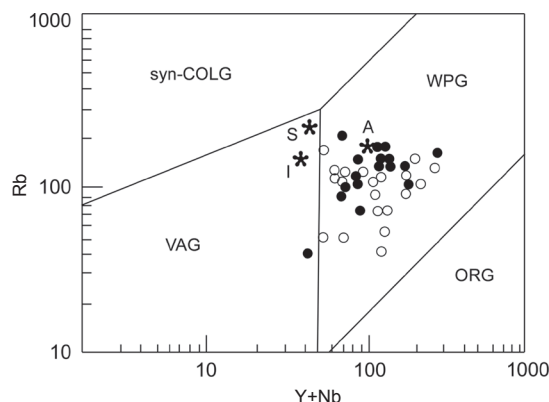


Figure 7. Plot of Byram (filled circles) and Lake Hopatcong (open circles) granites (Volkert et al., 2000b) on the Rb vs. Y + Nb diagram (Pearce et al., 1984). Granite fields are: syn-collision (syn-COLG); volcanic arc (VAG); within-plate (WPG); and ocean ridge (ORG). Average composition of A, I, and S-type granites from Whalen et al. (1987).

variants are rare compared to the Byram Suite, likely reflecting the relatively anhydrous composition of Lake Hopatcong magma. Lake Hopatcong rocks are foliated and contain hedenbergite, although fayalitic olivine and sparse ferrosilite and Fe-pigeonite locally coexist with hedenbergite in some rocks (Young and Cuthbertson, 1994).

Byram and Lake Hopatcong rocks are metaluminous to weakly peraluminous (Al saturation index = 0.7 to 1.1) and have similar concentrations of major and most trace elements. Both suites have near continuous ranges of SiO_2 from 58 to 75 wt. % and distinctive chemical compositions that are characterized by high total alkalis ($\text{K}_2\text{O} + \text{Na}_2\text{O} = 7$ to 11 wt. %), Fe, F, Cl, Nb, Zr, Y, Ga, light REE, and low Mg, Ca, and Sr. They form an overlapping and continuous differentiation trend from monzonite through granite on Harker variation diagrams and an AFM diagram (Volkert et al., 2000b). Both suites plot exclusively within the A-type granite field

on major- and trace-element discrimination diagrams (Volkert, 1995) and have similar Ga/Al ratios of 4 to 6 that are characteristic of A-type granites (Whalen et al., 1987). Both suites plot in the field of within-plate granites on tectonomagmatic discrimination diagrams (Fig. 7).

Intrusion of A-type granite and related rocks of the Byram and Lake Hopatcong Intrusive Suites at 1.18 Ga (Volkert et al., in press) followed the termination of calc-alkaline arc-related magmatism in the Highlands. Emplacement of these granite suites was synchronous with the older parts of AMCG suites in the Adirondack Mountains and southeastern Canada, documenting a similar style of magmatism resulting from collision-related delamination of lithospheric mantle.

Postorogenic Intrusive Rocks

Postorogenic rocks include the Mount Eve Granite, widespread thin sheets of weakly foliated, mildly discordant, microperthite alaskite, undeformed pegmatites, and locally occurring, thin, discordant, undeformed felsic intrusive bodies. The postorogenic Mount Eve Granite in northwestern New Jersey and adjacent New York (Fig. 6) occurs as nearly thirty, mostly small bodies. Mount Eve Granite bodies lack metamorphic foliation but display a weak magmatic fabric (Drake et al., 1991), exhibit discordant contacts with adjacent Mesoproterozoic rocks, contain sparse xenoliths of country rock, and are responsible for production of local thermal aureoles, some of which contain wollastonite, where intrusive into Franklin Marble (Volkert and Drake, 1999). The 1.02 Ga age of Mount Eve Granite (Drake et al., 1991) is consistent with the field relationships that indicate postorogenic emplacement. The Mount Eve Granite is metaluminous to weakly peraluminous syenogranite (Drake et al., 1991) with 65-72 wt. % SiO_2 , and A-type geochemical characteristics (Gorring et al., 2004). Overall, Mount Eve Granite is mineralogically and geochemically similar to granite of the Byram Intrusive Suite.

Other postorogenic granites occur as small, irregular-shaped felsic intrusions that lack metamorphic fabric, display discordant contacts with, and contain xenoliths of, foliated country rock, consistent with postorogenic emplacement. A U-Pb zircon age of 1004 Ma was obtained from an alaskite body that intrudes Franklin Marble (Volkert et al., 2005). The youngest postorogenic rocks are widespread granite pegmatites that have strongly discordant contacts with foliated country rock. A U-Pb zircon age of 986 Ma was obtained from a pegmatite that intrudes Franklin Marble (Volkert et al., 2005). These postorogenic rocks indicate that magmatic activity continued well past the peak of granulite-facies metamorphism. Postorogenic intrusions and pegmatites are felsic, contain 75-79 wt. % SiO_2 , and are geochemically similar to A-type granite (Volkert et al., 2005). The pegmatites display geochemical characteristics of the niobium-yttrium-fluorine (NYF) class of rare-earth pegmatites of Černý (1992), namely metaluminous to weakly peraluminous compositions, LREE enrichment, mineralization by U, Th, and REE, and postorogenic emplacement.

OTTAWAN OROGENY

Calc-alkaline magmatism in the New Jersey Highlands ceased by ca.1.25 Ga, but evidence for an accretionary suture or tectonic boundary reflecting a collisional event correlative to the Elzevirian orogeny has not been recognized (Volkert, 2004). Metamorphic overgrowths on zircon and monazite of 1.2 Ga are also unknown in the Highlands, providing little support for a high-grade tectonothermal event at this time in the north-central Appalachians.

Mesoproterozoic rocks of the New Jersey Highlands were metamorphosed to granulite-facies conditions during the Ottawa orogeny (Volkert, 2004 and references therein; Peck et al., 2006). The pervasive northeast-trending, southeast-dipping metamorphic fabric and dominant northwest-verging isoclinal folds in the Mesoproterozoic rocks were acquired during this tectonothermal event. Timing of Ottawa metamorphism is constrained by overgrowths on zircon and monazite that yield ages between 1.04 and 1.02 Ga (Volkert et al., in press). The undeformed, postorogenic 1.02 Ga Mount Eve Granite places a temporal constraint on the close of Ottawa orogenesis in the New Jersey Highlands and contiguous areas. Pressure and temperature estimates during regional Ottawa metamorphism are in reasonably good agreement from throughout the New Jersey Highlands and yield temperatures of 670°-780° C and pressures of 410-620 MPa (Volkert, 2004; Peck et al., 2006). Prograde mineral assemblages of metasedimentary gneisses are the same in both parts of the Highlands, and orthopyroxene is present in mafic and intermediate orthogneisses, indicating similar peak P-T conditions representative of granulite facies (Volkert, 2004). The available geochronological and geochemical data do not support a metamorphic discontinuity between the two parts of the Highlands related to the Ottawa collisional event, and faults partitioning the region into structural blocks do not appear to be tectonic boundaries separating disparate terranes. Therefore, the metamorphic pressure and temperature data are consistent with the lithologic, geochemical, and isotopic commonality of the western and eastern Highlands, and suggest that the entire region represents a single, although tectonically shortened, terrane that experienced a common geologic evolution.

MESOPROTEROZOIC ORE DEPOSITS

Iron

Perhaps the most important commodity mined in the New Jersey Highlands was low-Ti magnetite (\pm minor hematite). Although earliest mining predates the Revolutionary War, the industry did not reach its peak until the late 1800's. By the turn of the 20th century, more than 400 mines had been developed (Bayley, 1910) and New Jersey led the nation in the production of iron ore. This distinction lasted until the early 1900's when production from deposits in the Lake Superior region finally surpassed that of the Highlands. At the close of the industry in New Jersey in 1966, the estimated total production of iron ore had exceeded 50 million tons.

Magnetite deposits are both widespread and abundant in the Highlands (Fig. 8) where they are hosted by a variety of Mesoproterozoic rock types. The deposits have characteristic tabular to lenticular shapes that locally pinch and swell. While Fe in the deposits could arguably be due to post-metamorphic replacement, the geologic evidence strongly suggests that most of the Fe was present in the host rocks of the deposits prior to the metamorphic peak at ca. 1.04 Ga. This evidence includes the following: 1) all of the deposits are structurally conformable to the foliation in the host rocks; 2) the deposits plunge toward the northeast, parallel to the regional lineation; 3) the deposits have the same fold geometries as their host rocks; 4) field relationships provide little evidence for fault control on the distribution and occurrence of the deposits; 5) the ores appear to have undergone the same granulite facies metamorphism as the country rocks as determined by thermometric calculations of ore (Baker and Buddington, 1970; Puffer et al., 1993); and 6) reaction rims are absent between ore minerals and adjacent silicates (Baker and Buddington, 1970) suggesting equilibration at the same high metamorphic grade. At the present level of erosion few of the deposits are fault proximal but their linear trend does argue for some degree of structural control. A speculative, but reasonable explanation may involve the localization

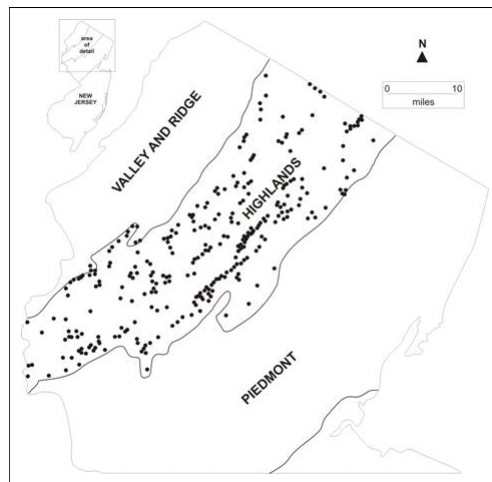


Figure 8. Distribution of some of the more than 400 low-Ti magnetite mines in the New Jersey Highlands.

of Fe along pre-metamorphic shear zones that were subsequently overprinted by recrystallization during granulite facies metamorphism. The host rocks and ores were deformed together during the end of the Ottawa orogeny and acquired their present structural conformity.

Exceptions to the above deposits are discordant magnetite-bearing pegmatites and thin, discordant veins and anastomosing networks of magnetite that clearly postdate the regional granulite facies metamorphism. Most of these occurrences are small and close enough to known magnetite deposits that the Fe may have been mobilized from the pre-existing orebodies. The age of this mineralization ranges from 998 to 940 Ma and overlaps the timing of emplacement of regional pegmatites that locally are characterized by minor concentrations of U and REE (Volkert et al., 2005). At the Scrub Oaks mine in the eastern Highlands, REE are hosted mainly in xenotime, synchysite (doverite), and bastnaesite within irregular-shaped, locally discordant bodies in the magnetite ore and in pegmatite adjacent to the ore (Klemic et al., 1959) implying a post-metamorphic age for this mineralization.

Zinc

Nearly as significant historically as the Mesoproterozoic Fe deposits, and certainly more widely studied and internationally renowned, are the Zn deposits at Franklin and Sterling Hill in the western Highlands (Fig. 9). Both mines have a long history, with the Franklin mine having been active from about 1770 until 1954 and the Sterling Hill mine from about 1772 until 1986. The economic importance of this industry in New

Jersey is underscored by the state's ranking as second in the nation in zinc ore production during the early 1900's. Approximately 34 million tons of zinc ore were mined from these two deposits between 1868 and the close of the industry in 1986. Unfortunately, production figures prior to 1868 are nonexistent. Excellent descriptions of these orebodies are found in Spencer (1908), Ries and Bowen (1922), Metsger et al. (1958), Frondel and Baum (1974), Dunn (1995), and Metsger (1997).

The marble-hosted deposits of zincite, willemite, and franklinite at Franklin and Sterling Hill are mineralogically similar. Both deposits occur within the Franklin Marble layer, and stratigraphically beneath ~300 m of interlayered

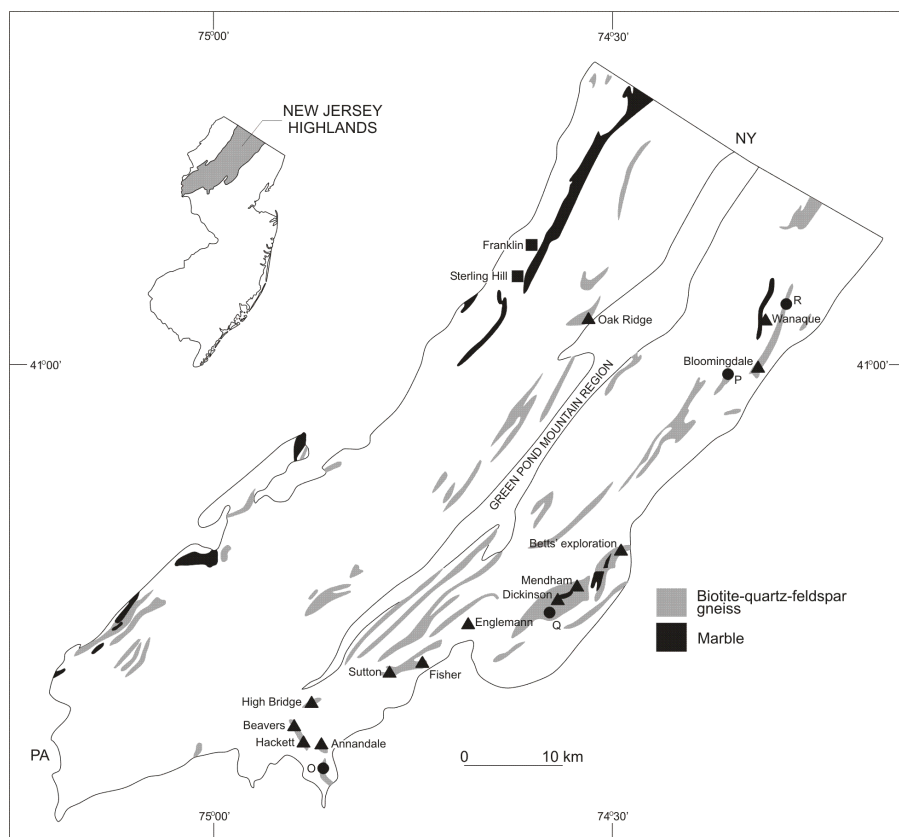


Figure 9. Distribution of marble in the New Jersey Highlands and locations of the Franklin and Sterling Hill zinc ore deposits in the western Highlands. Triangles locate graphite deposits in the eastern Highlands hosted by rusty biotite-quartz-feldspar gneiss and quartzite (Volkert et al., 2000a).

metasedimentary and metavolcanic gneisses informally named Cork Hill Gneiss (Fig. 5) by New Jersey Zinc Company geologists. Evidence that metals in the Zn deposit at Sterling Hill were in place prior to the granulite facies metamorphism is provided by the work of Metsger (1962) who recognized the conformity of the c-axis of willemite to the regional foliation, as well as by stable isotope studies of the orebodies (Johnson et al., 1990; Peck et al., 2009).

In addition to the Franklin and Sterling Hill deposits, a small marble-hosted deposit of sphalerite was worked about 1883 at the Raub mine in the southwestern Highlands. Zinc deposits at the Franklin and Raub mines are spatially associated with thin stratiform magnetite layers that occur in amphibolite structurally just beneath the marble contact. Manganese content of the magnetite is high at the Franklin and Raub mines (1.83 to 7.40 % Mn; Bayley, 1910), and comparable to that of other marble-hosted Fe deposits in the Highlands that lack Zn (0.40 to 11.28 % Mn), but considerably higher than the Mn contents of the non-marble-hosted Fe deposits (typically <0.60 %) (Bayley, 1910). The genetic significance of the association Fe-Mn and Zn in marble-hosted deposits in the Highlands has long been recognized (Spencer, 1908; Hague et al., 1956; Frondel and Baum, 1974) and has been addressed recently by several workers (e.g., Puffer, 1997; Johnson, 1997; Johnson and Skinner, 2003; Peck et al., 2009).

Graphite

Between 1848 and 1931, graphite was mined at 14 locations in the eastern Highlands (Fig. 9). Except for a single deposit just west of the Green Pond Mountain region these deposits all occur along the same trend. The deposits are continuous along strike for hundreds of meters and can be traced discontinuously for as much as several kilometers. At all locations the ore is hosted by rusty, sulfidic, biotite-quartz-feldspar gneiss and quartzite that are interlayered with diopside-bearing calc-silicate gneiss and marble. Graphite occurs in host rocks as disseminated flakes that make up from 17 to 40 volume percent of the rock in ore zones, as well as massive graphite in conformable layers. Ore zones range in width from <1 m to 9 m.

Protoliths of metasedimentary rocks in the New Jersey Highlands hosting the graphite deposits constitute a Mesoproterozoic-age marine sequence of locally organic-rich sand and mud. In a detailed study of the graphite deposits, Volkert et al. (2000a) obtained $\delta^{13}\text{C}$ values of -26 ± 2 per mil and -23 ± 4 per mil for graphite ore occurring within quartzite and gneiss, respectively, and $\delta^{34}\text{S}$ values of 0 to 9 per mil for associated pyrrhotite. They found no evidence for structural control of the positions of the ore zones, or for the large-scale synorogenic or postorogenic mobilization of carbon. Thus, the field, petrographic, and isotopic data indicate the graphite deposits formed in situ through the granulite facies metamorphism of Proterozoic accumulations of biogenic carbon, possibly in the form of algal mats (Volkert et al., 2000a).

CONCLUSIONS

The Losee Suite forms a regionally widespread calc-alkaline to tholeiitic assemblage of 1.28 to 1.25 Ga metaplutonic (dioritic gneiss and quartz dioritic gneiss, tonalitic gneiss) and metavolcanic (amphibolite, andesitic gneiss, dacitic gneiss, minor rhyolitic gneiss) rocks generated in a continental-margin magmatic arc along the eastern Laurentian margin. These calc-alkaline rocks are the southern continuation of magmatic arcs formed along the Laurentian margin prior to 1.2 Ga.

Synchronous sedimentation and bimodal volcanism characterized a back-arc basin that formed by extension of continental crust along the Laurentian margin at 1.29 to 1.25 Ga. The ages of these rocks constrain the timing of Zn-Fe-Mn metal input in the carbonate protolith of the Franklin Marble. Back-arc felsic volcanic rocks are interpreted as formed in an extensional, within-plate tectonic setting through partial melting of a crustal source inboard of the Losee arc.

Emplacement of A-type granite and related rocks of the Byram and Lake Hopatcong Intrusive Suites at ca. 1.18 Ga followed the termination of calc-alkaline, arc-related magmatism in the Highlands. Emplacement of the Byram and Lake Hopatcong suites was synchronous with the older parts of AMCG suites in the Adirondack Mountains and southeastern Canada, documenting a similar style of magmatism resulting from collision-related delamination of lithospheric mantle.

Mesoproterozoic rocks throughout the Highlands were metamorphosed to granulite-facies conditions during the Ottawa orogeny at ca. 1.04 Ga, as recorded in metamorphic overgrowths on zircon and monazite. Zn-Fe-Mn metals were metamorphosed to granulite-facies assemblages at this time, along with the enclosing carbonate protolith of the Franklin Marble.

Following the Ottawa orogeny, A-type granite magmatism continued sporadically in the Highlands. It is represented in the waning stages by the undeformed 1.02 Ga Mount Eve Granite, which provides a temporal constraint for the close of the Ottawa tectonothermal event in the north-central Appalachians, and widespread, small volumes of felsic magmatism at 1004 to 986 Ma. The latter is coeval with formation of small, subeconomic deposits of low-Ti magnetite, U-Th-REE mineralization, and hydrothermal activity in the Highlands.

Mesoproterozoic rocks of the Highlands correlate temporally to other Grenvillian terranes in the northern Appalachians and in southeastern Canada. Available data also indicate that the western and eastern New Jersey Highlands share a common magmatic, tectonic, sedimentary, and metamorphic history and that they represent a single, tectonically shortened terrane.

REFERENCES

- Aleinikoff, J.N., Volkert, R.A., and Fanning, M.C., 2007, SHRIMP U-Pb geochronology of zircon and monazite from ca.1.3 Ga arc-related rocks, New Jersey Highlands: Geological Society of America Abstracts with Programs, v. 39, No. 1, p. 37.
- Baker, D.R., and Buddington, A.F., 1970, Geology and magnetite deposits of the Franklin quadrangle and part of the Hamburg quadrangle, New Jersey: U.S. Geological Survey Professional Paper 638, 73 p.
- Bayley, W.S., 1910, Iron mines and mining in New Jersey. Final Report of the State Geologist, VII, p. 220-223.
- Blake, C.M., 2003, Geochemistry and continental arc origin of the Middle Proterozoic Losee Metamorphic Suite, New Jersey Highlands [MS thesis]: Upper Montclair, New Jersey, Montclair State University, 52 p.
- Černý, P., 1992, Rare-element granitic pegmatites. Part II, Regional to global environments and petrogenesis: Geoscience Canada, v. 18, p. 68-81.
- Corfu, F., and Easton, M., 1995, U-Pb geochronology of the Mazinaw terrane, an imbricate segment of the Central Metasedimentary Belt, Grenville Province, Ontario: Canadian Journal of Earth Sciences, v. 32, p. 959-976.
- Drake, A.A., Jr., 1969, Precambrian and lower Paleozoic geology of the Delaware Valley, New Jersey-Pennsylvania, in Subitsky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guidebook of excursions: New Brunswick, New Jersey, Rutgers University Press, p. 51-131.
- Drake, A.A., Jr., 1984, The Reading Prong of New Jersey and eastern Pennsylvania: an appraisal of rock relations and chemistry of a major Proterozoic terrane in the Appalachians, in Bartholomew, M.J., ed., The Grenville Event in the Appalachians and Related Topics: Boulder, Colorado, Geological Society of America Special Paper 194, p. 75-109.
- Drake, A.A., Jr., Aleinikoff, J.N., and Volkert, R.A., 1991, The Mount Eve Granite (Middle Proterozoic) of northern New Jersey and southeastern New York: U.S. Geological Survey Bulletin 1952, p. C1-C10.
- 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 New Jersey: U.S. Geological Survey Miscellaneous Investigations Series Map I-2540-A, scale 1:100,000.
- Dunn, P.J., 1995, Franklin and Sterling Hill, New Jersey: The world's most magnificent mineral deposits. Franklin, New Jersey, Franklin-Ogdensburg Min. Soc. 755p.
- Fronzel, C., and Baum, J.L., 1974, Structure and mineralogy of the Franklin zinc-iron-manganese deposit, New Jersey: Economic Geology, v. 69, p. 157-180.
- Gates, A.E., and Volkert, R.A., 2004, Vestiges of an Iapetan Rift Basin deposit in the New Jersey Highlands: Implications for the Neoproterozoic Laurentian Margin: Journal of Geodynamics, v. 37, p. 381-409.
- Gorring, M.L., Estelle, T.C., and Volkert, R.A., 2004, Geochemistry of the Late Mesoproterozoic Mount Eve Granite suite: Implications for late to post-Ottawan tectonics in the New Jersey-Hudson Highlands, in Tollo, R.P., Corriveau, L., McLelland, J., and Bartholomew, J., eds., Proterozoic tectonic evolution of the Grenville orogen in North America: Boulder, Colorado, Geological Society of America Memoir 197, p. 505-523.
- Hague, J.M., Baum, J.L., Herrman, L.A., and Pickering, R.J., 1956, Geology and structure of the Franklin-Sterling area, New Jersey: Geological Society of America Bulletin, v. 67, p. 435-474.
- Hotz, P.E., 1952, Magnetite deposits of the Sterling Lake, N.Y.-Ringwood, N.J. area: U.S. Geological Survey Bulletin 982-F, p. 153-244.
- Johnson, C.A., 1997, Genesis of the marble-hosted zinc deposits and iron deposits at Sterling Hill and Franklin, NJ, and comparison with gneiss-hosted iron deposits at Cornwall, NY, in Benimoff, A.I., and Puffer, J.H., eds., The economic geology of northern New Jersey: Field Guide and Proceedings of the 14th Annual Meeting of the Geological Association of New Jersey, p.47-59.
- Johnson, C.A., and Skinner, B.J., 2003, Geochemistry of the Furnace Magnetite Bed, Franklin, New Jersey, and the relationship between stratiform iron oxide ores and stratiform zinc oxide-silicate ores in the New

- Jersey Highlands: Economic Geology, v. 98, p. 837-854.
- Johnson, C.A., Rye, D.M., and Skinner, B.J., 1990, Petrology and stable isotope geochemistry of the metamorphosed zinc-iron-manganese deposit at Sterling Hill, New Jersey: *Economic Geology*, v. 85, p. 1133-1161.
- Klemic, H., Heyl, A.V., Jr., Taylor, A.R., and Stone, J., 1959, Radioactive rare-earth deposit at Scrub Oaks mine, Morris County, New Jersey: *U.S. Geological Survey Bulletin* 1082-B, p. 29-59.
- Kula, J., and Gorrington, M.L., 2000, Tectonic setting and protolith for Middle Proterozoic amphibolites within the Losee Gneiss of the New Jersey Highlands: *Geological Society of America Abstracts with Programs*, v. 32, no. 7, p. A-455.
- Maxey, L.R., 1971, Metamorphism and origin of Precambrian amphibolites of the New Jersey Highlands [Ph.D. dissertation]: New Brunswick, New Jersey, Rutgers University, 156 p.
- McLelland, J.M., and Chiarenzelli, J.R., 1990, Geochronological studies in the Adirondack Mountains and the implications of a Middle Proterozoic tonalitic Suite, *in* Gower, C.F., et al., eds., *Mid-Proterozoic Laurentia-Baltica*: St John's, Newfoundland, Geological Association of Canada Special Paper 38, p. 175-194.
- Metsger, R.W., 1962, Notes on the Sterling Hill ore body, Ogdensburg, N.J., *in* Northern Field Excursion Guidebook, International Mineralogical Association, 3rd General Congress, Washington D.C., p. 12-21.
- Metsger, R.W., 1997, The geology and mining of the Sterling Hill zinc deposit Ogdensburg, Sussex County, New Jersey, *in* Benimoff, A.I., and Puffer, J.H., eds., *The economic geology of northern New Jersey: Field Guide and Proceedings of the Fourteenth Annual Meeting of the Geological Association of New Jersey*, p. 33-46.
- Metsger, R.W., Tennant, C.B., and Rodda, J.L., 1958, Geochemistry of the Sterling Hill zinc deposit, Sussex Co., N.J.: *Geological Society of America Bulletin*, v. 69, p. 775-788.
- Offield, T.W., 1967, Bedrock geology of the Goshen-Greenwood Lake area, N.Y.: New York State Museum and Science Service, Map and Chart Series No. 9, 78p.
- Pearce, J.A., Harris, N.B.W., and Tindle, A.G., 1984, Trace element diagrams for The tectonic interpretation of granitic rocks: *Journal of Petrology*, v. 25, p. 956-983.
- Peck, W.H., Volkert, R.A., Meredith, M.T., and Rader, E.L., 2006, Calcite-Graphite thermometry of the Franklin Marble, New Jersey Highlands: *The Journal of Geology*, v. 114, p. 485-499.
- Peck, W.H., Volkert, R.A., Mansur, A.T., and Doverspike, B.A., 2009, Stable isotope and petrologic evidence for the origin of regional marble-hosted magnetite deposits and the zinc deposits at Franklin and Sterling Hill, New Jersey Highlands, United States: *Economic Geology*, v. 104, No. 7, p. 1037-1054.
- Puffer, J.H., 1997, Genesis of New Jersey Highlands magnetite deposits on the basis of geochemical and structural evidence, *in* Benimoff, A.I., and Puffer, J.H., eds., *The economic geology of northern New Jersey: Field Guide and Proceedings of the 14th Annual Meeting of the Geological Association of New Jersey*, p.71-96.
- Puffer, J.H., and Volkert, R.A., 1991, Generation of trondhjemite from partial melting of dacite under granulite facies conditions: An example from the New Jersey Highlands, USA: *Precambrian Research* v. 51, p. 115-125.
- Puffer, J.H., Pamganamamula, R.V., and Davin, M.T., 1993, Precambrian iron Deposits of the New Jersey Highlands, *in* Puffer, J.H., ed., *Geologic traverse across the Precambrian rocks of the New Jersey Highlands: Field Guide and Proceedings of the Tenth Annual Meeting of the Geological Association of New Jersey*, p. 56-95.
- Rankin, D.W., Chiarenzelli, J.R., Drake, A.A., Jr., Goldsmith, R., Hall, L.M., Hinze, W.J., Isachsen, Y.W., Lidiak, E.G., McLelland, J., Mosher, S., Ratcliffe, N.M., Secor, D.T., Jr., and Whitney, P.R., 1993, Proterozoic rocks east and southeast of the Grenville front, *in* Reed, J.C., Jr., et al., eds., *Precambrian: Conterminous US, The Geology of North America*: Boulder, Colorado, Geological Society of America, v. C-2, p. 335-461.
- Ratcliffe, N.M., Aleinikoff, J.N., Burton, W.C., and Karabinos, P., 1991, Trondhjemitic 1.35-1.31 Ga gneisses of the Mount Holly Complex of Vermont: evidence for an Elzevirian event in the Grenville basement of the United States Appalachians: *Canadian Journal of Earth Sciences*, v. 28, p. 77-93.
- Ries, H., and Bowen, W.C., 1922, Origin of the zinc ores of Sussex County, N.J.: *Economic Geology*, v. 17, p. 517-571.
- Sims, P.K., 1958, Geology and magnetite deposits of the Dover district, Morris County, New Jersey: U.S.

- Geological Survey Professional Paper 287, 162 p.
- Spencer, A.C., Kummel, H.B., Wolff, J.E., Salisbury, R.D., and Palache, C., 1908, Franklin Furnace, New Jersey: U.S. Geological Survey Geologic Atlas Folio 161, 27p.
- Volkert, R.A., 1995, The Byram and Lake Hopatcong Intrusive Suites: Geochemistry and petrogenetic relationship of A-type granites from the New Jersey Highlands: *Northeastern Geology and Environmental Sciences*, v. 17, p. 247-258.
- Volkert, R.A., 2004, Mesoproterozoic rocks of the New Jersey Highlands, north-central Appalachians: Petrogenesis and tectonic history, *in* Tollo, R.P., Corriveau, L., McLelland, J., and Bartholomew, J., eds., *Proterozoic tectonic evolution of the Grenville orogen in North America*: Boulder, Colorado, Geological Society of America Memoir 197, p. 697-728.
- Volkert, R.A., and Drake, A.A., Jr., 1998, The Vernon Supersuite: Mesoproterozoic A-type granitoid rocks in the New Jersey Highlands: *Northeastern Geology and Environmental Sciences*, v. 20, p. 39-43.
- Volkert, R.A., and Drake, A.A., Jr., 1999, Geochemistry and stratigraphic relations of Middle Proterozoic rocks of the New Jersey Highlands: Reston, Virginia, U.S. Geological Survey Professional Paper 1565-C, 77 p.
- Volkert, R.A., and Puffer, J.H., 1995, Late Proterozoic diabase dikes of the New Jersey Highlands-A remnant of Iapetan rifting in the north-central Appalachians: U.S. Geological Survey Professional Paper 1565-A, 22 p.
- Volkert, R.A., Aleinikoff, J.N., and Fanning, M.C., in press, Tectonic, magmatic and metamorphic history of Mesoproterozoic rocks, New Jersey Highlands: New insights from SHRIMP U-Pb geochronology, *in* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P. M., eds., *From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region*: Geological Society of America Memoir 206.
- Volkert, R.A., Johnson, C.A., and Tamashausky, A.V., 2000a, Mesoproterozoic Graphite deposits, New Jersey Highlands: Geologic and stable isotopic evidence for possible algal origins: *Canadian Journal of Earth Sciences*, v. 37, p. 1665-1675.
- Volkert, R.A., Zartman, R.E., and Moore, P.B., 2005, U-Pb zircon geochronology of Mesoproterozoic postorogenic rocks and implications for post-Ottawan magmatism and metallogenesis, New Jersey Highlands and contiguous areas, USA: *Precambrian Research*, v. 139, p. 1-19.
- Volkert, R.A., Feigenson, M.D., Patino, L.C., Delaney, J.S., and Drake, A.A., Jr., 2000b, Sr and Nd isotopic compositions, age and petrogenesis of A-type granitoids of the Vernon Supersuite, New Jersey Highlands, USA: *Lithos*, v. 50, p. 325-347.
- Whalen, J.B., Currie, K.L., and Chappell, B.W., 1987, A-type granites: geochemical characteristics, discrimination and petrogenesis: *Contributions to Mineralogy and Petrology*, v. 95, p. 407-419.
- Young, D.A., and Cuthbertson, J., 1994, A new ferrosilite and Fe-pigeonite occurrence in the Reading Prong, New Jersey, USA: *Lithos*, v. 31, p. 163-176.

