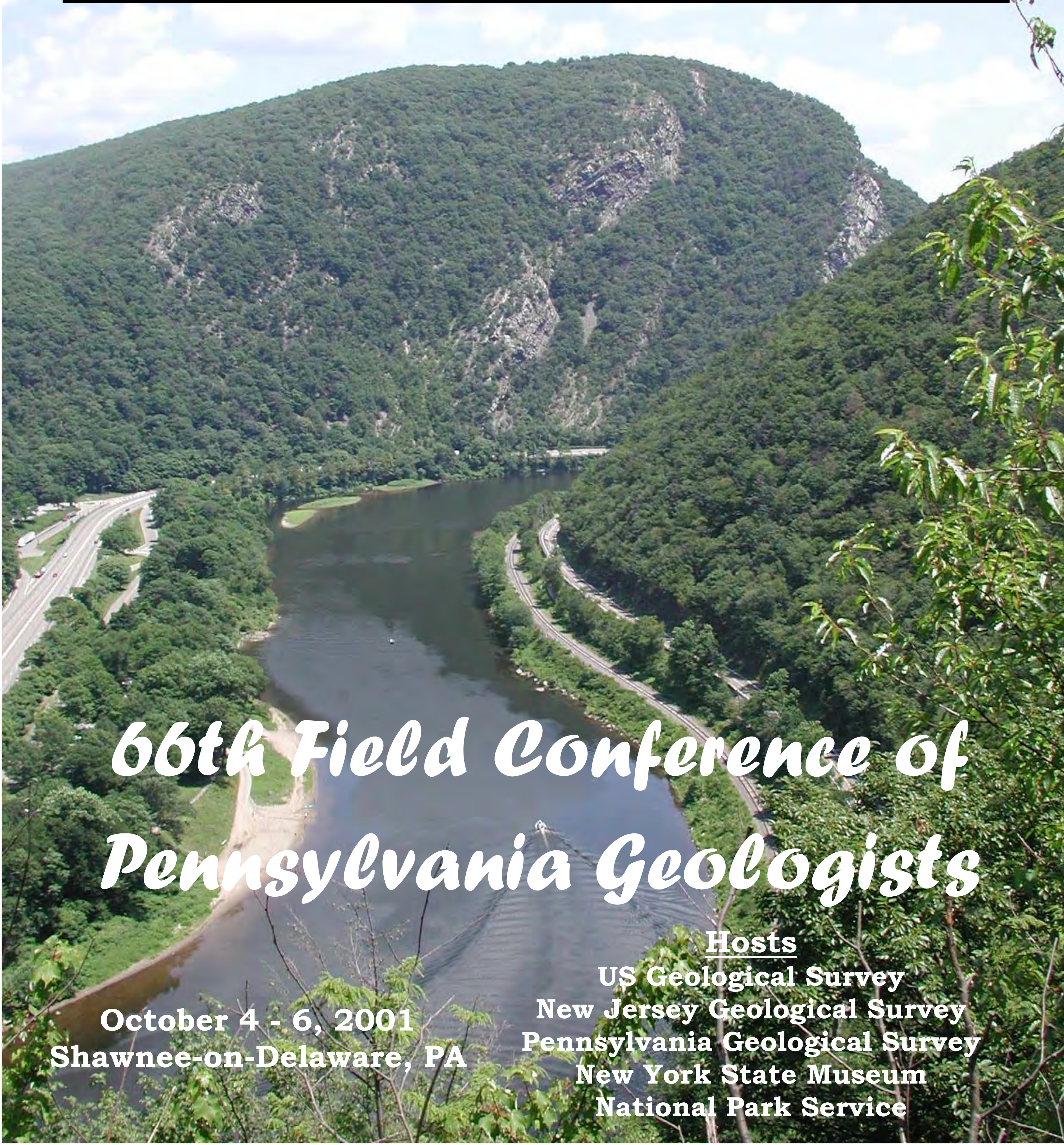
# 2001: A DELAWARE RIVER ODYSSEY



# 66th Field Conference of Pennsylvania Geologists

**Hosts**

US Geological Survey
New Jersey Geological Survey
Pennsylvania Geological Survey
New York State Museum
National Park Service

October 4 - 6, 2001
Shawnee-on-Delaware, PA

Guidebook for the
66th Annual Field Conference of Pennsylvania Geologists

# 2001: A DELAWARE RIVER ODYSSEY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October 4-6, 2001

*Hosts:* U.S. Geological Survey, New Jersey Geological Survey, Pennsylvania Geological Survey, New York State Museum, and National Park Service.

*Headquarters:* Shawnee Inn, Shawnee-on-Delaware, PA

*Cover:* Delaware Water Gap, view from the Pennsylvania side. (Photo by Gary M. Fleeger.)


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**Group photo of the 2000 Field Conference of Pennsylvania Geologists.**
**North Park, Allegheny County**

1967

1973

1977

2000

**This Field Conference and Guidebook are dedicated—by his many friends—to**

# Jack B. Epstein, U.S. Geological Survey,

**for his voluminous contributions over the years to our understanding of the geology of Pennsylvania and New Jersey.**

1984

1983

1982

# A (Very) Few Words about Jack Epstein

Jack Epstein was born a long time ago, probably beyond the memory of most of us. He started his geological career during the soporific '50's, graduating from Brooklyn College with a B.S. in geology in 1956 and from the University of Wyoming with an M.A. in 1958. His Master's thesis, *Geology of the Fanny Peak quadrangle, Wyoming-South Dakota*, probably isn't as exiting as it sounds, but it did give him a good background for some of his early work with the U.S. Geological Survey in the Black Hills and Montana's Gallatin Range. Things got a bit more exciting in the '60's, as Jack made permanent status with the U.S.G.S. and was transferred east to begin what one could almost call his "life's work"—mapping in the Valley and Ridge of eastern Pennsylvania and adjacent New Jersey. He started in the Stroudsburg quadrangle (work he eventually turned into a 1970 Ph. D. dissertation, *Geology of the Stroudsburg quadrangle and adjacent areas, Pennsylvania-New Jersey*, at Ohio State University) and then moved on to the Wind Gap and Lehighton-Palmerton quadrangles. These early projects all showed that Jack had a great proficiency for mapping both bedrock and surficial deposits—something that very few of us have developed to such a high degree. In those early days, he also proved himself to be quite a geomorphologist, writing one of the landmark papers in Appalachian drainage development. *Structural control of wind gaps and water gaps…in the Stroudsburg area* (1966).

His subsequent career has certainly lived up to this early promise. He not only prepared numerous bedrock and surficial geologic maps of various scales, but also authored or co-authored many topical stratigraphic and structural reports (see *Bibliography* of this guidebook). The "crown jewel" of these reports is—arguably at least—U.S.G.S. Map I-1422, *Geologic map of Cherry and Godfrey Ridges in the Saylorsburg, Stroudsburg, and East Stroudsburg quadrangles, Monroe County, Pennsylvania* (1989). Just since 1997, Jack has been involved in karst projects in both Pennsylvania and the Midwest, aquifer vulnerability studies in the Black Hills, building subsidence in the Philadelphia area, and geologic mapping in various National Parks. Whew! The latter brings us down to the present and to the immediate subject of this Field Conference—the geology of the Delaware Water Gap National Recreation Area. It seems appropriate that Jack returns to the site of one of his earliest field areas with the USGS.

Over the last few years Jack has often been known to mutter about retiring to unknown destinations—to being a founding member of "Ollie's Army" and filling his time as a handyman, it is said. Yes, he may retire one of these days, finally finding time to cut his yard and fix his pool—and when he does, geology would seem to have lost one of its real "Rock Stars." But, knowing Jack, he'll probably just keep working that much harder, seeing that he has two jobs to do. And be sure to look for him at the 2002 Field Conference!

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# PREFACE

The Field Conference last visited this part of northeastern Pennsylvania 34 years ago (Epstein and Epstein, 1967). Then, the geology between Delaware Water Gap and Lehigh Gap was highlighted, although a couple of stops were in the Delaware Water Gap National Recreation Area (DEWA in National Park Service parlance). On this trip we will concentrate on the area immediately surrounding and within the DEWA. The National Park Service is a sponsor of this year’s excursion, along with the U.S. Geological Survey, New Jersey Geological Survey, Pennsylvania Geological Survey, and New York State Museum.

The rocks that we will see range in age from Middle Ordovician to Middle Devonian, overlain by a complex of Pleistocene and Holocene surficial deposits. As befitting a recreation area visited by several millions of citizens yearly, thirsty for a user-friendly interpretation of the geologic environment around them, this trip will cover a wide variety of topics: Pleistocene glacial history, Paleozoic stratigraphy and its interpretation, geomorphic development of this part of the Appalachian Mountains, origin of waterfalls, structural geology including age of deformation (Taconic-Acadian-Alleghanian), mineral resources, geologic hazards, vertebrate and invertebrate paleontology, and geoarcheology. In addition to detailed road logs of the geologic and historic features along several of the main highway corridors through the park, this guidebook also includes several geologic guides to scenic and historic trails and a river guide for canoeists from Bushkill to Smithfield Beach.

In short, there’s not much about the landscape, geology, and history of the Delaware Water Gap area that won’t be discussed sometime during this Field Conference. So, take a good window seat on the bus (if you’re lucky), and, to paraphrase Harry Golden, “Enjoy, Enjoy,!”

# ACKNOWLEDGMENTS

The organizers and editors are indebted to many individuals, businesses, and organizations for their assistance and cooperation in making this 2001 Field Conference possible. First of all, we thank the many contributors to the guidebook, the majority of whom are representatives of federal and state agencies that have a public charge to elucidate—and preserve—our common geological and environmental heritage. Then there are the numerous private individuals and businesses who generously allowed us to use sites on their properties for STOPS or to investigate geologic features pertinent to our investigations: John W. Briggs, Resorts USA, Inc.; Ron Pickel and Brig Burgess, Norfolk Southern Corp.; Yu Tian Cheu, E-Hi Art Studio; Frank Riccobono, North Park Development; and J. J. Geuther, Yards Creek Generating Station. Much historical and geographical information was obtained from the Minisink Valley Historical Society (Peter Osborne, Executive Director), and the Pike County Historical Society (Lori Strelecki, Curator). William Moore (Susquehanna University) and Justin Gindlesperger (Shippensburg University), student interns with the Pennsylvania Geological Survey, provided yeomanly field assistance during the summers of 2000 and 2001, respectively. Helen Delano, Pennsylvania Geological Survey, prepared the route map for Day 2. This Field Conference would not have been possible without the enthusiastic cooperation of Bill Laitner, Rab Cika, and John Wright of the DEWA National Park Service. And last, but far from least, we all owe a great debt of gratitude to Gary Lambert, sales manager at Shawnee Inn and Golf Resort, whose imaginative assistance made possible the use of Shawnee Inn as our 2001 headquarters.

# SHAWNEE INN AND GOLF RESORT

The 66th Annual Field Conference of Pennsylvania is being held in one of the most storied hostelries and resorts in eastern Pennsylvania. Long famous for its beautiful surroundings, PGA-sanctioned golf course and hospitality, the Shawnee Inn and Golf Resort now includes a nearby ski resort and an on-site ice rink. Located on a triangularly shaped alluvial terrace at the mouth of Shawnee Creek, the inn was originally opened as the “Buckwood Inn” under the ownership of C. C. Worthington (of Worthington Pump and Machinery Corp.) in May of 1911. To help attract the “crème de la crème” of 20th-century business leaders, Worthington himself designed a nine-hole golf course adjacent to the inn—and soon after hired A. W. Tillinghast (now famous for his later designs at Baltusrol in Springfield, New Jersey, and Winged Foot in Mamaroneck, New York) to lay out an additional 18 holes on Shawnee Island. In 1912, Buckwood was the site of the founding meeting of what became the present-day Professional Golfers Association (PGA); and in 1938, it gained fame as the site of the PGA Tournament in which Sam Snead lost to Paul Runyon on the last hole of match play.

In 1943, Worthington sold the Buckwood to musician Fred Waring, who renamed it the Shawnee Inn. Being at the height of his popularity, Waring made the Inn the center of all his musical activities. From there, he broadcasted his famous radio programs—and Shawnee became known as the home of “Fred Waring and his Pennsylvanians.” (Waring also gained fame and the everlasting appreciation of cooking enthusiasts for designing the “Waring Blender.”) Many of his celebrity friends—Jackie Gleason, Milton Berle, Art Carney, Ed Sullivan, Eddie Fisher, and Perry Como, among others—came to Shawnee as guests. In the ‘50’s and ‘60’s, Arnold Palmer often played at Shawnee. (A delightful picture of a bemused Palmer looking on as Gleason does his trademark dance-step holding a golf club—“And awaaaaayyyyyy we go!”—can be seen on the wall near the entrance to the Inn, as well as on page 156 of this guidebook.)

Waring sold the entire property to Karl Hope, a successful real estate developer from Philadelphia, in 1974. Hope turned the Inn into a year-round resort and conference center. He built Shawnee Mountain Ski Area on the Pennsylvania extension of Wallpack Ridge about two miles to the northeast, and hired Jean-Claude Killy, Olympic gold medallist, to head ski operations there.

After just three years, Hope sold his operations to Charles and Virginia Kirkwood, the present owners. Over the next two decades, the Kirkwoods expanded Shawnee’s golfing facilities, added an indoor pool, opened Shawnee Canoes, developed 100 percent snowmaking capabilities at Shawnee Mountain, and built the ice rink. As time permits, Field Conference participants are encouraged to take advantage of such seasonally appropriate facilities as are available.

Shawnee’s scenic location on the shore of the Delaware River between “Wallpack Ridge” on the northwest and Kittatinny Mountain on the southeast does have some disadvantages. Both the inn and golf course are situated on the floodplain terrace of the river, only 20 to 40 feet above normal water level. At several times in the past century, the area has been inundated by floodwaters, notably in 1936, 1955 (Hurricane Diane), 1972 (Tropical Storm Agnes), and 1996. Though the 1955 flood was the worst on record, the flood of January 20, 1996 was particularly destructive, not only because it occurred just after extensive renovations had been completed on the ground floor of the inn but also because rampaging ice blocks tore up the fairways and greens of the golf course to an incredible degree. The water was about 6 feet deep around the inn, and ice was piled up to a depth of 4 feet on the course. Shawnee Inn was closed for 10 weeks—from late January to early April, and at one point it was thought that the golf course was too badly scarred to be saved. Total damage amounted to more than $2 million. Historical pictures of the golf course and the 1996 devastation can be viewed on a wall in the Inn’s Golf Pro Shop.

**Shawnee Inn and Golf Resort, Shawnee-on-Delaware, Pennsylvania.**

# STRATIGRAPHY IN THE REGION OF DELAWARE WATER GAP NATIONAL RECREATION AREA

by

Jack B. Epstein

## INTRODUCTION

Field mapping in the folded Appalachian Mountain and Great Valley sections of the Valley and Ridge physiographic province of eastern Pennsylvania and northern New Jersey by the U.S. Geological Survey, New Jersey Geological Survey, and Pennsylvania Geological Survey has led to a better understanding of all aspects of Appalachian geology. Disagreements have been common since H. D. Rogers first described the geology of the area in 1858. Many differing opinions still exist regarding the stratigraphy, structural geology, geomorphology, and glacial geology. The rocks in the area range from Middle Ordovician to Late Devonian in age. This diversified group of sedimentary rocks was deposited in many different environments, ranging from deep sea, through neritic and tidal, to alluvial. In general, the Middle Ordovician through Lower Devonian strata are a sedimentary cycle related to the waxing and waning of Taconic tectonism. The sequence began with a graywacke-argillite suite (Martinsburg Formation) representing synorogenic basin deepening. This was followed by basin filling and pro-gradation of a sandstone-shale clastic wedge (Shawangunk Formation and Bloomsburg Red Beds) derived from the erosion of the mountains that were uplifted during the Taconic orogeny. The sequence ended with deposition of many thin units of carbonate, sandstone, and shale on a shelf marginal to a land area of low relief. Another tectonic-sedimentary cycle, related to the Acadian orogeny, began with deposition of Middle Devonian rocks. Deep-water shales (Marcellus Shale)

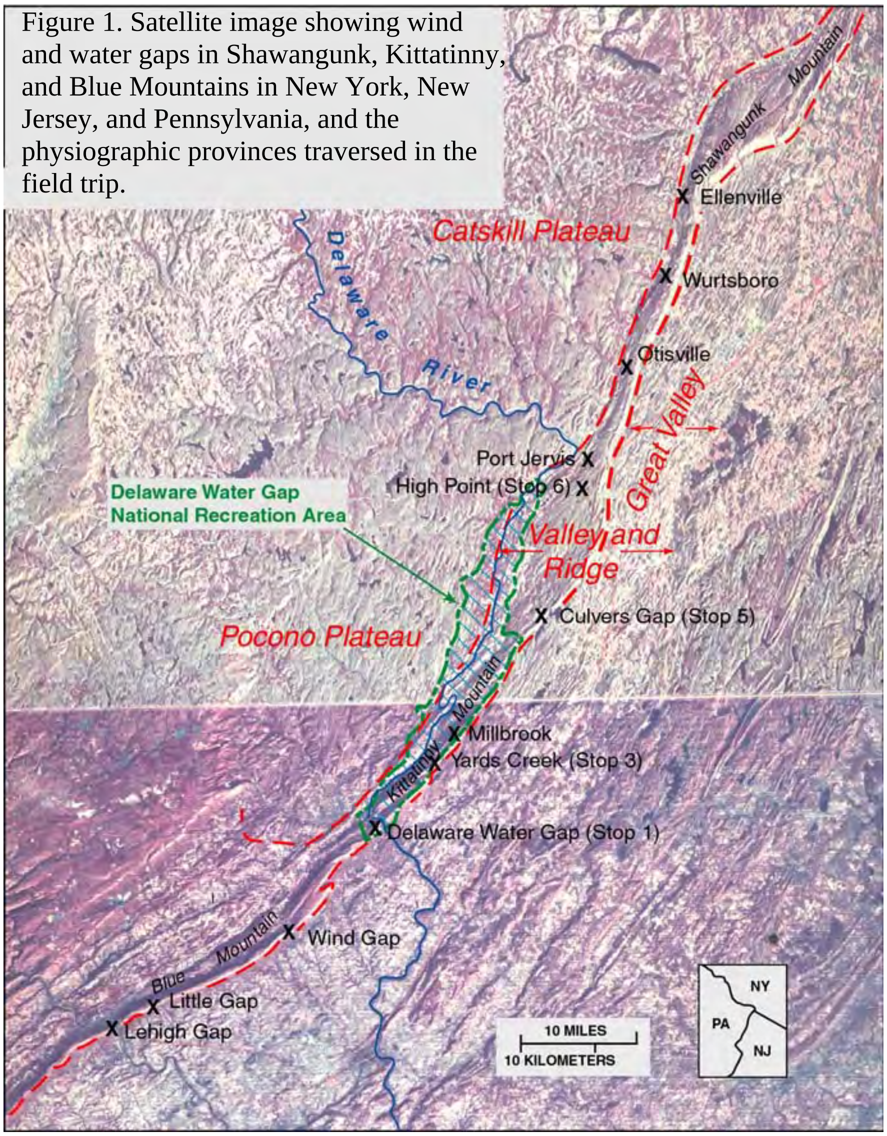



Figure 1. Satellite image showing wind and water gaps in Shawangunk, Kittatinny, and Blue Mountains in New York, New Jersey, and Pennsylvania, and the physiographic provinces traversed in the field trip.

Epstein, J. B., 2001, Stratigraphy in the region of the Delaware Water Gap National Recreation Area, *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. 1 - 13 .

preceded shoaling (Mahantango Formation) and turbidite sedimentation (Trimmers Rock Formation) followed by another molasse (Catskill Formation).

## STRATIGRAPHY

The Ordovician, Silurian, and Devonian rocks, and overlying surficial deposits that will be seen on this Field Conference, lie mainly within the Valley and Ridge physiographic province and partly within the Great Valley of northeastern Pennsylvania and northwestern New Jersey (Figure 1). The first significant geologic study of the area was the magnificent treatise of H. D. Rogers (1858), although other less comprehensive reports appeared before. Since that time, abundant studies have resulted in an understanding of many stratigraphic details, but they have also spawned many controversies that appear to become more numerous as years go by. What follows is a terse summary of results of stratigraphic investigations and remaining problems that provide interesting research potential.

The stratigraphic sequence from the Martinsburg Formation of Ordovician age through the Catskill Formation of Devonian age comprises more than 25,000 feet (7600 m) of shale, siltstone, sandstone, conglomerate, limestone, and dolomite. The salient lithic types and the thickness of the stratigraphic units, other than the Catskill, are given in Table 1. Correlation charts are given for Lower through Upper Silurian clastic rocks (Figure 2) and for complex Upper Silurian and Lower Devonian rocks (Figure 3A). Our understanding of the physical stratigraphy has been increased by reports and dissertations on the Catskill Formation (Glaeser, 1963; Epstein et al., 1974; Berg et al., 1977), Onesquethawan rocks (Inners, 1975; Epstein, 1984; Ver Straeten, 1996a, 1996b), Upper Silurian and Lower Devonian rocks (Epstein et al., 1967), the Shawangunk Formation (Epstein and Epstein, 1972; Epstein, 1993), and the Martinsburg Formation (Drake and Epstein, 1967; Lash et al., 1984), to name a few.

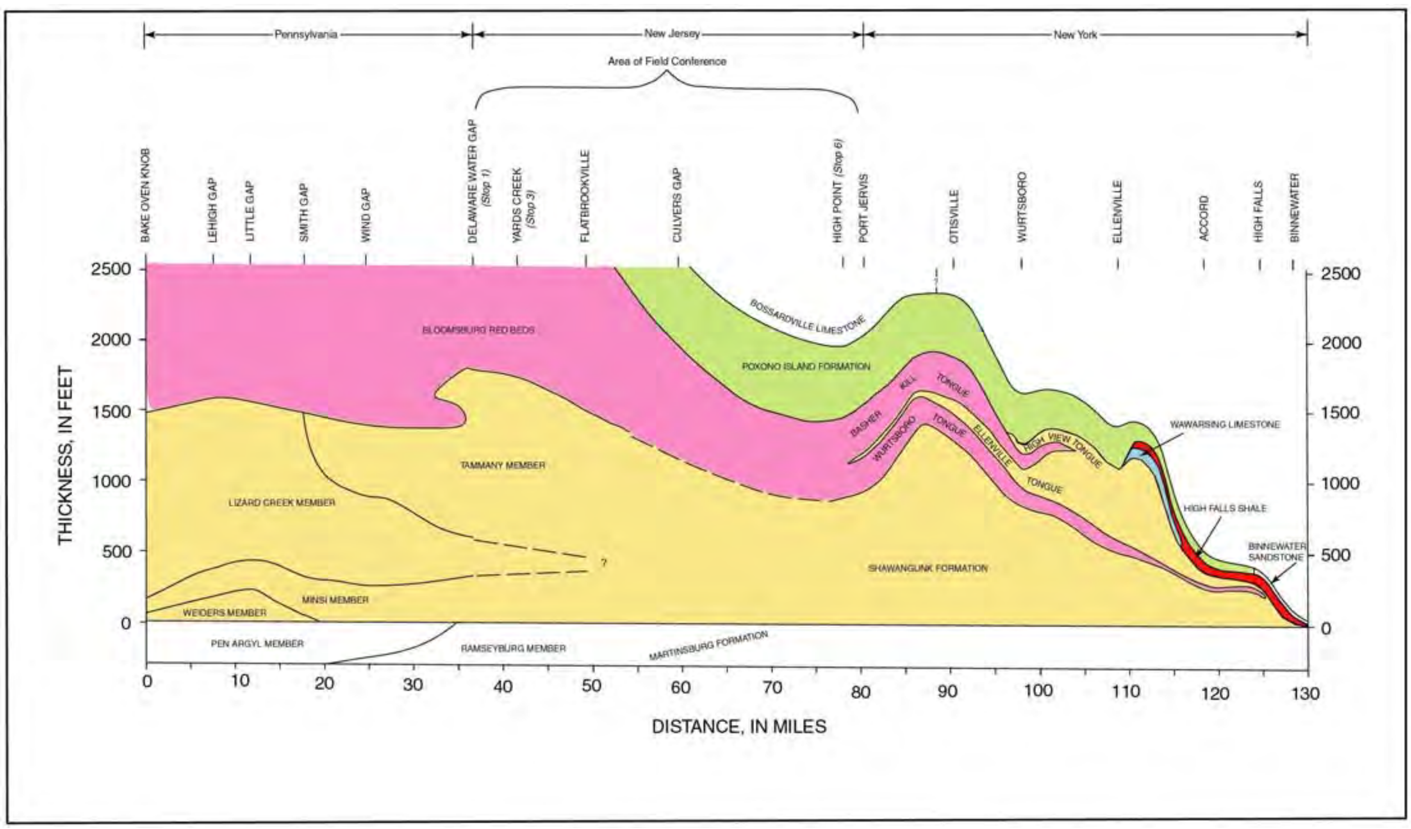



Figure 2. Correlation chart of Silurian rocks from southeastern New York, through New Jersey, and into eastern Pennsylvania. Modified from Epstein (1972, 1993). See Figure 1 for location of most of the sections.

Table 1. Description of rock units in the Field Conference area

| System | Series | Group | Formation | Member | Description | Thickness (feet) |
|---|---|---|---|---|---|---|
| | Upper | | Trimmers Rock | Millrift | Dark-gray to medium-dark gray siltstone, shale, and very fine-grained sandstone, coarsening upwards. Fossiliferous (brachiopods). | 720-1,825 |
| | | | | Sloat Brook | | |
| | Middle | | Mahantango | | Medium-dark-gray siltstone and silty shale. Fossiliferous, biostromes (corals, brachiopods, pelecypods, bryozoans). | 1,300-2,450 |
| | Middle | | Marcellus | Brodhead Creek | Dark-gray, laminated to poorly bedded silty shale; depauperate brachiopods. Medium-dark gray shaly limestone. | 800-950 |
| | | | | Stony Hollow | Medium-dark-gray to medium-gray, laminated to thin-bedded, shaly limestone, fossiliferous (brachiopods). | 150 |
| | | | | Union Springs | Medium-dark-gray to dark-gray laminated shale; sheared along detachment. | 50 |
| | | | Onondaga (Buttermilk Falls) | Seneca | Fossiliferous cherty limestone. Contains TIOGA ash bed. | 15 |
| | | | | Moorehouse (Stroudsburg) | Medium-gray limestone and argillaceous limestone with beds, pods and lenses of dark-gray chert. Fossiliferous (brachiopods, ostracodes), burrowed. | 135 |
| | | | | Nedrow (McMichal) | Medium-dark-gray calcareous argillite with lenses of light-medium gray fossiliferous limestone. | 40 |
| | | | | Edgecliff (Foxtown) | Medium-dark-gray calcareous siltstone and argillaceous limestone containing lenses of dark-gray chert. Fossiliferous, one-inch diameter crinoid "colum-nals" in lower half. | 80 |
| | Lower | | Schoharie | | Medium-to medium-dark gray argillaceous calcareous siltstone. Fossiliferous (brachiopods, *Taonurus* burrows in lower half, vertical burrows in upper half). | 100-150 |
| | | | Esopus | | Medium- to dark-gray silty shale and shaly to finely arenaceous siltstone. Poorly fossiliferous. Burrowed (*Taonurus*). | 180-300 |
| | | Oriskany Group | Ridgeley | | Light- to medium-gray, fine- to coarse-grained calcareous sandstone and quartz-pebble conglomerate with minor siltstone, arenaceous limestone, and dark-gray chert. Fossiliferous (brachiopods). | 0-16 |
| | | | Shriver Chert | | Medium-dark-gray siliceous calcareous shale and siltstone and beds, lenses, and pods of dark-gray chert and minor calcareous sandstone. Fossiliferous (brachiopods), burrowed. | 50-85 |
| | | Helderberg Group | Port Ewen Shale | | Medium-dark-gray poorly fossiliferous, irregularly laminated calcareous shale and siltstone grading up to fossiliferous, burrowed, irregularly bedded calcareous siltstone and shale. | 150 |
| | | | Minisink Limestone | | Dark- to medium-gray argillaceous fossiliferous limestone. | 11-14 |
| | | | New Scotland | Maskenozha | Dark-gray silty calcareous laminated fossiliferous shale with lenticular argillaceous fossiliferous limestone. | 45 |
| | | | | Flatbrookville | Medium-dark-gray silty and calcareous fossiliferous shale with lenticular medium-gray argillaceous, very fossiliferous limestone. | 20-33 |
| | | | Coeymans | Stormville | Medium-gray, fine- to coarse-grained, biogenic limestone, fine-to medium-grained arenaceous limestone, fine- to coarse-grained, crossbedded and planarbedded calcareous sandstone and quartz-pebble conglomerate, with some dark-gray chert. Fossiliferous (brachiopods, crinoids). | 0-20 |
| | | | | Shawnee Island / Thacher Mbr of Manlius Fm | Shawnee Island: Medium-gray, argillaceous and arenaceous irregularly bedded fossiliferous and burrowed limestone with chert at top. Contains bioherms of medium-light-gray very coarse grained crudely bedded biogenic limestone with corals, stromatoporoids, and shelly fauna (*Gypidula*). Thacher: Dark-gray, unevenly bedded limestone with yellowish-gray shale partings. | 0-60 / 0-35 |
| | | | | Kalkberg Limestone | Medium-dark gray argillaceous massive fossiliferous limestone (diversified fauna) with nodules and lenses of dark-gray chert. | 0-60 |
| | | | | Peters Valley | Medium-gray arenaceous limestone to light-medium-gray fine- to coarse-grained pebbly calcareous sandstone. Cross bedded, fossiliferous. | 0-9 |
| | | | | Depue Limestone | Medium- to dark-gray arenaceous and argillaceous fossiliferous Limestone. | 13-29 |

| | | | | | |
|---|---|---|---|---|---|
| | | | Ravena | Medium-dark-gray slightly argillaceous, fossiliferous limestone. | 0-30 |
| SILURIAN AND DEVONIAN | Up. Silurian & Low. Devonian | Rondout | Mashipacong | Medium-dark- to light-gray shale, calcareous shale, and very fine- to medium-grained argillaceous limestone. Mudcracks, cut and fill. | 8-15 |
| | | | Whiteport Dolomite | Dark- to medium-gray mud-cracked laminated dolomite. | 5-10 |
| | | | Duttonville | Dark- to medium-gray calcareous shale and argillaceous limestone. Mud-cracked intervals and biostromal limestone beds. | 10-20 |
| SILURIAN | Upper | Decker | Wallpack Center / Clove Brook | Wallpack Center: Lenticular and evenly bedded quartz-pebble conglomerate, calcareous sandstone and siltstone, argillaceous and arenaceous limestone and dolomite. Cross bedded, planar bedded, flaser bedded, fossiliferous.<br>Clove Brook: Medium-gray to medium-dark gray fossiliferous (crinoidal) limestone with light-olive-gray shale partings near top. | 0-85 / 0-50 |
| | | Bossardville Limestone | | Dark- to medium-gray, laminated argillaceous limestone locally containing deep mud cracks (as much as 20 feet deep) grading up to dark-gray laminated limestone. Poorly fossiliferous (ostracodes). | 12-110 |
| | | Poxono Island | | Light-olive-gray to green, calcareous and dolomitic, laminated, fissile to nonfissile shale, olive-green dolomite, sandstone, and siltstone. | 500-800 |
| | Middle & Upper | Bloomsburg Red Beds | | Red, green, and gray siltstone, shale, sandstone, and conglomeratic sandstone in upward-fining sequences. Cross-bedded and laminated, mud cracks, cut and fill, scattered ferroan dolomite concretions. Partly burrowed. Fish scales locally. | 1,500 |
| | Lower and Middle | Shawangunk (Members loose their identity several miles northeast of Delaware Water Gap) | Tammany | Gray, fine- to coarse-grained, partly crossbedded, pyritic conglomerate, evenly bedded quartzite, and about 2% dark-gray argillite. | 800 |
| | | | Lizard Creek | Gray to olive-gray, fine- to coarse-grained, partly crossbedded, pyritic, thin- to thick-bedded quartzite interbedded with thin-to thick bedded, gray argillite. | 275 |
| | | | Minsi | Gray to olive-gray, fine- to coarse-grained, partly crossbedded, pyritic and feldspathic, thin- to thick-bedded quartzite, conglomeratic quartzite, and conglomerate. Locally contains mud-cracked argillite. | 300 |
| ORDOVICIAN | Middle and Upper Middle and Upper | Martinsburg | Pen Argyl | Dark-gray to grayish black, thick- to thin-bedded (some beds more than 20 feet thick), evenly bedded claystone slate, rhythmically interbedded with quartzose slate, subgraywacke, and carbonaceous slate. Taconic unconformity at top. Disappears under Shawangunk about one mile west of Delaware Water Gap. | 3,000-6,000 |
| | | | Ramseyburg | Medium- to dark-gray claystone slate alternating with light- to medium-gray, thin- to thick-bedded graywacke and graywacke siltstone. | 2,800 |
| | | | Bushkill | Dark- to medium-gray thin-bedded (beds do not exceed six Inches thick), claystone slate with thin interbedded quazrtzose slate and graywacke siltstone and carbonaceous slate. Not exposed in Delaware Water Gap National Recreation Area. | 4,000 |

Several problems remain. Whereas stratigraphic relationships of Upper Silurian through lower Middle Devonian strata are well known between Delaware Water Gap and Lehigh Gap, 29 miles southwest of Delaware Water Gap, this group of rocks is poorly understood for at least 20 miles southwest of Lehigh Gap. The entire sequence becomes thinner and more clastic, and several units disappear as an ancient low-lying positive area is approached to the southwest near Harrisburg, PA. This positive area was termed the “Harrisburg axis” by Ulrich (1911) and Willard (1941), and named the

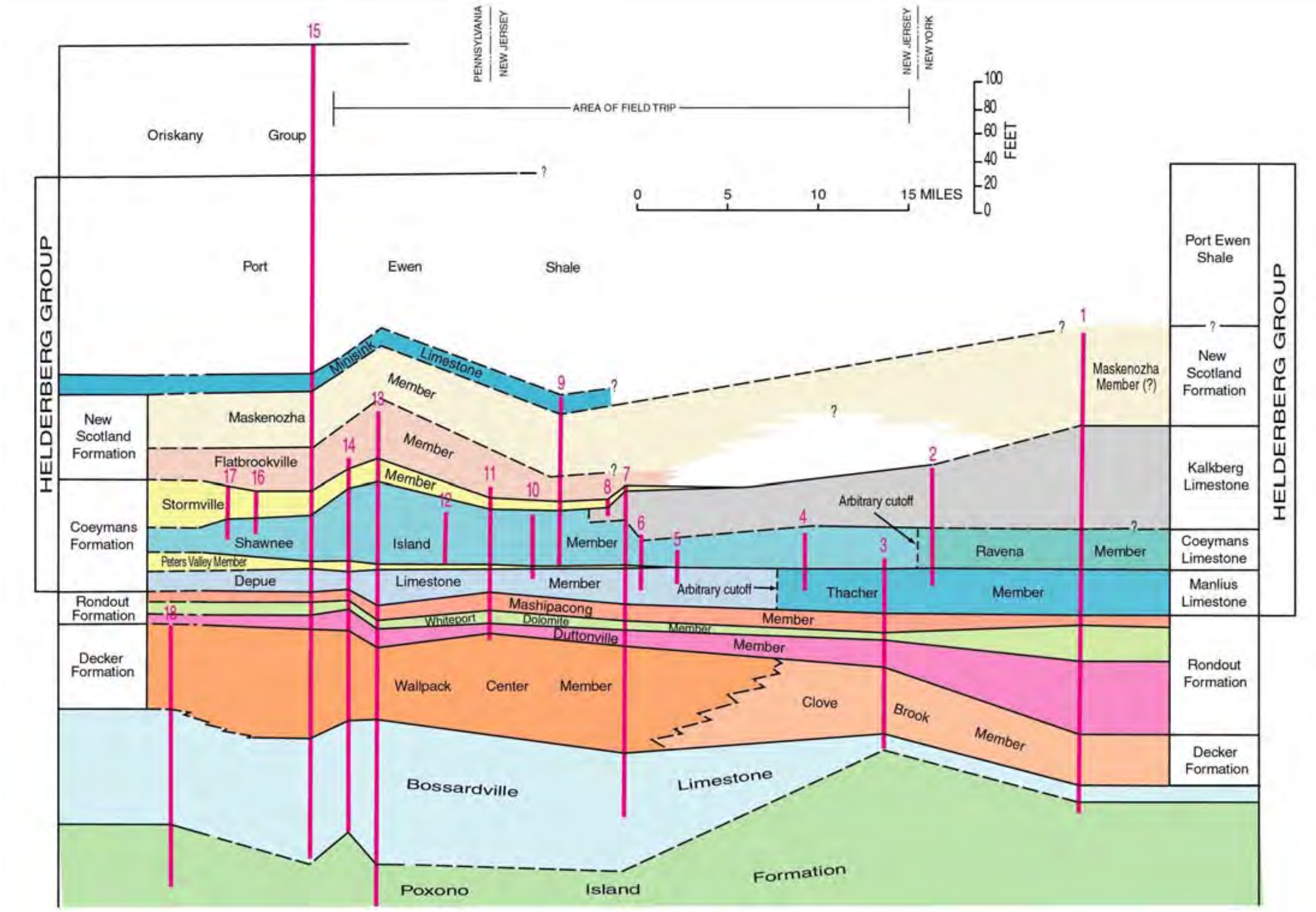



Figure 3A. Correlation chart of Upper Silurian and Lower Devonian formations in the vicinity of the Delaware Water Gap National Recreation Area, northeastern Pennsylvania, northern New Jersey, and southeasternmost New York. Modified from Epstein et al., 1967. Section locations shown on Figure 3B.

"Auburn Promontory" by Swartz (in Willard et al., 1939). Poor exposures, abrupt facies changes, and limited paleontologic data have hampered our understanding of these rocks.

The ages of the rocks in the Valley and Ridge of eastern Pennsylvania are generally well known, but the exact locations of the two systemic boundaries within the sequence are still a bit speculative. The Ordovician-Silurian boundary is generally accepted (incorrectly?) as being at the unconformable contact between the Martinsburg and Shawangunk Formations. The uppermost beds of the Martinsburg are late Middle Ordovician in age (Berry, 1970), but the basal beds of the Shawangunk have not yielded diagnostic fossils. On the basis of regional considerations, the basal Shawangunk could be uppermost Ordovician in age (Epstein and Epstein, 1972). Thus, the dating of the Taconic unconformity in eastern Pennsylvania is still open to question.

The location of the Silurian-Devonian boundary is a somewhat lesser, but nonetheless important, problem. The lowest Coeymans Formation is definitely known to be Devonian, on the basis of conodonts studies (A. G. Harris, oral communication, 1982), and the Decker Formation is possibly Silurian in age. Thus, the boundary may lie within about 30 feet (9 m) of poorly fossiliferous dolomite, shale, and limestone of the intervening Rondout Formation, or possibly within sandstone, limestone, and dolomite of the Decker Formation. Recent attempts to collect conodonts from this interval have so far

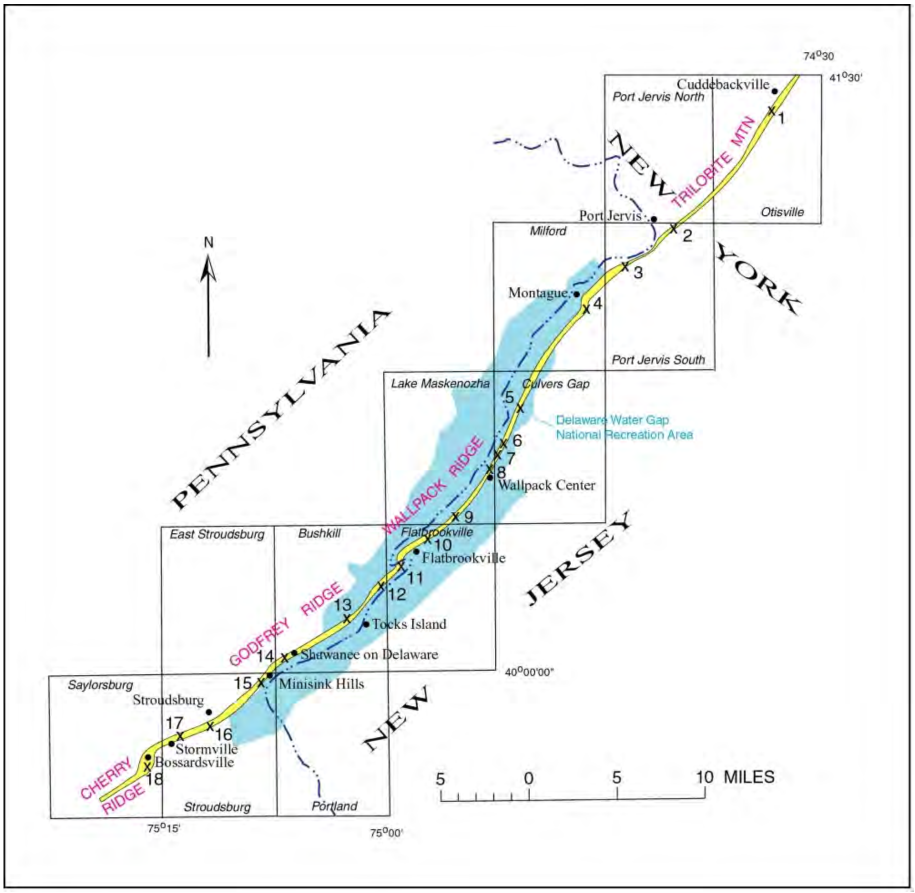



Figure 3B. Outcrop belt (yellow) of uppermost Silurian and lowermost Devonian rocks in the Delaware Water Gap National Recreation Area and location of measured sections shown in Figure 3A. Modified from Epstein et al., 1967.

failed to resolve this problem. A discussion of the long-standing debate of the position of the Silurian-Devonian boundary is given in Epstein et al. (1967).

One of the most vexing sedimentological problems in the folded Appalachians is the source of debris for many of the thick clastic wedges in the Paleozoic succession. The Shawangunk Formation of Silurian age, with its abundant quartz sand and quartz pebbles, is one example. It overlies a thick lower Paleozoic section of slate and carbonate rocks of the Great Valley, and Precambrian metasedimentary rocks, amphibolite, marble, and granitic rocks in the Reading Prong. A comparison of the mineralogy of these rocks does not make the rocks beneath the Shawangunk an enticing source for the Shawangunk. It is possible that pre-Silurian structural shuffling may have brought a source terrane in juxtaposition with Shawangunk depositional basin that was more quartz rich than the rocks presently south of the Shawangunk outcrop belt (Epstein and Epstein, 1972).

A major controversy that still exists after nearly a century of debate concerns the number of members within the Martinsburg Formation. The arguments have been based on both faunal and structural evidence. In general, those workers who have studied the Martinsburg west of the Lehigh River have divided it into two parts: a lower slate unit and an upper sandstone unit (e.g., Stose, 1930; Willard, 1943; Wright and Stephens, 1978). In the Delaware Valley many geologists favor a tripartite subdivision: two slate belts separated by a middle sandstone-bearing unit. Behre's (1933) work was the most detailed in the slate belt, but his threefold subdivision was not accepted on the 1:250,000-scale, 1960-vintage Pennsylvania state geologic map (Gray et al., 1960), although the three belts of rock are clearly shown. Those who support a two-member division maintain that the northern slate belt is actually the southern slate belt repeated by folding. Detailed stratigraphic and structural evidence presented later by Drake and Epstein (1967) showed that the Martinsburg can be divided into three mappable members (see Table 1) in almost the same way as defined by Behre (1933). This should not be surprising because the best geologist of all, the slate quarrymen who have toiled over the Martinsburg since the first half of the 19th century have long recognized two distinct slate belts in the Martinsburg Formation of eastern Pennsylvania and northwestern New Jersey—the "hard slate" belt in the south and the "soft slate" belt in the north. They are separated by a zone that contains a poorer quality of slate because appreciable grace (dirty sandstone) is interbedded with the slate.

Differences in stratigraphic interpretation have led to various thickness estimates. Those who accept a two-fold interpretation have estimated that the Martinsburg is as thin as 3000 feet (Stose, 1930), whereas those who support the idea of three members have estimated thicknesses of more than 10,000 feet (Behre, 1933; Drake and Epstein, 1967). Wright et al. (1979) recognized five graptolite zones in the Martinsburg Formation in the Lehigh River area and suggested that the Pen Argyl and Bushkill Members are the same age and are simply repeated by folding. This contradicts detailed mapping in the Lehigh area (Epstein et al., 1974; Lash, 1978), as well as in the Delaware Water Gap Area (Epstein, 1973, 1990) which clearly shows that the Bushkill, Ramseyburg, and Pen Argyl Members are part of a progressively younging sequence—the Pen Argyl stratigraphically overlies the Ramseyburg as is demonstrated wherever there are adequate exposures at or near the contact. Where the Ramseyburg structurally overlies the Pen Argyl, it can be shown that the contact is overturned (e.g., Figure 6 in Epstein, 1980). Furthermore, the lithic characteristics of the Bushkill and Pen Argyl are very different. The Bushkill is a ribbon slate: beds are never more than 6 inches thick and are generally less than 2 inches thick (Figure 4*A*). This laminated to thin-bedded characteristic is present everywhere in the member over an outcrop width of nearly 5 miles in places and an outcrop length of more than 30 miles. The overlying Ramseyburg Member comprises about 25 percent graywacke. The slates in the graywacke are thin bedded at the base and become thicker bedded upwards. The first thousand feet or so of the Pen Argyl Member, immediately overlying the Ramseyburg, is well exposed in a belt of quarries in the Wind Gap and Bangor area, and is characterized by thick-bedded slates, some of which are more than 20 feet thick (Figure 4*B*). In 1967 the Field Conference visited a quarry in the Pen Argyl, but most of these quarries are now inactive and flooded. This is the area from which most slate for pool tables is mined. The thick-bedded material is not repeated south of the Ramseyburg outcrop belt, a fact long known to the slate quarrymen of the area. They recognized the difference between the Pen Argyl and Bushkill belts, and named them the “soft slate” and “hard slate” belts, respectively.

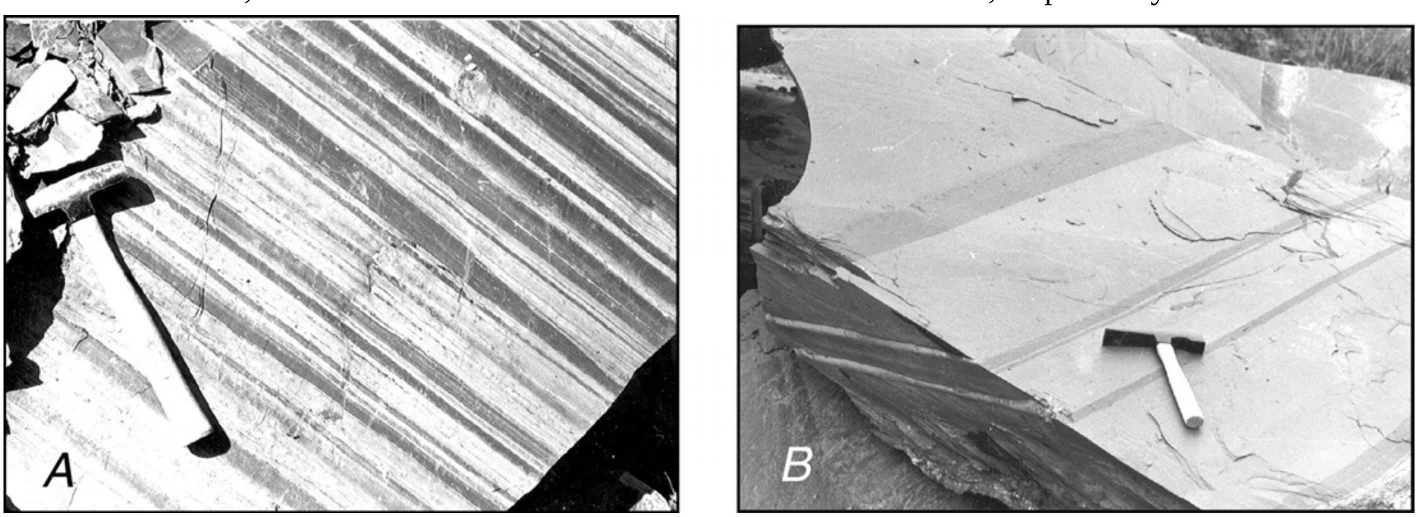



Figure 4. Typical exposures of thin-bedded "hard slate" (*A*, Chapman quarries, 20 miles southwest of Delaware Water Gap) and the much thicker bedded "soft slate" (*B*, Penn Big Bed Quarry, 19 miles WSW of Delaware Water Gap). Some of the beds in the Penn Big Bed quarry exceed 20 feet in thickness (Lash et al., 1984, Stop 9) making it appropriate for such uses as billiard table tops.

The patterns of graptolite distribution of Wright et al. (1979) are used as evidence that the upper and lower Martinsburg members are the same age. An alternate explanation is that graptolites suffer from facies control just as do all paleontologic groups, and there are recurrent faunas in the two slate members (see Lash et al., 1984, p. 80-81). In any event, the final answer to the question of the number of members in the Martinsburg must await a final verdict based on additional paleontologic studies (see Finney, 1985). At present, the three-member interpretation is favored. A more recent study of graptolites in the Delaware Water Gap area supports this interpretation (Parris and Cruikshank, 1992). The most recent geologic map of Pennsylvania (Berg et al., 1980) avoids the issue by showing the three belts on the map, with the northern and southern belts apparently repeated by folding, but also by showing slate units both above and below the greywacke-bearing Ramseyburg Member in the explanation! Lyttle and Epstein (1987) have depicted the regional relations of the three members of the Martinsburg in eastern Pennsylvania.

## SEDIMENTOLOGICAL HISTORY

An interpretation of the depositional environments and paleogeography of the rocks in the Delaware Water Gap area may be made by study of their sedimentary structures, regional stratigraphic relations, petrographic characteristics, and faunal content, and by comparing these rocks with sediments that are being deposited today. The environments of deposition represented by the rocks in the Field Conference area and the paleogeography from Silurian through early Middle Devonian time is depicted in Figure 5.

Few modern studies have been made of the rocks in the Martinsburg Formation, but it appears that these sediments were deposited in a rapidly subsiding flysch-turbidite basin (Van Houten, 1954) formed during Middle Ordovician continental plate collision. The highland source for the Martinsburg was *Appalachia* to the southeast, and the sediments covered a foundered Cambrian and Ordovician east-facing carbonate bank. Basin deepening actually began during deposition of the muddy carbonate rocks of the underlying Jacksonburg Limestone. The thin-bedded graded sequences of siltstone, siliceous slate, and carbonaceous slate of the Bushkill Member of the Martinsburg are probably distal turbidites and pelagic sediments of a deep-sea submarine plain that were later overrun by thicker turbidites and submarine fan deposits of the Ramseyburg Member. Paleocurrent studies indicate that the turbidites flowed down the regional slope to the northwest and turned longitudinally in a northeast direction along the basin axis (McBride, 1962). The deepest part of the basin appears to be in northeasternmost Pennsylvania. Many thick intervals (possibly more than 100 feet thick) of lenticular packets of coarser graywacke were probably deposited in submarine channels that fed the fans. Many of the turbidites in the Ramseyburg were undoubtedly triggered by seismic events related to Ordovician tectonism in the source area. These events may have become less severe during Pen Argyl time, so that much thicker pelagic muds and silt-shale turbidites were deposited between more widely spaced, coarser grained turbidites. The Martinsburg of eastern Pennsylvania and New Jersey lies begging for a detailed petrologic study to decipher the intricacies of its sedimentological history. The contact between the Pen Argyl and Ramseyburg Members disappears under the Shawangunk just within the confines of Delaware Water Gap National Recreation Area (DEWA) one mile west of Delaware Water Gap (Epstein, 1973). The Pen Argyl does not reappear in New Jersey. Several small slate quarries and prospects in the Ramseyburg Member, all long since abandoned, are found within the DEWA boundaries (Epstein, 1974).

Rapid shallowing of the Ordovician basin was accomplished by deposition of the thick Martinsburg detritus and by tectonic uplift reflecting intense Taconic mountain building, which peaked with emergence of the area during the Late Ordovician. This period of orogenic activity and regional uplift was followed by deposition of a thick clastic wedge, the lowest unit of which consists of coarse terrestrial deposits of the Shawangunk Formation. The contact between the Shawangunk and Martinsburg is a regional angular unconformity. The discordance in dip is not more than $15^{o}$ in the area of the Field Conference. Taconic structural relations will be discussed at STOPS 1, 3, and 6 of Day 1.

The conglomeratic sandstone members of the Shawangunk Formation, the Weiders, Minsi, and Tammany (Figure 2) are believed to be fluvial in origin and are interposed by a transitional marine-continental facies (the Lizard Creek Member). The fluvial sediments are characterized by rapid alternations of polymictic conglomerate with quartz pebbles more than 6 inches long, conglomeratic sandstone, and sandstone (cemented with silica to form quartzite), and subordinate siltstone and shale. The bedforms (planar beds, crossbedding, and possibly antidunes) indicate rapid flow conditions. Crossbed trends are generally unidirectional to the northwest. The minor shales and siltstones are thin, and at least one is mudcracked, indicating subaerial exposure. These mudcracks may be seen at the south entrance to Delaware Water Gap on the New Jersey side by looking up about 60 feet at an overhanging ledge (Figure 6). These features indicate that deposition was by steep braided streams with high competency and erratic fluctuations in current flow and channel depth. Rapid runoff was undoubtedly aided by lack of vegetation cover during the Silurian. The finer sediments present are mere relicts of any that may have been deposited in overbank and backwater areas. Most of these were flushed away downstream to be deposited in the marine and transitional environment represented by the Lizard Creek Member of the Shawangunk Formation.

The Lizard Creek Member contains a variety of rock types and a quantity of sedimentary structures that suggest that the streams represented by the other members of the Shawangunk flowed into a complex transitional (continental-marine) environment, including tidal flats, tidal channels, barrier bars and beaches, estuaries, and shallow neritic shelves. These are generally highly agitated environments, and many structures, including flaser bedding (ripple lensing), uneven bedding, rapid alternations of grain size, and deformed and reworked rock fragments and fossils support this interpretation (Epstein et al., 1974). The occurrence of collophane, siderite, and chlorite nodules, and Lingula fragments indicate nearshore marine deposition. Many of the sandstones in the Lizard Creek are supermature, laminated, rippled, and contain heavy minerals concentrated in laminae. These are believed to be beach or bar deposits associated with the tidal flats. The outcrop pattern of the Shawangunk Formation and the coarseness of some of the sediments, suggest that they were deposited on a coastal plain of alluviation with a linear source to the southeast and a marine basin to the northwest (Figure 5). Erosion of the source area was intense and the climate, based on study of the mineralogy of the rocks, was warm and at least semi-arid (Epstein and Epstein, 1972). The source was composed predominantly of sedimentary and low-grade metamorphic rocks with exceptionally abundant quartz veins and small local areas of gneiss and granite. As the source highlands were eroded, the braided streams of the Shawangunk gave way to gentler streams of the Bloomsburg Red Beds.

The rocks in the Bloomsburg are in well to poorly defined, upward-fining cycles that are characteristic of meandering streams (Allen, 1965). These cycles may be seen at STOP 2, Day 1. The cycles are as much as 13 feet thick and ideally consist of a basal crossbedded to planar-bedded sandstone that truncates finer rocks below. These sandstones were deposited in stream channels and point bars through lateral accretion as the stream meandered. Red shale clasts, as much as 3 inches long were derived from caving of surrounding mud banks. These grade up into laminated finer sandstone and siltstone with small-scale ripples indicating decreasing flow conditions. These are interpreted as levee and crevasse-splay deposits. Next are finer overbank and floodplain deposits containing irregular carbonate concretions. Burrowing suggests a low-energy tranquil environment; mudcracks indicate periods of desiccation. The concretions are probably caliche precipitated by evaporation at the surface. Fish scales in a few beds suggest possible minor marine transgressions onto the low-lying fluvial plains (Epstein, 1971). The source for the Bloomsburg differed from that of the Shawangunk because the red beds required the presence of iron-rich minerals (Miller and Folk, 1955), suggesting an igneous or metamorphic source. Evidently, the source area was eroded down into deeper Precambrian rocks.

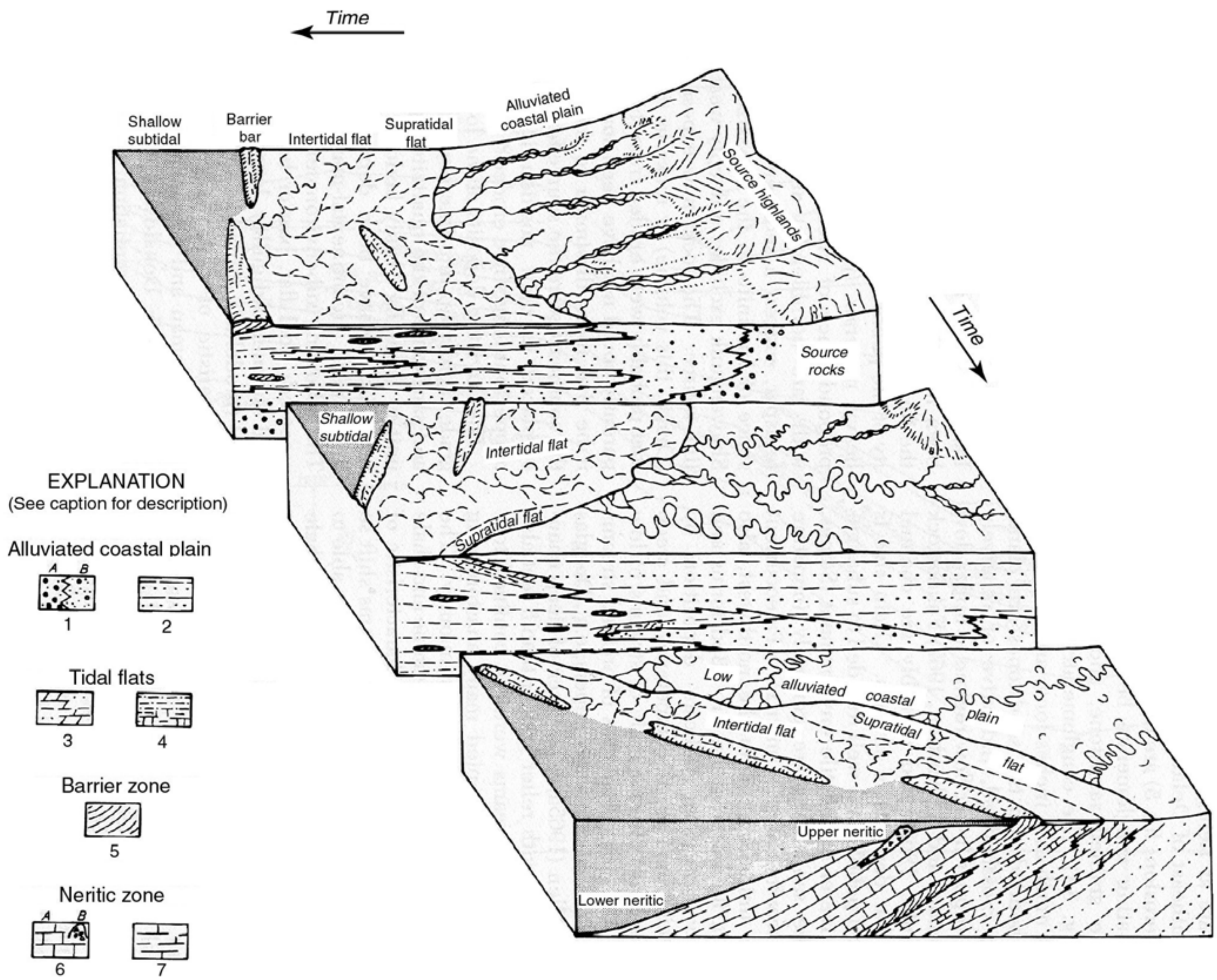



**Figure 5. Generalized block diagram showing sedimentary environments and major lithofacies in northeasternmost Pennsylvania from Silurian through Early Devonian time. Modified from Epstein and Epstein (1969).**

In general, the environments and facies are younger towards the bottom and to the left of the diagram as indicated by the arrows showing time. For example, the upward succession of deposits and environments during Shawangunk time are represented in the uppermost block going from right to left. Taconic orogenesis uplifted the highlands shown on the right, which were the source for the coarse braided stream sediments of the Weiders Member of the Shawangunk Formation. As we go to the left, the slightly finer clastic sediments of the Minsi Member are shown on the alluviated coastal plain. The Minsi overlies the Weiders in eastern Pennsylvania, and the diagram demonstrates the stratigraphic axiom that a vertical rock succession at one place indicates lateral variations over an area. Continuing to the left (and upwards in the stratigraphic succession into younger rocks), we see a variety of transitional deposits typical of the Lizard Creek Member of the Shawangunk Formation (shallow subtidal to supratidal flat). In a similar way, the diagram depicts the changing environments from terrestrial fluvial of the Shawangunk in the upper block, through various tidal and neritic environments of the Silurian and Devonian rocks in the lower block.

*Alluviated coastal plain:*

1. Streams of high gradient, coarse load, low sinuosity (braided).

   A. Bedforms in upper flow regime (planar beds, possible antidunes) and lower flow regime (dunes). Chiefly conglomerate and sandstone. Weiders Member of the Shawangunk Formatlon.

   B. Bedforms in lower upper flow regime (planar beds) and upper lower flow regime (dunes). Chiefly conglomeratic quartzite and quartzite. Minsi Member of the Shawangunk Formation, Lizard Creek Member of the Shawangunk Fm.

2. Streams of low gradient, medium load, and fine floodplain deposits, high sinuosity (meandering). Bedforms in lower flow regime (dunes and ripples). Sandstone, siltstone, and shale. Bloomsburg Red Beds and possibly Decker Formation and Andreas Red Beds in Lehigh Gap area.

*Tidal flats:*

3. Supratidal flat, may include tidal creeks. Dolomite, limestone, shale, sandstone. Laminated (algal), massive, mud cracked, intraclasts, sparse fauna. Lizard Creek Member of the Shawangunk Formation, Poxono Island Formation, Decker Formation, Rondout Formation.
4. Intertidal flat, may include tidal channel and gully, estuary, lagoon, beach. Shale, siltstone, sandstone, and limestone in areas of low terrigenous influx, minor nodules and oolites of collophane, siderite, and chlorite. Irregularly bedded and laminated, graded, rippled, flaser-bedded, cut-and-fill, ball-and-pillow structure, burrowed, restricted fauna (abundant leperditiid ostracodes in carbonates; Lingula and eurypterids in noncarbonates). Lizard Creek Member of the Shawangunk Formation, Poxono Island Formation, Bossardville Limestone, Decker Formation.

*Barrier zone:*

5. Offshore bar and beach. Sandstone, siltstone, and conglomerate. Foreshore laminations, cross-bedding, some burrowing, scouring, wave-tossed shell debris, some textural maturity. Lizard Creek Member of the Shawangunk Formation, Decker Formation, Stormville Member of the Coeymans Formation, Ridgeley Sandstone, Palmerton Sandstone.

*Neritic zone:*

6. Cherty calcareous shale and siltstone, laminated to unevenly bedded, partly burrowed, diverse fauna. Decker Formation, Stormville Member of the Coeymans Formation, New Scotland Formation, Minisink Limestone, Port Ewen Shale, Shriver Chert, Esopus Formation, Schoharie Formation, Buttermilk Falls Limestone.

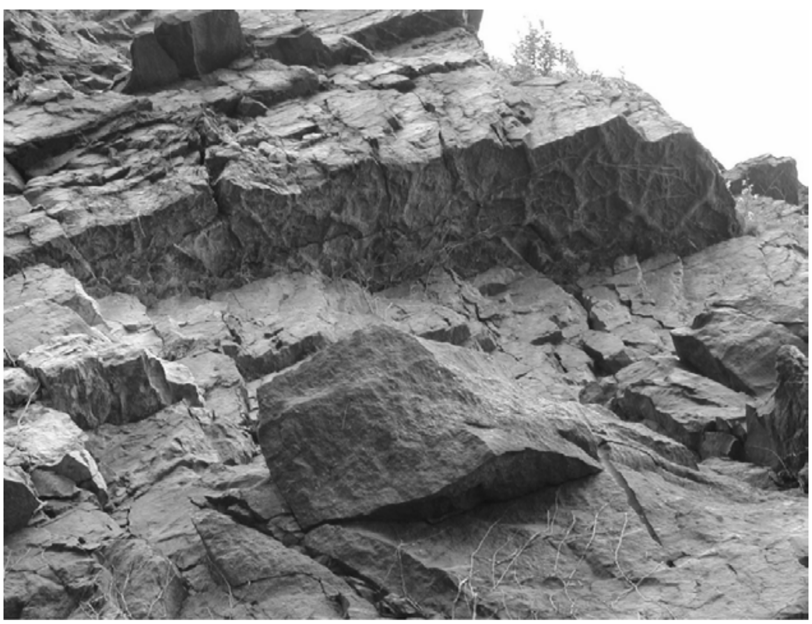


Figure 6. Mudcrack casts at the base of a sandstone bed in the Minsi Member of the Shawangunk Formation, New Jersey side of the Delaware Water Gap. Mudcrack polygons are about 1 foot across.

From Poxono Island time through Oriskany time, the fluvial deposits of the Bloomsburg gave way to transgression of a shallow marine shelf. The area was maintained near sea level and a complex of alternating supratidal and intertidal flats, barrier bars, and subtidal zones was maintained (Figure 5). Sediments indicative of supratidal flats contain laminations of probable algal origin, fine-grained laminated to very thin-bedded massive dolomite and limestone, very restricted fauna (mainly leperditiid ostracodes) or no fossils at all, and mudcracks (Figure 7). Supratidal sediments are found in the Poxono Island Formation, Bossardville Limestone, Decker Formation, and Rondout Formation. Intertidal flat sediments are characterized by graded, laminated, and thin-bedded partly quartzose limestone, cut-and-fill structures, small-scale crossbedding, intraclasts (edgewise conglomerates), abundant leperditiid ostracodes, storm-tossed shell debris, and some mudcracks. Intertidal flat sediments are found in the Poxono Island Formation, Bossardville Limestone, Decker Formation, and Rondout Formation. Barrier-bar and beach deposits are distinguished by calcareous sandstone, conglomerate, and quartzose limestone with foreshore laminations and crossbedding, cut-and-fill structures, intraclasts, skeletal debris of a variety of marine organisms, and scattered burrows. These deposits are common in the Decker Formation, Ridgeley Sandstone of the

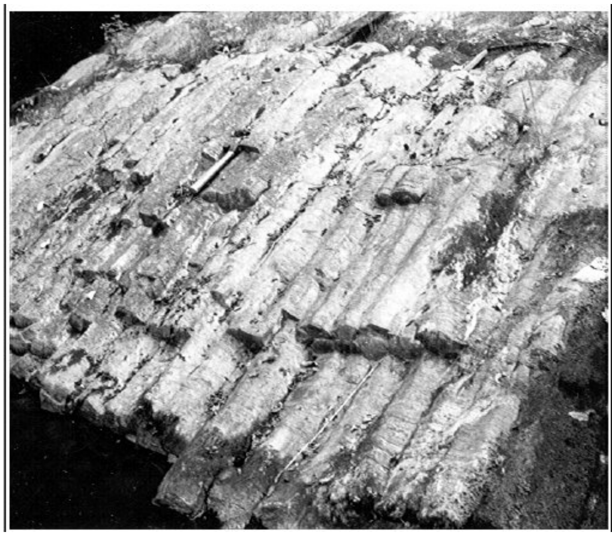


Figure 7. Mudcrack columns, about 13 feet deep, in the Bossardville Limestone along creek in Shawnee-on-Delaware.

Oriskany Group, and Coeymans Formation. Neritic deposits consist predominantly of calcareous shale and limestone that may contain abundant chert. Fauna are diverse and abundant, and burrowing may be extensive. Reefs developed locally in the Shawnee Island Member (and equivalent strata) of the Coeymans Formation in both Pennsylvania and New Jersey (STOP 5, Day 1). Neritic units include the Decker Formation, Coeymans Formation, New Scotland Formation, Minisink Limestone, Port Ewen Shale, and Shriver Chert of the Oriskany Group.

The wandering shoreline during deposition of the Poxono Island through the Oriskany migrated northwestward and the area became emergent following Oriskany deposition. Next came a rapid change to moderate to deep neritic conditions during deposition of the Esopus and lower Schoharie Formations. These rocks are characterized by persistence over a wide geographic area (eastern Pennsylvania to east-central New York), lack of abundant skeletal debris, abundant hexactinellid sponge spicules, and abundant *Taonurus,* a trace fossil typical of the *Zoophycus* facies of Seilacher (1967). A regressive phase followed from Schoharie into Onondaga (Buttermilk Falls) time as indicated by an upward transition from horizontal to vertical burrows, an increase in marine fauna (including corals), and an increase in limestone (STOPS 7 and 8, Day 2; Ver Straeten, this guidebook p. 35). These features indicate water depths within the photic zone and warm, well-oxygenated, and gently circulating water. The Palmerton Sandstone, found about 10 miles southwest of DEWA, lying between the Schoharie Formation and Onondaga Limestone, was most likely a marginal marine (bar or beach) linear sand body. It is massive and generally lacks distinctive internal structures, making a precise interpretation of its depositional environment a bit uncertain.

The black pyritic shales of the Marcellus Shale, with its depauperate fauna, reflects development of an anoxic basin below wave base in the deeper part of a prodelta plain, heralding the arrival of the regressive deposits of the Catskill delta. The Tioga ash-bed zone, a series of altered volcanic tuffs (Dennison, 1969; Smith and Way, 1983) occurs in the upper Onondaga of this area (and extends up into the Marcellus Shale to the west). The Tioga B bed at the base of the Seneca Member of the Onondaga marks the top of the Onesquethaw Stage (STOP 7, Day 1). The ash beds record a period of volcanism that presages the onset of the Acadian orogeny later in the Devonian (see Ver Straeten, this guidebook, p. 35).

The overlying Mahantango Formation, which contains coarser siltstones than the Marcellus, and very diverse fauna (brachiopods, corals, bivalves, bryozoans, trilobites, etc.), indicates a shallower marine environment with better circulation (STOP 10, Day 2—though the shale and siltstone at this spot are not particularly fossiliferous!). Local biostromes containing abundant corals attest to the return of more “normal” marine conditions. One of these biostromes, the so-called “Centerfield coral reef” (actually a coral biostrome), is well exposed in several places outside the DEWA, providing excellent fossil collecting for amateur paleontologists. These localities are along PA 191, about 2.7 miles north of Stroudsburg in the East Stroudsburg quadrangle (Wilt, this guidebook p. 72); along PA 115, 0.5 miles northwest of Saylorsburg in the Saylorsburg quadrangle (Hoskins et al., 1983); and along I-80, about 2.5 miles west of Stroudsburg in the Stroudsburg quadrangle. The first two localities have a wide shoulder

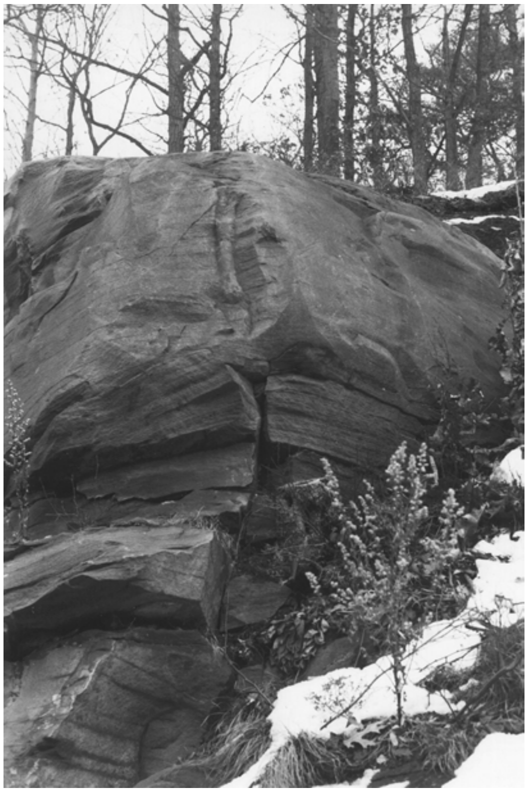


Figure 8.  Burrow of *Archanodon catskillensis* (Vanuxem) in the lower part of the Towamensing Member of the Catskill Formation at Hawks Nest on NY 97, 1.3 miles north-northwest of Sparrow Bush, Orange Co., NY.

along the road for parking. The I-80 locality—though spectacular—is unsafe for collecting because of high-speed traffic.

The Trimmers Rock Formation contains many features suggesting deposition from turbidity currents, including graded sequences, scoured bases, transported fossil hash, and sole marks. These pro-delta slope deposits are transitional up into sediments of the Catskill Formation, and were probably deposited in the pro-deltaic apron in front of the advancing Catskill delta.

The Catskill delta advanced northwestward and the shoreline shifted in response to tectonic uplift and sinking, and to fluctuating loci of deposition.  Fine sandstones of the Towamensing Member of the Catskill Formation (lowest member) are rippled, partly burrowed, and contain plant fragments, as well as burrows of the clam *Archanodon* (Figure 8; Sevon et al., 1989).  These sediments grade up from the Trimmers Rock Formation and are shallower delta front sandstones of the advancing delta, reworked by ocean currents after being transported to the site of deposition along distributary channels.  They may also be partly fluvial in origin.  The overlying Walcksville Member contains red beds that have features similar to the older Bloomsburg Red Beds—crossbedded sandstones with scoured bases, mudcracks, carbonate concretions, and upward-fining cycles, as well as roots—suggesting that a low-lying subaerial fluvial plain of the Catskill delta prograded over the underlying transitional sediments.  Subsidence and marine incursion followed with deposition of fossil-bearing siltstones, shales, and sandstones of the superjacent Beaverdam Run Member.  Conditions similar to those of the Walcksville returned during Long Run time, followed by a thick and coarser sequence of sandstones and conglomerates, suggesting that during upper Catskill time the area was overwhelmed by braided stream deposits as Acadian orogenic uplift to the southeast raised a linear mountain chain similar to the source area that supplied earlier sediments during Shawangunk time.

More sediments were deposited during the later Paleozoic, but the younger rocks formed from these sediments have been eroded away from the Delaware Water Gap area and the nearby Pocono Plateau. The sedimentological record begins again with the deposition of glacial deposits and alluvium during the Pleistocene and Holocene—but that's a wholly different story covered extensively elsewhere in this guidebook.

# STRUCTURAL GEOLOGY OF THE DELAWARE WATER GAP NATIONAL RECREATION AREA

by
Jack B. Epstein

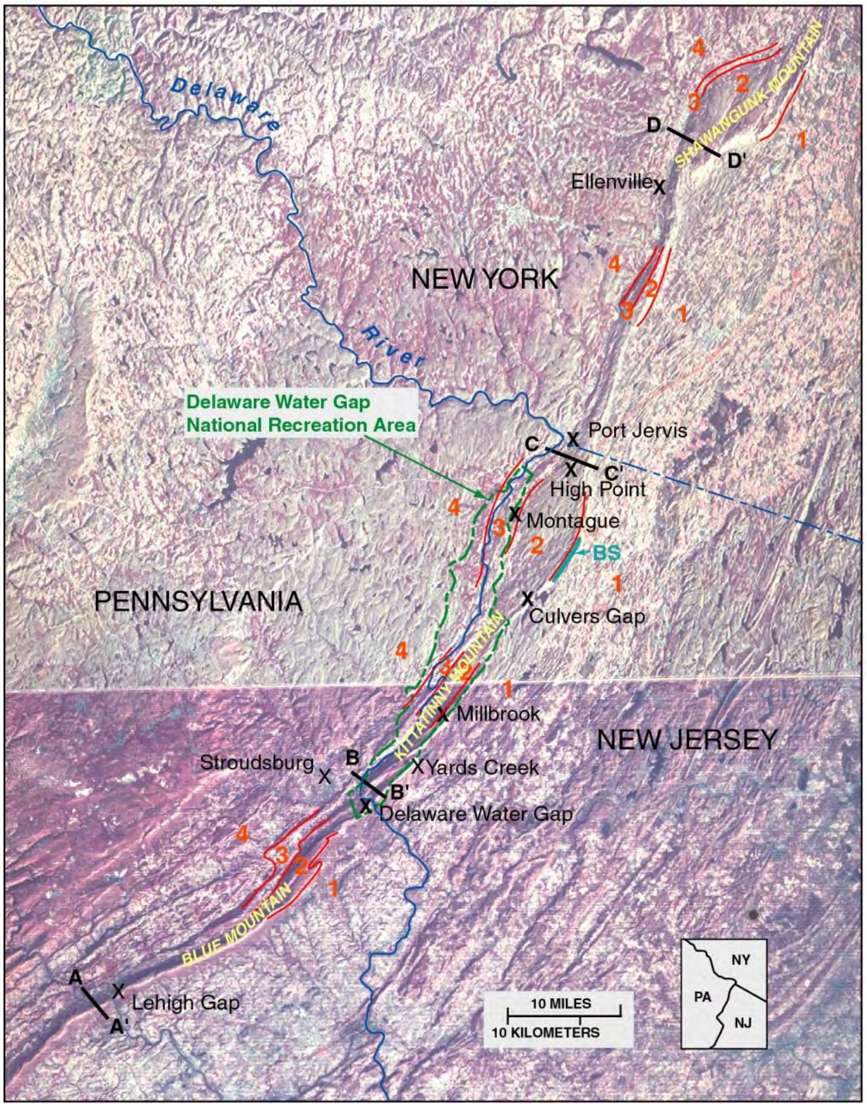



Figure 9. Satellite image of the Delaware Water Gap National Recreation Area showing lithotectonic units (in red) and locations of cross sections shown in Figure 10. BS is the nepheline syenite intrusion near Beemerville, NJ. The exposed syenite is exposed at the north end; the remainder is buried. Its significance is discussed at STOP 6, High Point, NJ.

Field mapping in rocks of Ordovician to Devonian age in the Valley and Ridge province of eastern Pennsylvania and northwesternmost New Jersey indicates that rocks of differing lithology and competency have different styles of deformation. Folding is thus disharmonic. Four rock sequences, lithotectonic units, have been recognized. Each sequence is presumably set off from those above and below by decollements (detachments along a basal shearing plane or zone). Type and amplitude of folds are controlled by lithic variations within each lithotectonic unit. The lithotectonic units, their lithologies, thicknesses, and styles of deformation are listed in Epstein and Epstein, (1969, Table 3) and their distribution is shown in Figure 9. In general the intensity of folding diminishes to the northeast, from overturned and faulted folds in the southwest to northwest-dipping monoclines with superimposed gentle folds in the northeast (Figure 10).

Lithotectonic unit 1 comprises the Martinsburg Formation. It will be seen at STOP 3, Day 1. Slaty cleavage is generally well developed in its pelitic rocks. In the interbedded graywackes, a coarser fracture cleavage is refracted to steeper angles. Cleavage tends to die out within hundreds of feet of the contact with the overlying Shawangunk Formation. The significance of this is discussed at STOP 3. The angular discordance at the Martinsburg-Shawangunk contact within the area of the Field Conference is less than $15^{o}$. Bedding and fold patterns in the Martinsburg within several thousand feet of the

Epstein, J. B., 2001, Structural geology of the Delaware Water Gap National Recreation Area, *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. 14 - 21.

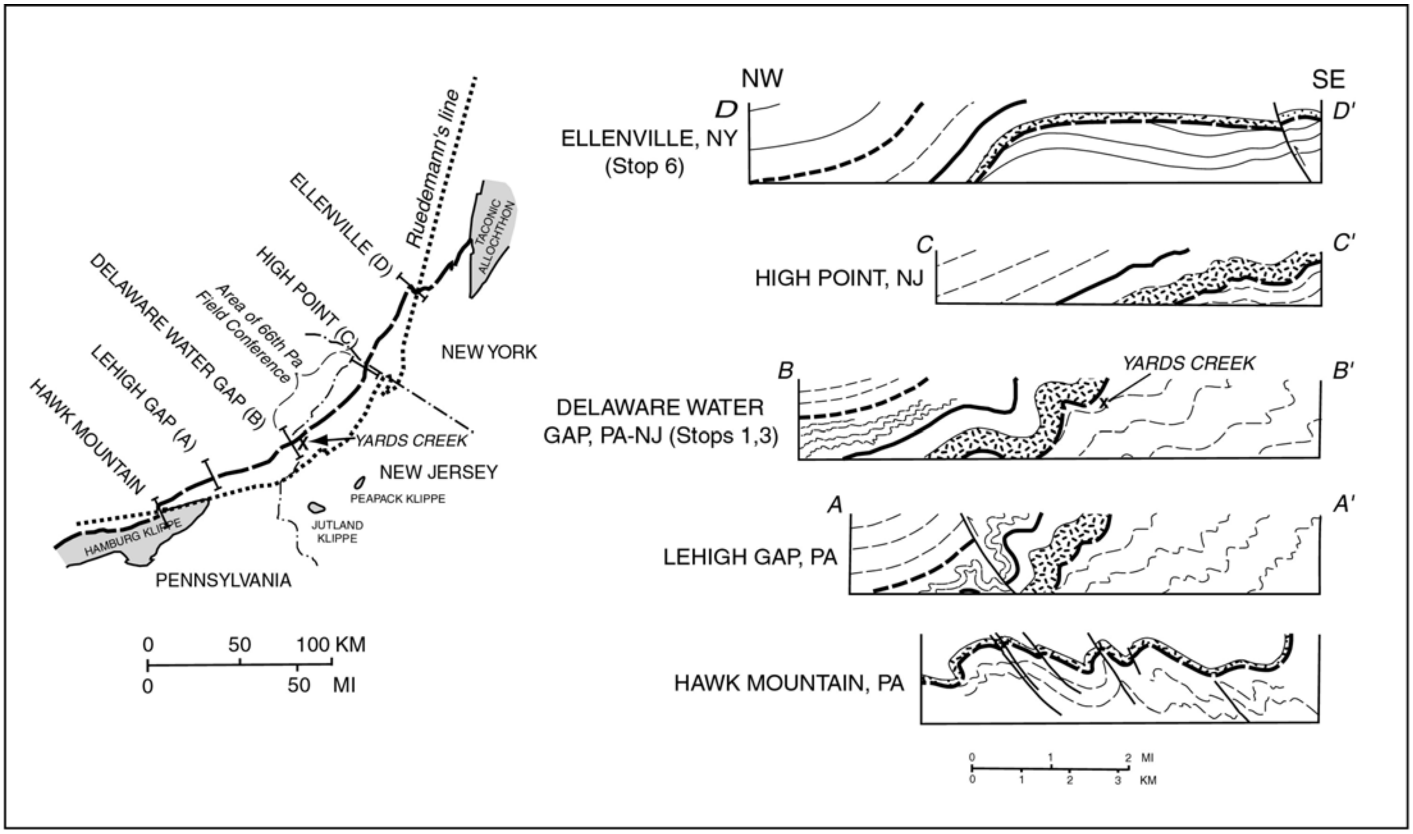
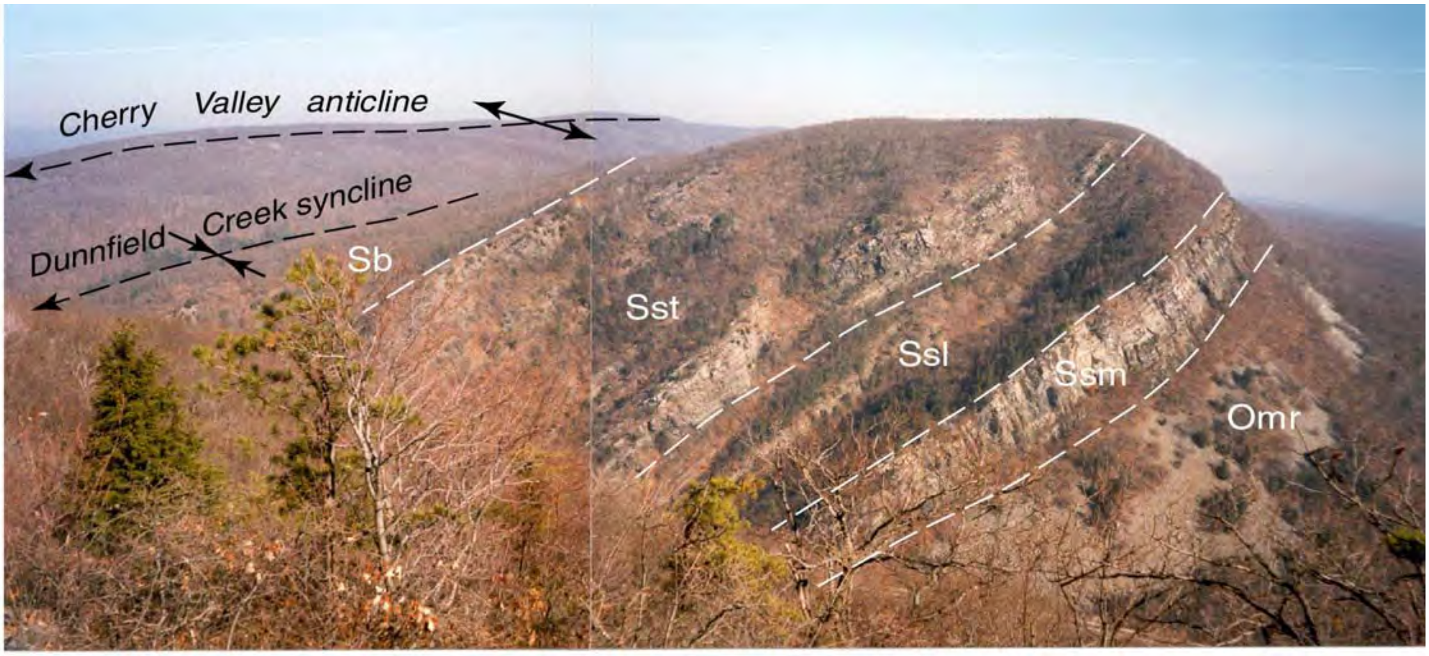



Figure 10. Generalized tectonic map and cross sections along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and New York. Dashed heavy line is the Taconic (Ordovician) unconformity separating the Silurian clastics of lithotectonic unit 2 from the underlying Martinsburg Formation (unit 1). Solid heavy line separates lithotectonic units 2 and 3. Short-dashed heavy line separates lithotectonic units 3 and 4. Dotted heavy line, including Ruedemann's line, separates broad open Taconic folds to the north and west from more intense structures to the south and east. Dotted unit in the cross sections is the Tuscarora Sandstone – Shawangunk Formation. Lettered cross sections also shown in Figure 9. Modified from Epstein and Lyttle, 1987.

Shawangunk mimic those in the younger rocks, suggesting that the folding is Alleghanian in age. At STOP 6, Day 1, we will discuss how Taconic structures may be isolated from those of the Alleghanian. At Delaware Water Gap (STOP 1, Day 1) we will discuss the slight diverge in dip at the contact. The angular discordance in strike is more than $7^{o}$, the Martinsburg beds striking more northerly than those in the Shawangunk, so that the contact between the upper Pen Argyl and middle Ramseyburg Members of the Martinsburg heads under the Shawangunk slightly more than one mile southwest of the Water Gap (Epstein, 1973).

Lithotectonic unit 2 is made up of resistant, competent quartzites and conglomerates of the Shawangunk Formation overlain by finer clastics of the Bloomsburg Red Beds. These underlie Blue and Kittatinny Mountains in Pennsylvania and New Jersey, and Shawangunk Mountain in New York. Concentric folding by slippage along bedding planes is common. Cleavage is found within the shales and siltstones of this unit, but it is not so well developed as in the Martinsburg where slates have been commercially extracted. The reason for this is not because of different time of formation (e.g., Taconic or Alleghanian), but because of slight lithologic differences—the Martinsburg shales were more uniform and of finer grain than those in the Silurian clastic rocks. Folds are generally open and upright (Figure 11), but some limbs are overturned. In the Water Gap, the Bloomsburg is thrown into many small folds in the core of the Dunnfield Creek syncline. These can be seen between mileage 3.4 and 4.3 of the Day-



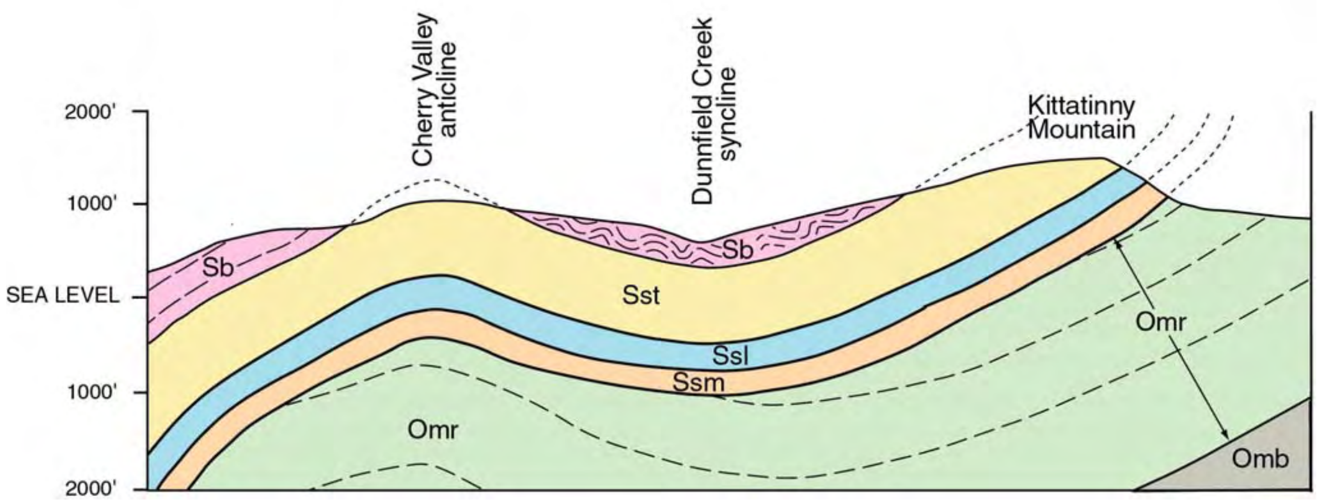




Figure 11. Delaware Water Gap in New Jersey as viewed from atop Kittatinny Mountain (Mt. Minsi) on the Pennsylvania side. Omb, Bushkill Member of the Martinsburg Formation; Omr, Ramseyburg Member of the Martinsburg Formation; Ssm, Minsi Member of the Shawangunk Formation; Ssl, Lizard Creek Member of the Shawangunk Formation; Sst, Tammany Member of the Shawangunk Formation; Sb, Bloomsburg Red Beds. Small-scale folds in the Bloomsburg are located only in the Dunnfield Creek syncline. The angular discordance at the Ss-Om Taconic contact is about one degree (Beerbower, 1956).

1 road log. Cleavage in the Bloomsburg dips to the southeast and appears to have been rotated during later folding. Numerous bedding-plane faults (Figure 12), many with small ramps, in the Bloomsburg contain slickensides with steps that indicate northwest translation of overlying beds, regardless of position within a given fold. Dragging of cleavage along some of these faults indicate that faulting postdated cleavage development, which in turn, predated folding.

The home for rocks of lithotectonic unit 3 is in a narrow ridge (Godfrey, Wallpack) northwest of Kittatinny and Blue Mountains. Folds in this sequence in the southwestern part of the area are of smaller scale than surrounding units (Figure 13 and 14). Axes of these folds are doubly plunging and die out within short distances, making for complex outcrop patterns (Epstein, 1973, 1989). Folding

Figure 12. Bedding thrust and ramp at base of sandstone in the Bloomsburg Red Beds along Old Mine Road in New Jersey.

becomes less intense and in the northeast part of the DEWA where units 2 and 3 dip uniformly to the northwest.

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

Three decollements, or zones of decollement in relatively incompetent rocks, are believed to separate the four lithotectonic units. The Martinsburg-Shawangunk contact is interpreted to be a zone of detachment between lithotectonic units 1 and 2 and can be seen at Yards Creek (STOP 3, Day 1). Thin fault gouge and breccia, about 2 inches thick, are present at the contact. Elsewhere, such as at Lehigh Gap (Epstein and Epstein, 1967, 1969), and at exposures in southeastern New York (Epstein and Lyttle, 1987), thicker fault gouge, bedding-plane slickensides containing microscarps or steps, and drag folds indicate northwest movement of the overlying Shawangunk Conglomerate.

The change in style of deformation between lithotectonic units 2 and 3 takes place in the Poxono Island Formation, but considerable northwest movement is indicated by wedging and bedding slip in the Bloomsburg Red Beds (Figure 12).

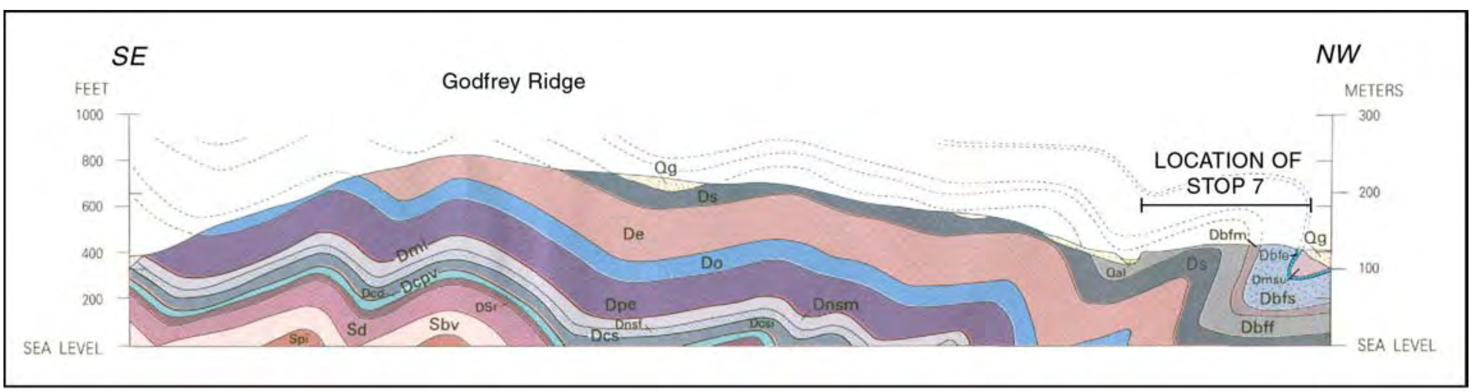



Figure 13. Cross section through Godfrey Ridge showing the overturned anticline in the railroad cut at STOP 7, Day 2, south of I-80 in East Stroudsburg, Pa. Spi, Poxono Island Formation; Sbv, Bossardville Limestone; DSr, Rondout Formation; Dcpv, Peters Valley Member of the Coeymans Formation; Dcd, Depue Limestone Member of the Coeymans Formation; Dcs, Stormville Member of the Coeymans Formation; Dcs, Shawnee Island Member of the Coeymans Formation; Dnsf, Flatbrook Member of the New Scotland Formation; Dnsm, Maskenozha Member of the New Scotland Formation; Dmi, Minisink Limestone; Dpe, Port Ewen Shale; Do, Oriskany Group; De, Esopus Formation; Ds, Schoharie Formation; Dbff, Foxtown Member of the Buttermilk Falls Limestone; Dbfm, McMichael Member of the Buttermilk Falls Limestone; Dbfs, Stroudsburg member of the Buttermilk Falls Limestone; Dbfe, Echo Lake Member of the Buttermilk Falls Limestone; Dmsu, Stony Hollow and Union Springs Members of the Marcellus Shale; Qg, Wisconsinan glacial deposits; Qal, alluvium. Modified from Epstein, 1989.

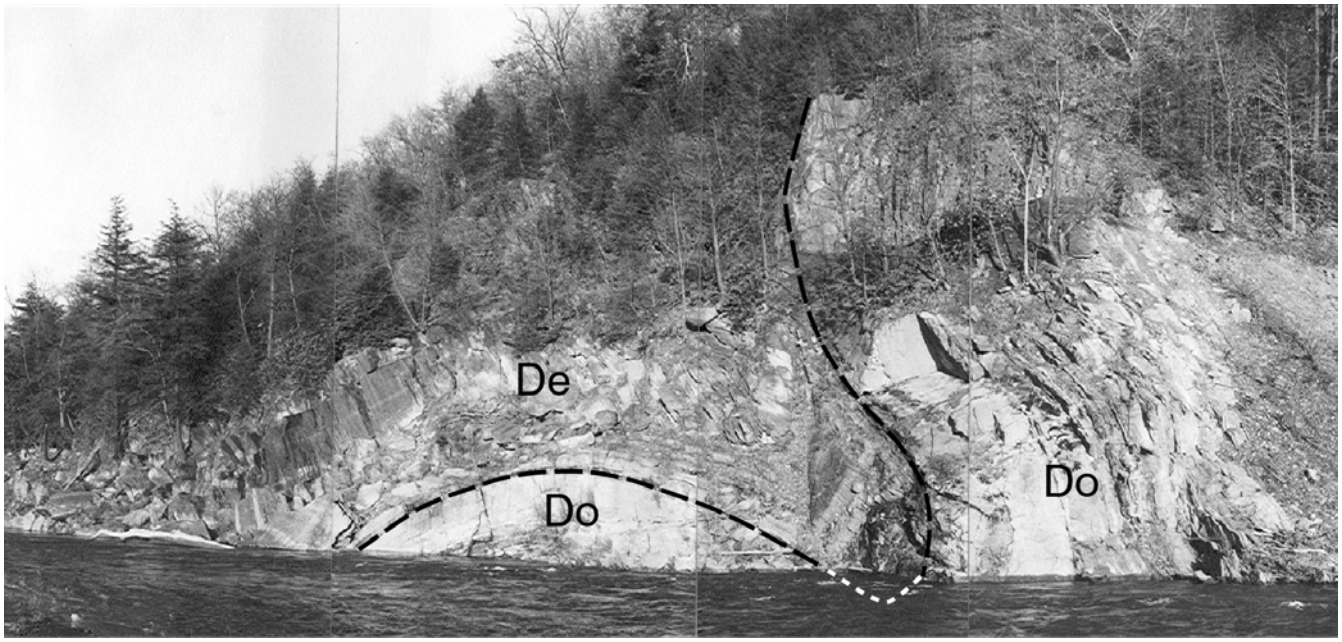



Figure 14. Typical overturned fold in lithotectonic unit 3 along Brodhead Creek, one mile west of the village of Delaware Water Gap and 2000 feet east of STOP 7, Day 2. Do, Oriskany Group, comprising the Ridgeley Sandstone at the top and Shriver Chert in slope to right; De, Esopus Shale. View looking south. Outcrop is the same structure as in the northwesternmost syncline at depth in the cross section in Figure 13.

The movement between lithotectonic units 3 and 4 occurred within the Marcellus Shale. A 10-foot shear zone will be seen in the base of the Marcellus at STOP 8, Day 2. In the Lehigh Gap area, about 30 miles southwest of Delaware Water Gap, the Marcellus is extensively faulted (Epstein et al., 1974) in a zone several hundred feet thick.

The intensity of deformation in the Valley and Ridge in the Field Conference area decreases to the northeast from Pennsylvania, through New Jersey and into New York (Figure 10). Southwest of Delaware Water Gap, as at Lehigh Gap, many of the folds are recumbent and isoclinal, and continued tightening has produced faults in lithotectonic units 1 and 2 because of insufficient space in the cores of anticlines (Epstein et al., 1974). The amplitudes of folds are greater in this area. The Appalachian Mountain section of the Valley and Ridge province is far wider than to the northeast (40 miles wide west of Lehigh Gap compared to not more than 5 miles wide east of Delaware Water Gap), and slaty cleavage is developed to a greater degree in younger and younger rocks. For example, whereas slate has been quarried in the Martinsburg throughout eastern Pennsylvania, is has been quarried in a shaly interval in the Mahantango Formation near Lehigh Gap (Behre, 1933, p. 121).

## A WORD OR TWO ABOUT SLATY CLEAVAGE

Slaty cleavage is the property of a rock that allows it to be split into very thin slabs of slate. It is controlled by parallelism of platy minerals in the rock. For many years geologists did not argue that slaty cleavage was formed during folding, the stress having rearranged the orientation of minerals, particularly micas, parallel to the cleavage direction. It was considered a metamorphic process, occurring during elevated temperature and pressure. Slaty cleavage is well developed in pelitic rocks of the Martinsburg Formation. The Martinsburg has been quarried for slate since it was discovered about 1808.

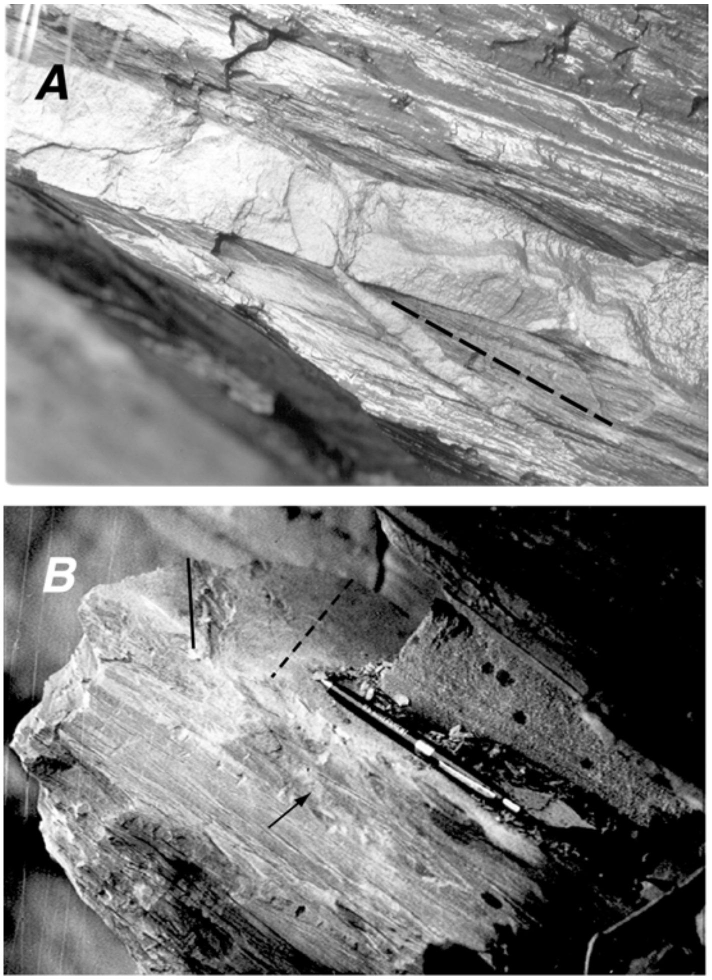



Figure 15. Sandstone dikes in the Martinsburg Formation along US Route 46 south of Delaware Water Gap.
*A.* Dike #1 in Figure 16, same dike as shown by Maxwell (1962, Fig. 4A). Cleavage (dashed line) dips 8° less than the dike. Mud from the overlying bed replaced the evacuated sand and formed a mud dike in the graywacke. A poorly developed cleavage in the graywacke is about 10° steeper than the mud dike.
*B.* Another sandstone dike at this locality (# 2 of Figure 16).

South of Columbia, NJ, the Martinsburg Formation is exposed in continuous outcrops along US 46. Here, about 5 miles south of Delaware Water Gap, is an exposure of interbedded graywacke and slate in the Ramseyburg Member of the Martinsburg. Based on interpretation of a sandstone dike intruded down from a graywacke bed and into the cleavage of the underlying slate, Maxwell (1962) concluded that the slaty cleavage in the Martinsburg Formation in the Delaware Water Gap area was produced by tectonic dewatering during the Taconic orogeny, and the cleavage was the result of only slight stress on pelitic sediments with high porewater pressures. The slate that was produced, therefore, is not a metamorphic rock, but is rather a product of diagenesis. As a consequence, Maxwell concluded that the Taconic orogeny was minor in comparison to the later more intense Alleghanian orogeny, during which time a metamorphic fracture cleavage was produced in the Martinsburg and younger rocks. Maxwell's ideas served the geologic profession very well because they stimulated a flood of papers on the origin of slaty cleavage (a recent search of a the GeoRef geologic data base for articles after 1965 resulted in 450 hits for *slaty cleavage*).

Figure 15*A* is a photo of the dike that Maxwell first discovered that stimulated his interpretation of a nonmetamorphic origin of slaty cleavage. He reasoned that high pore pressures in the sand beds caused the fluid expulsion of sandstone dikes parallel to already-formed slaty cleavage in the water-bearing muds. Note, however, that the dike is not parallel to the slaty cleavage in Figure 15*A*. There are several other dikes extending down from the parent bed (Figure 16). None of these are parallel to the cleavage. They vary considerably in dip, dip direction, and strike. In one case (dike #2, Figure 15*B*) the strike of the dike on the graywacke-bedding surface does not parallel the strike of cleavage on the bedding surface (the intersection of bedding and cleavage; *IBC*). A thin section of one of the dikes (the specimen was loose and about ready to fall when collected in 1970) is shown in Figure 17. Note the lack of parallelism between the dike and slaty cleavage.

Clearly, the supposed parallelism between sandstone dikes and slaty cleavage, which formed the basis for the non-metamorphic origin of cleavage, is incorrect. Field relations also show that variation in cleavage development in the younger rocks is controlled by lithologic differences and not age differences.

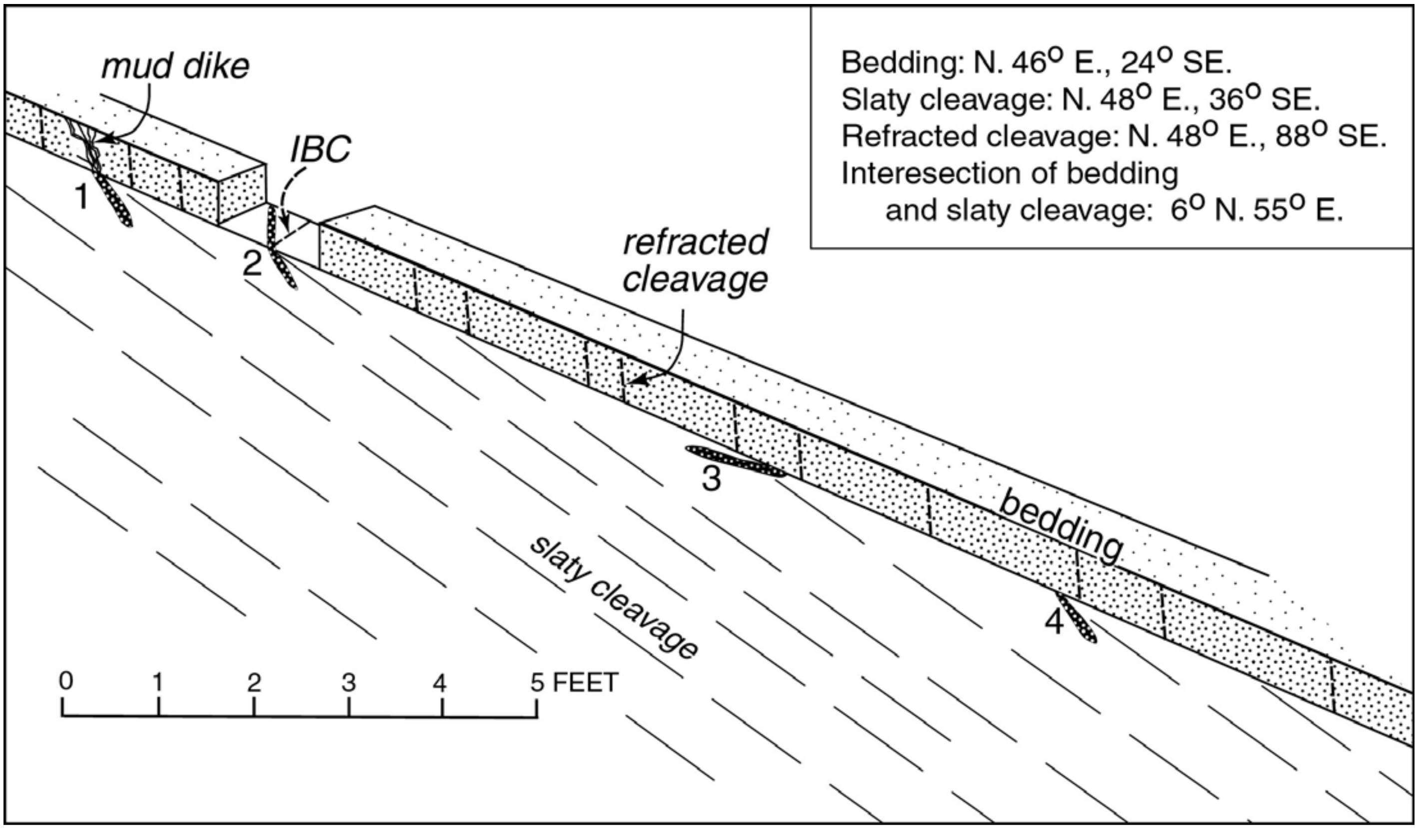



Figure 16. Sandstone dikes extending down from a graywacke bed into slate in the Ramseyburg Member of the Martinsburg Formation, along US 46, five miles south of Delaware Water Gap. North is to the left.

1. Sandstone dike shown in Figure 15*A* and portrayed by Maxwell (1962, p. 287). The dike does not parallel cleavage (dips 8° steeper than cleavage). A mud dike extends into the graywacke bed and dips 10° less than the refracted cleavage.
2. Sandstone dike dips 5° steeper than cleavage and is shown in Figure 15*B*. The strike of the dike (N28°E) is more northerly than the strike of cleavage. This difference is reflected in the divergence of the trend of the intersection of bedding and cleavage (IBC) with the trend of the intersection of the dike and bedding.
3. This sandstone dike differs from the others in that it dips more gently than slaty cleavage. Figure 17 shows the details.
4. The strike of this dike is also more northerly than the strike of cleavage (N25°E) and it dips 10° more steeply than cleavage.

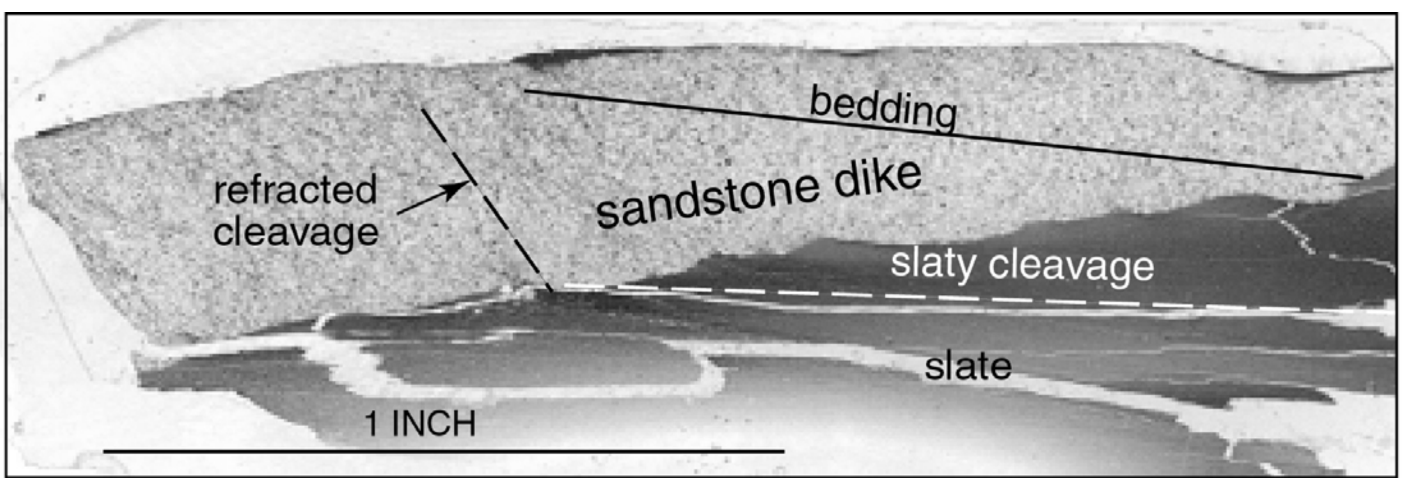



Figure 17. Scanned thin section of sandstone dike # 3 shown in Figure 16. The dike dips to the southeast (to the right) 4° less than bedding and slaty cleavage dips 9° more than the dip of the dike. Irregular fracture at word "slate" is pull apart in thin section.

I have concluded (Epstein and Epstein, 1967; Epstein, 1974) that the dominant northwest-verging folds and related regional slaty cleavage were produced during the Alleghanian orogeny and are superimposed upon Taconic structures in pre-Silurian rocks. The regional slaty cleavage formed after the rocks were indurated at, or just below, conditions of low-grade metamorphism. Estrangement of the effects of the two orogenies is still the subject of considerable debate, but we try a stab at it under *Structural relations along the Taconic unconformity between New York, New Jersey, and Pennsylvania* (Epstein and Lyttle, this guidebook, p. 22). Some of the thoughts and data used to reach this conclusion are listed here, without going into great detail:

(1) An Alleghanian age for the regional slaty cleavage is supported by $^{40}Ar/^{39}Ar$ whole-rock analysis from the Martinsburg Formation at Lehigh Gap (Wintsch et al., 1996).

(2) The arching of cleavage in different stratigraphic levels and at different places in the Martinsburg Formation as the contact with the overlying Shawangunk is approached (Delaware Water Gap, STOP 1, and Yards Creek, STOP 3, Day 1) demonstrates a post Silurian age for the cleavage as described at STOP 1. It is not due to Alleghanian folding of a Taconic cleavage as suggested by Maxwell (1962) and Drake et al. (1960).

(3) There are many examples of bedding-plane slickensides that are cut by cleavage in the Martinsburg as well as in younger formations. This indicates that the Martinsburg was competent enough to deform by flexural-slip prior to passive deformation and does not support the hypothesis that the cleavage was imposed upon a water-bearing pelite.

(4) The mica in the slate is 2M muscovite as shown by X-ray analyses. This, along with chlorite porphyroblasts, shows that the slate is a product of metamorphism. This is also corroborated by high length-width ratios of quartz grains, the result of pressure-solution.

(5) Slaty cleavage is not confined to the Martinsburg. All post-Ordovician pelitic units contain cleavage. Rocks in the Mahantango Formation have been quarried for slate near Aquashicola, PA, a fact noted many years ago by Dale (1914, p. 108) and Behre (1933, p. 119).

(6) In some exposures of the Martinsburg, a later slip cleavage has nearly obliterated the earlier slaty cleavage. This second cleavage has nearly perfect mineral alignment along which the rock can be split into thin laminae. If transposition had been more complete, a perfectly respectable slate would have resulted as suggested by Broughton (1946, p. 13).

In summary, in easternmost Pennsylvania and northern New Jersey the prominent slaty cleavage in the Martinsburg Formation is not Taconic in age, but formed during the latest Paleozoic deformation at the same time cleavage formed in post-Ordovician rocks. The folds associated with the regional cleavage are Alleghanian. However, with some difficulty, as discussed by Epstein and Lyttle (this guidebook) folds of Taconic age can be resolved from the complex fold package in this part of the Appalachians.

# STRUCTURAL RELATIONS ALONG THE TACONIC UNCONFORMITY BETWEEN NEW YORK, NEW JERSEY, AND PENNSYLVANIA

by

Jack B. Epstein and Peter T. Lyttle

## INTRODUCTION

From atop High Point in New Jersey at STOP 6, Day 1 of the Field Conference, we will be able to peer off 30 miles into New York State and see a broad fold in the Shawangunk Formation, which overlies a wide tract of decently exposed rocks of the Martinsburg Formation. Examination of these exposures by Epstein and Lyttle (1987) yielded the story that follows, a tale of structural zones at the limits of Taconic deformation and the relative effects of Taconic, Acadian, and Alleghanian deformation. We unabashedly plagiarize and modify much of our writings here.

The Ordovician Martinsburg Formation was folded and faulted during the complex deformation of the Taconic Orogeny. Following Taconic deformation, mountains rose to the east and coarse sediments were transported westward, and sandstone and conglomerates of the Shawangunk Formation were deposited across beveled folds of the Martinsburg Formation. A thin diamictite containing exotic pebbles, seen at a couple of localities in southeastern New York, records a heretofore unreported geologic episode which occurred during the Taconic hiatus; this will not be discussed further here. As the mountains were worn down, finer clastic sediments and carbonates were deposited more or less continuously into the Middle Devonian. Clastic influx during the Middle Devonian records a later orogeny, the Acadian. The structural effects of the Acadian orogeny did not extend as far southwest as northern New Jersey and southeastern New York; the limit of Acadian folds, faults, and igneous intrusions lies to the east. Finally, near the end of the Paleozoic, continental collision deformed all rocks, down to and below the Martinsburg. The trends of these later (Alleghanian) structures are more northeasterly than those of the Taconic in southeastern New York.

## STRUCTURAL GEOLOGY

The timing and degree of deformation of both Ordovician and younger rocks in northern New Jersey and southeastern New York in the area of STOP 6 has been the subject of considerable long-standing debate. The four most important questions are: (1) what is the geographic distribution of Taconic structures in pre-Silurian rocks; (2) what are the intensities of Taconic and post-Taconic deformations in pre-Silurian rocks, (3) what is the age of the folds, faults, and cleavage in these pre-Silurian rocks; and (4) is the age of the post-Taconic deformation Acadian or Alleghanian, or both?

Our field work suggested that (1) zones of Taconic deformation can be recognized which decrease in intensity from northwest to southeast; (2) northwest and west of "Ruedemann's line" (Figure 18) the Ordovician rocks were more severely affected by post-Taconic deformation than by Taconic deformation; (3) the age of the later deformation is Alleghanian and not Acadian; (4) Alleghanian deformation decreases in intensity from Pennsylvania, through New Jersey, and into southeastern New York; (5) the regional slaty cleavage in this area, where present, is Alleghanian in age; (6) more intense, later, Alleghanian deformation overlaps the earlier Alleghanian deformation in the eastern part of the area; and (7) the strike of Taconic structures is more northerly (by as much as $20^{o}$) than Alleghanian structures in southeastern New York.

Epstein, J. B. and P. T. Lyttle, 2001, Structural relations along the Taconic unconformity between New York, New Jersey, and 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. 22 - 27.

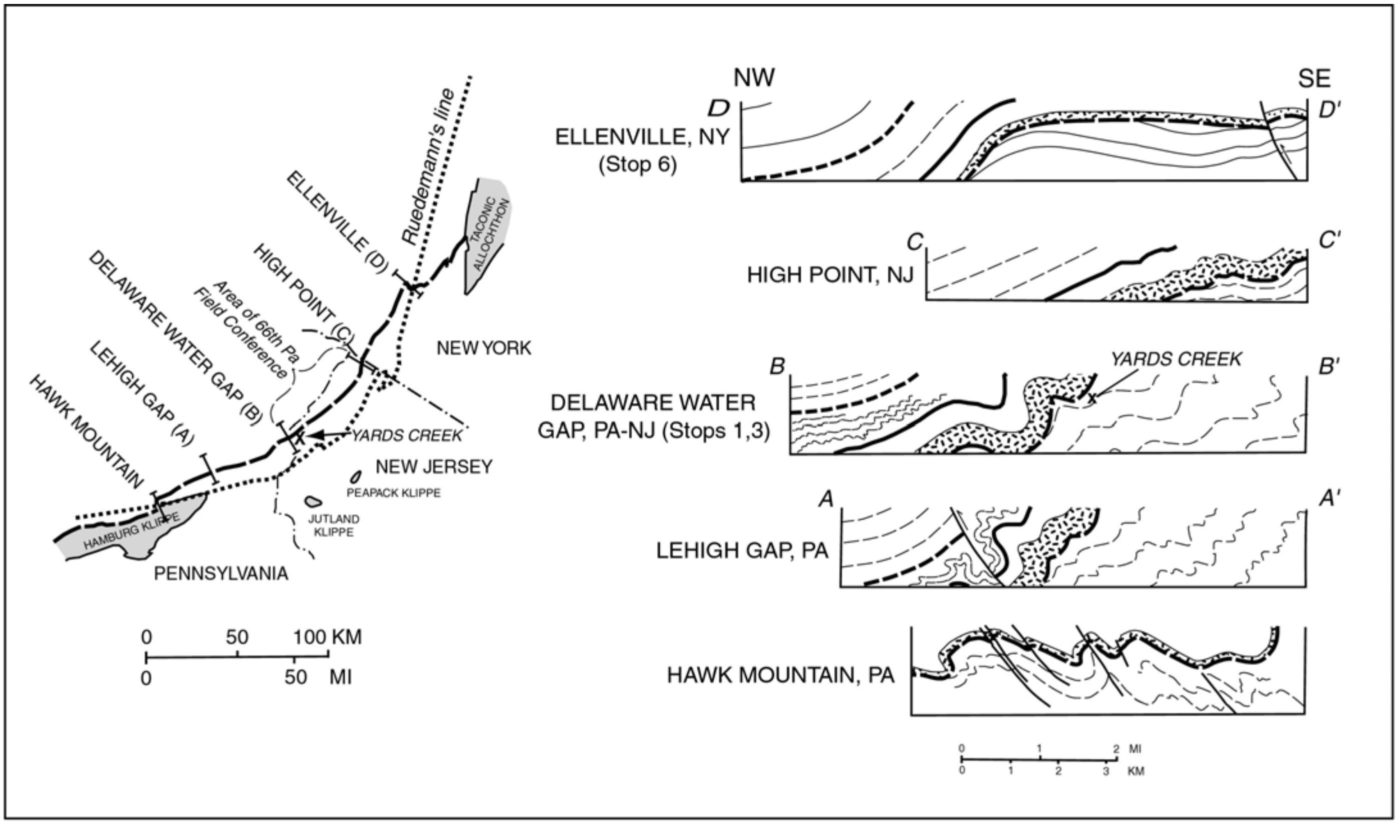



Figure 18. Generalized tectonic map and cross sections along the Taconic unconformity, northeastern Pennsylvania, New Jersey, and New York. Dashed heavy line is the Taconic (Ordovician) unconformity separating the Silurian clastics of lithotectonic unit 2 from the underlying Martinsburg Formation (unit 1). Solid heavy line separates lithotectonic units 2 and 3. Short-dashed heavy line separates lithotectonic units 3 and 4. Dotted heavy line, including Ruedemann's line, separates broad open Taconic folds to the north and west from more intense structures to the south and east. Dotted unit in the cross sections is the Tuscarora Sandstone – Shawangunk Formation. Modified from Epstein and Lyttle, 1987.

## Taconic Tectonic Zones

Epstein and Lyttle (1987) identified three tectonic zones of Taconic age in southeastern New York based on field mapping and compilation of data from geologic archives of the Delaware and Catskill aqueduct tunnels (which supply water to the hordes in New York City). These zones strike more northerly (about N10-20$^{o}$E) than the strike of overlying Silurian rocks and they progressively emerge to the southwest along the contact with the overlying Shawangunk Formation. The structure is more complex to the east.

The zones are, from west to east: (1) broad open folds in slight angular unconformity with the overlying Shawangunk Formation; (2) a belt of less severe folds and faults with bedding in high angularity with overlying Silurian rocks; and (3) thrusts, steep dips, overturned folds, and melange. The melange is definitely Taconic because in places Silurian rocks truncate the scaly cleavage in them. The Taconic thrust faults that produced these melanges are abundant to the southeast of the unconformity and become rarer as the unconformity is approached. The differences between zones 1 and 2 were noted in the Albany, NY, area by Ruedemann (1930) and that contact has been termed "Ruedemann's line" by subsequent workers. It trends southerly and is overlapped by Silurian rocks southwest of Albany (Bosworth and Vollmer, 1981; Epstein and Lyttle, 1987, Figure 9). This line passes under the Catskill Plateau and emerges from beneath the Shawangunk Mountains about 5 miles east of Ellenville (Figure 18). To the east of Ruedemann's line in New York lies the complex structural terrain of the Taconic klippen. In Pennsylvania, Ruedemann's line passes beneath Silurian rocks near Hawk Mountain and complex Taconic structures again appear to the southwest of that (i.e., the Hamburg klippe (Figure 18). Farther west in central Pennsylvania the angular unconformity gives way to a conformable Ordovician-Silurian sequence, and orogenic uplift is reflected only by the Taconic clastic wedge (i.e., Bald Eagle, Juniata, Tuscarora).

Ignoring for the moment all faults and folds of Taconic age, the structure of the Martinsburg belt in eastern Pennsylvania and northern New Jersey can be characterized as a northwest-dipping sequence. The oldest member is always on the south side of the Great Valley and the youngest on the north side. Lyttle and Epstein (1987) show that this monoclinal sequence is actually the north limb of a very broad anticline that involves rocks as far south as the Pennsylvania Piedmont and that this structure is probably Alleghanian in age. Going northeastward into New Jersey the middle member of the Martinsburg is found in the trough of several smaller-scale synclines, but still the very broad and general structure is one of a northwestward-dipping monocline. In southern New York State, the Wallkill Valley has long been recognized as a very broad open anticline (e.g., Offield, 1967). This anticline is highly faulted in places, and many of these faults cut Silurian rocks. We interpret them to be Alleghanian in age.

**Relative Effects of Alleghanian and Taconic Deformation at Ellenville, New York**

The tectonic effects in rocks above and below the Taconic unconformity in the central Appalachians has been the subject of considerable discussion and debate ever since the unconformity was recognized by H. D. Rogers (1838). We have been mapping selected areas along 120 miles of the unconformity from eastern Pennsylvania through New Jersey, and into southeastern New York (Figure 18). We have chosen areas where exposures are abundant enough to be able to determine structural relations in rocks on both sides of the contact. In general, going from Pennsylvania to New York, structures become simpler, from highly faulted and folded at Hawk Mountain, where the Tuscarora Formation rests on both the Martinsburg Formation and rocks of the Hamburg klippe, to overturned and faulted rocks at Lehigh Gap, to oversteepened folds at Delaware Water Gap, and upright to slightly overturned folds at High Point, New Jersey, and finally into a fairly simple arch at Ellenville, New York. Slaty cleavage in both Ordovician and younger rocks is common, particularly in the southwestern part of the study area.

The geology of the area near Ellenville, New York, where Alleghanian and Taconic structures are relatively simple, is an excellent place to distinguish the effects of Taconic and later deformations. The Ellenville arch is a northeast-plunging fold with a half wavelength of 4.2 miles. Folded rocks include the Martinsburg in the Great Valley, the Shawangunk in the Shawangunk Mountains, and rocks of Silurian and Devonian age in the Rondout Valley and Catskill Plateau. The broad arch is prominent in exposed cliffs of the Shawangunk Formation in the Ellenville area. It can be seen on a clear day from the High Point monument, looking northeast. The shales and graywackes of the Martinsburg are fairly well exposed, and they rarely contain slaty cleavage in this area. We are therefore able to draw an accurate cross section showing that the crest of the arch differs in position in the Martinsburg and in the Shawangunk (Figure 19). It is clear that this geometry is the result of the folding of an unconformable sequence. If we unfold the folds in the Shawangunk, we can reconstruct the pre-Alleghanian folds in the Martinsburg, as shown on the bottom of the diagram. Note that the Ellenville arch has been eliminated and we are left with only a broad syncline, Taconic in age.

Epstein and Lyttle (1987, Figure 12) prepared a similar reconstruction by rotating bedding in the Shawangunk back to horizontal using a stereonet and determining the retro-deformed Taconic attitudes in the Martinsburg. Thus, the Alleghanian folding (the Ellenville arch) was eliminated leaving only Taconic structures in the Martinsburg. Results were similar to those shown in Figure 19. These exercises prove that Taconic folds in this area are broad and open, and the Ellenville arch is a later

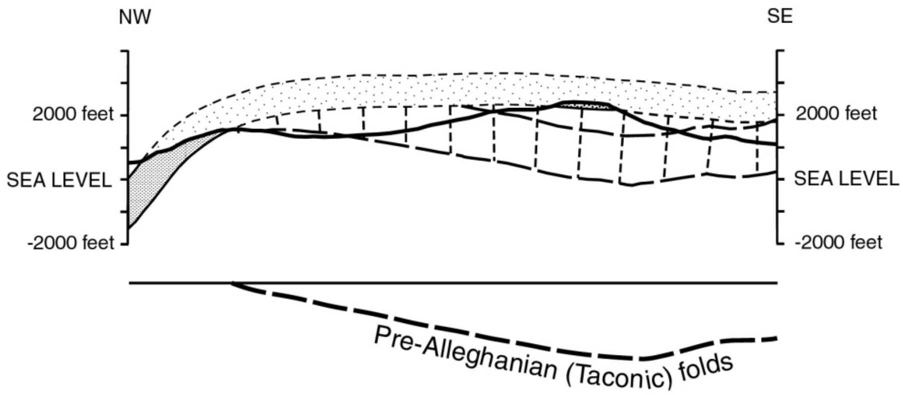



Figure 19. Cross section through the structural arch east of Ellenville, NY, 30 miles northeast of High Point State Park (STOP 6, Day 1). The angular unconformity between the Shawangunk Formation (stippled; light stippled where removed by erosion) and the underlying Martinsburg Formation is accentuated by configuration of key beds in the Martinsburg (long dashed). By measuring the orthogonal distance between the basal Shawangunk and the lower marker bed in the Martinsburg (near vertical short-dashed lines), and by dropping the lines from a horizontal plane, the configuration of Taconic (pre-Alleghanian) folds can be reconstructed, shown in the lower part of the diagram. The location of the cross section is shown in Figure 18. Modified from Epstein and Lyttle, 1987.

structure superimposed on the Taconic folds. Equal area plots of bedding in the Shawangunk and Martinsburg, and in the stereographically rotated Martinsburg defined both the Taconic and Alleghanian folds (Epstein and Lyttle, 1987, Figure 13). The axis of the Alleghanian Ellenville arch plunges 5$^{o}$ toward N32$^{o}$E. The Martinsburg fold trends, as we see them now, are more northerly, by about 10$^{o}$ than trends in the Shawangunk. Interestingly, when the retro-deformed Martinsburg bedding is plotted, the Taconic folds plunge to the southwest. Therefore, we conclude that Taconic folds trend more northerly than Alleghanian folds in this area, and plunge in the opposite direction. Thus, in the Ellenville area, we have been able to distinguish Taconic from Alleghanian folds, both in amplitude and trend.

From the data presented above, and from other considerations (Epstein and Lyttle, 1986; also see discussions at STOPS 1 and 3, Day 1), we draw the following conclusions for the area from Ellenville, New York, to Hawk Mountain, Pennsylvania, and north of Ruedemann's line near the Taconic unconformity:

1. The Shawangunk and equivalent Tuscarora Formation overlie the Martinsburg Formation with an angular unconformity that ranges between an angle that is barely discernible, to about 15$^{o}$.
2. The dominant regional folding in all rocks along the contact is Alleghanian in age.
3. The regional slaty cleavage is Alleghanian in age.
4. Taconic folds in the Martinsburg Formation below the unconformity are mostly broad and open along the entire 120-mile length of the contact that we have studied southwest of Ellenville. To the northeast of Ellenville in zones 2 and 3 the structures become more intense and the angular disparity between beds above and below the unconformity is greater.
5. The strike of Taconic structures trend a bit more northerly (by about 3-20$^{o}$) than later structures. This strike divergence can also be seen in the Delaware Water Gap area.

**Age of Post-Taconic Deformation**

Was the entire sequence of rocks exposed in southeastern New York and adjacent New Jersey affected by Acadian or Alleghanian deformation, or both? Marshak (1986, p. 366) gives a succinct summary of the controversy, which is briefly summarized here. An Acadian age was favored by some workers, based on the age of the youngest rock that has been deformed. Structures in Early Devonian and Upper Silurian rocks were also believed to be Acadian in age by others because these structures were thought to be different in style and trend from structures known to be Alleghanian in age in Pennsylvania. On the other hand, some workers argued that secondary structures, such as joints, could be traced from Pennsylvania into New York, and the fold-fault structures in the Hudson Valley area are Alleghanian in age. An Acadian age was inferred by some from dating of cleavages east of the Hudson River. We favor an Alleghanian age for the following reasons:

1. The Ellenville arch is a structure at the northeast end of a series of structures that extend from tight folds with abundant faults in east-central Pennsylvania, through tight folds with less-abundant faults in easternmost Pennsylvania, through upright folds in New Jersey, and into simple folds and monoclinal dips in southeastern New York (Figure 18). Since these folds in Pennsylvania involve rocks of Pennsylvanian age, the Ellenville arch is therefore believed to be Alleghanian in age. In New York rocks at least as young as the Plattekill Formation of Middle Devonian age are involved in the arch. Possibly even younger rocks, now eroded away, were likewise folded. To the east in New England, the age of Acadian intrusion and deformation is generally believed to be Middle Devonian age (Naylor, 1971). Clearly the Ellenville arch is a post-Acadian structure.

2. The structures of the Hudson-Valley trend in the Silurian and Devonian rocks in the Kingston area (Marshak, 1986) may extend southwest into structures that we have mapped in the Shawangunk Mountains of southeastern New York. We believe that these structures crosscut and post-date the Alleghanian Ellenville arch, and therefore formed during a later Alleghanian event.

3. Many workers have suggested that the youngest rocks that have been faulted are in the Hamilton Group, thus limiting the time of deformation to Middle Devonian (the Acadian orogeny). Epstein and Lyttle (1987) discussed two such fault zones. Slickenlines and verging of folds at these localities indicates northwestward translation along these faults. Similar faults have been reported in equivalent rocks in central New York as much as 100 miles west of Albany (Schneider, 1905; Long, 1922, Rickard, 1952, Bosworth, 1984). Thus, there is evidence for detachment within Middle Devonian shales under the rocks of the Catskill Plateau. Bosworth (1984) suggested that this movement may be linked to detachment in Salina salt under the Appalachian Plateau of central New York and Pennsylvania, described earlier by Prucha (1968) and Frey (1973). Bosworth placed no age constraints on the age of this movement, except to say that it is post-Middle Devonian, and could be Acadian or Alleghanian. If it is linked to the Salina horizon and all the rocks of the Catskill Plateau have moved on this decollement, then an Alleghanian age would be indicated.

Similar fault horizons are found in rocks even higher than the Middle Devonian shale interval. For example, one such fault was discussed by Pedersen et al. (1976, p. B4-B16) in the Plattekill Formation of Middle Devonian age, located along NY 28, 7 miles west of Kingston. The fault zone is a duplex about two feet high in which slickenlines, the verging of folds, and overlapping of structural blocks indicates translation of the overlying beds towards N23$^{o}$W. Well-developed cleavage is found just below the fault. All these data suggest that there has been movement of rocks of the Catskill Plateau above the horizon of the Hamilton Shale as well as within younger rocks. Perhaps many more similar faults zones are waiting to be discovered. If the structures within the Hamilton shales really mark the limit of Acadian deformation, as a number of geologists have suggested, then younger rocks should lie on the Hamilton with angular unconformity. So far as we know, no evidence for such an unconformity has ever been presented. If one recognizes structures such as small thrust zones or detachment horizons within the Hamilton shales, and does not see this sort of structure in any overlying unit, it is meaningless to say that the Hamilton is the youngest unit affected by these structures. There is plenty of evidence to suggest that these structures formed when the rocks were at least partially lithified. Therefore, some rocks younger than the affected beds must have been present and were transported to the west in the overlying block or thrust sheet. Therefore, we feel it is very important to examine the type of structure being discussed when important generalizations about the ages of regional deformations are being made.

4. Lineaments, which have a trend of about N20$^{o}$E, are very apparent on radar imagery and topographic maps. They extend northward into rocks as young as the Plattekill Formation of Middle Devonian age and probably extend into the Oneonta Formation of Late Devonian age. They also parallel faults that we have mapped in the Shawangunk Mountains to the south. In the Catskill Plateau, they are aligned along valleys, which preliminary investigations suggest are controlled by minor faulting and very closely spaced joints. The structures that cause these lineaments are post-Acadian in age, since they cut Upper Devonian rocks. The parallelism with the faults in the Shawangunk Mountains suggests, but does not prove, an age-equivalence.

5. Finally, the Acadian orogeny in New England involved deformation, metamorphism, pluton emplacement, and uplift. Dating of the late orogenic plutons places a minimum date of 380 million years (middle Middle Devonian) for the orogeny (Naylor, 1971). Therefore, Acadian deformation ceased by at least the time that the basal part of the Hamilton Group (Bakoven Shale) was being deposited, if not sooner. Thus, the response in the Field Conference area to Acadian deformation going on to the east was subsidence to form a basin in which Hamilton sediments were deposited. This was followed by shoaling and finally terrestrial deposition (Catskill Formation) as the Acadian mountains to the east were uplifted. Acadian folding may never have extended as far west as the Field Conference area! Faill (1985) likewise suggested that evidence for Acadian deformation of rocks in the Catskill depositional basin is either absent or ambiguous, at best. Catskill sediments are the result of Acadian orogenic uplift, and were not deformed during Acadian tectonism. Faulting in the Plattekill and Hamilton must therefore be the result of later (Alleghanian) deformation. This suggests that the flat-lying and gently dipping rocks of the Catskill Plateau may lie with fault contact on the highly deformed Upper Silurian and lower Middle Devonian rocks of the Hudson Valley. Alternatively, the severe deformation of these Silurian and Devonian rocks may not have extended as far west as the present Catskill front (Marshak, 1986, p. 366).

# STRAIN AND PALEOMAGNETISM IN THE BLOOMSBURG FORMATION AT THE DELAWARE GAP FOLD

by
John A. Stamatakos and Kenneth P. Kodama

## INTRODUCTION

The possibility that grain-scale deformation may alter the orientation of a rock's remanent vectors was first recognized by Graham (1949) who warned of the necessity of considering penetrative deformation as well as rigid body rotations when conducting the paleomagnetic fold test. Kligfield et al. (1981, 1983) and Cogne and Perroud (1985) showed that with increased strain the remanent directions in redbeds of the Alpes Maritimes were rotated progressively away from their Permian directions toward cleavage. Bossart et al. (1990) also noted rotation of the NRM directions toward cleavage in the Murree Formation. Hirt et al. (1986) argued that the great circle distribution of site-mean directions in Permian-Triassic redbeds of the Helvetic nappes was the result of penetrative simple shear associated with nappe-internal deformation.

To test if penetrative deformation has rotated the Bloomsburg's characteristic magnetization, the relationship between the strain geometry and the characteristic remanent directions around the Bloomsburg fold at the Delaware Water Gap in the central Appalachian Valley and Ridge was investigated. At this fold, we measured rock fabric, finite strain, and the remanent magnetization of the Bloomsburg Formation strata.

## GEOLOGIC SETTING AND SAMPLING

The Middle to Upper Silurian Bloomsburg Formation consists of deltaic fluvial sandstones, siltstones, and shales originating from Taconian highlands (Epstein and Epstein, 1969). The Bloomsburg Formation reaches its maximum thickness of 550 meters in eastern Pennsylvania and western New Jersey and can be traced to the northeast into New York and to the southwest into Maryland and West Virginia (Hoskins, 1961). Deformation of the Bloomsburg Formation is thought to be Alleghanian in age (Epstein and Epstein, 1969; Nickelsen and Cotter, 1983), and has resulted in a variety of structural features including asymmetric and concentric flexural folds with extensive bedding-parallel slip, bedding-parallel wedges (Figure 12), and a moderately to well-developed disjunctive cleavage.

Petrographic examinations of the Bloomsburg show that these rocks are composed of clastic fragments of quartz and feldspar as well as chlorite, biotite, muscovite, and opaque minerals including hematite (Epstein et al., 1974). The remanence is carried by hematite (Irving and Opdyke, 1965; Roy et al., 1967; Kent, 1988) and appears in three habits. Irving and Opdyke (1965) noted black crystalline material (20-30 μm in diameter) which they interpreted as specularite. Hematite (1-20 μm in diameter) also appears embedded within the basal planes of detrital chlorite grains (Epstein and Epstein, 1969) and as coatings (0.1 - 1.0 μm in diameter) on the framework quartz grains (Irving and Opdyke, 1965; Epstein et al., 1974).

The anticline at Delaware Water Gap sampled for this study is one of the many third-order broad upright folds located between the Dunnfield Creek Syncline and the Cherry Valley Anticline exposed along the New Jersey section of the Delaware Water Gap (Epstein et al., 1974). Its fold axis trends S72$^{o}$W with an 8$^{o}$ plunge. The medium to fine-grained units contain a moderately to well-developed spaced cleavage that is fanned by the fold and is either normal to bedding or dips to the south with

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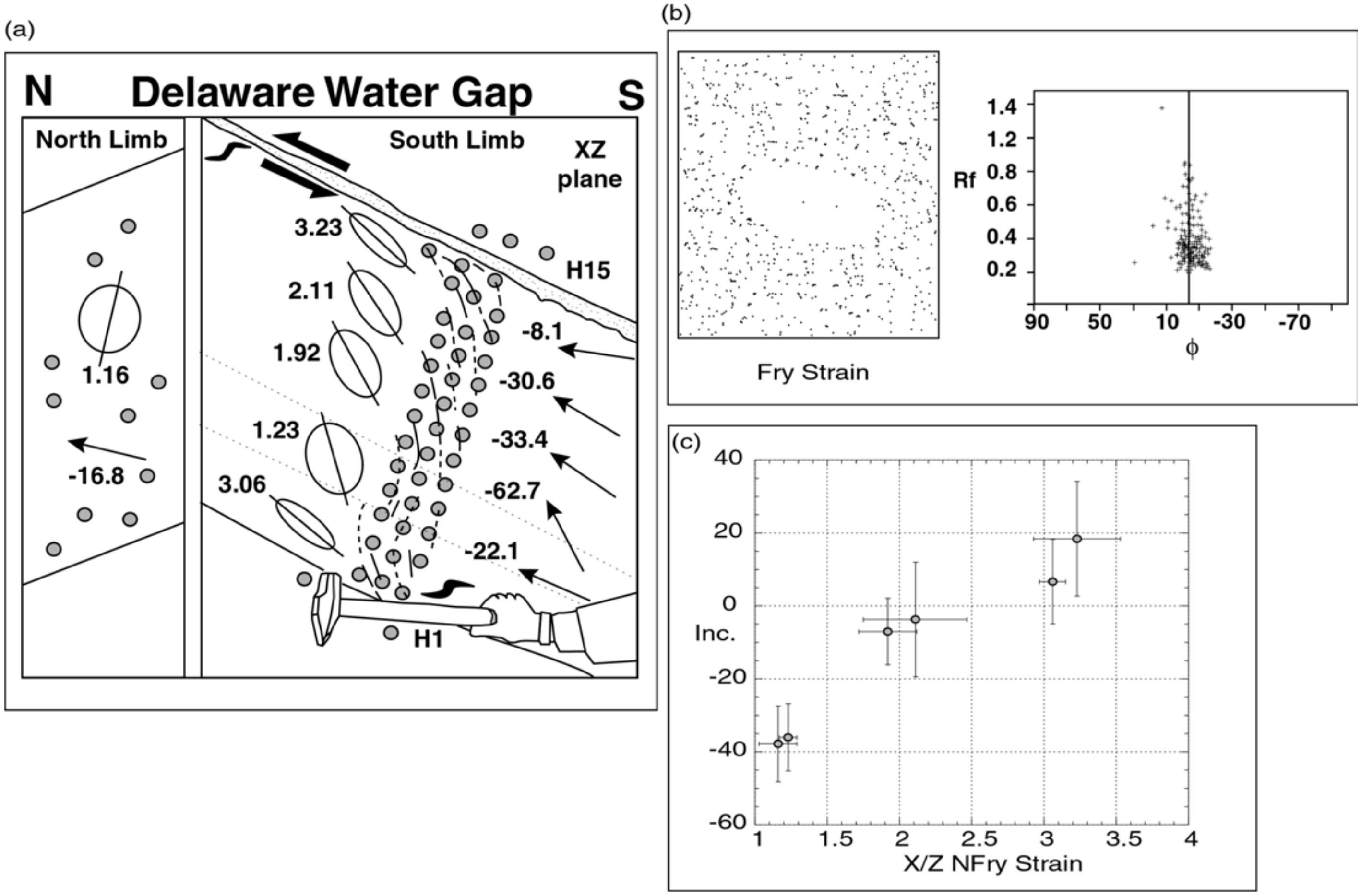


Figure 20. (a) Diagram of the southern limb outcrop face at Delaware Water Gap showing the paleomagnetic and finite strain results. Ellipses with numbers show the magnitude and orientation of the mean finite strains. Black arrows and numbers show the projection of the C-component inclinations, in geographic coordinates, on the XZ principle plane. Dashed lines show the trace of cleavage on the outcrop face. Dotted lines show the approximate location of the 12 cm thick coarser-grained bed within the section. H1 and H15 illustrate the numbering scheme used for the 15 horizons. (b) Representative normalized Fry and $R_f$-$\phi$ diagrams (vertical plane). $R_f$ is the ratio of the long and short axes of the ellipse. $\phi$ is the acute angle between long axis of the ellipse and horizontal, clockwise positive. (c) Delaware Water Gap C component inclinations, in stratigraphic coordinates, plotted against finite strain ratios for normalized Fry results. Inclination error-bars are from the horizon's or site's $\alpha_{95}$. The finite strain error-bars are two standard errors around the mean finite-strain value. Figure modified from Stamatakos and Kodama (1991).

respect to bedding (Figure 20). Bedding-plane faults, delineated by quartz-chlorite shear fibers, can be traced around the folds. Wedges are also common and often appear in conjugate pairs (M. B. Gray, personal communication, 1990). However, most of the bedding plane faults show top-to-the-north (top-to-the-foreland) shear, even those on the north-dipping limb of the anticline.

Structures similar to Nickelsen's (1986) cleavage duplexes are also present. These are 0.5 to 2 m thick zones of concentrated strain bounded above and below by bedding-parallel faults. Within these zones is a strongly to very strongly spaced cleavage oriented at a high angle to bedding near the zone's center but dragged nearly parallel to bedding near the bounding faults producing a sigmoidal cleavage pattern.

Samples for this study come from one of these highly strained zones, located on the south-dipping limb of the anticline. In addition to the well-developed sigmoidal cleavage, this 0.7 m thick zone contains several offset veins near its center and numerous sigmoidal en-echelon tension gashes near its roof and base. Within the lower half of the zone is a 12 cm bed of coarser-grained material that appears less deformed than the rest of the zone. Here cleavage is less pronounced and is oriented at a high angle to bedding. Shear sense for both the offset veins and tension gashes is also top-to-the-north. Strain measurements (Hancock, 1972; Ramsay and Huber, 1983) of six tension gashes yield an average shear strain of $1.0 \pm 0.2$, which corresponds to a strain ellipse with an axial ratio of 2.7. The highly strained zone disappears near the hinge of the fold and either climbs up-section or dies out there. On the northern limb of the anticline, the same bed contains only a weakly to moderately spaced cleavage oriented roughly normal to bedding.

In thin sections, the southern limb samples show patchy undulatory extinction, sutured quartz contacts, overgrowths and curved fibrous beards resulting in preferred elliptical grain-shapes orientated parallel to the trace of cleavage. Detrital chlorite grains are asymmetrically folded and occasionally draped over quartz grains. The orientation of the axial planes of the crenulated chlorite grains are also parallel to the trace of cleavage. The asymmetry of the chlorite microfolds and the curved fibers on the quartz grains suggest top-to-the-north shear. There is also secondary chlorite neocrystallized parallel to cleavage. There is no evidence for a preferred quartz-lattice orientation except for samples nearest the zone's margin, which show moderate lattice alignment when viewed through the gypsum plate. Approximately 10% of the quartz grains appear to have subgrains. Samples from the northern limb of the fold show only a moderately developed spaced cleavage, sutured quartz-grain contacts, and slightly folded chlorite grains.

On the southern limb of the anticline fifteen horizons (designated H1 through H15) consisting of three cores per horizon were collected across a 0.7 meter thick zone of relatively high strain, including two horizons from the beds immediately above and below this zone (Figure 20a). Ten samples were also collected from the same bed on the northern limb of the fold (Figure 20a).

## METHODS

### Paleomagnetism

All paleomagnetic samples were drilled in the field using a portable gasoline-powered drill fitted with a 2.5 cm diameter diamond-coring bit. Each sample was oriented with a standard orienting device and magnetic compass. Cleavage and bedding orientations were made in the field with a Brunton compass with an estimated accuracy of $\pm 3^{o}$.

Progressive thermal demagnetization (12-15 steps) was performed on all samples using a Schonstedt TSD-1 thermal demagnetizer. Remanence measurements were made on a CTF two-axis cryogenic magnetometer at Lehigh University. Characteristic magnetizations were obtained from principal component analysis (Kirschvink, 1980). The distribution of directions about the site-mean and horizon-mean directions were determined using Fisher (1953) statistics. Incremental fold tests were performed by rotating the magnetic directions about the strike of bedding in progressive increments of 5% of bedding dip until the beds were fully restored to horizontal. The statistical significance of the fold test, for each increment of unfolding, was determined by the methods suggested in McFadden and Jones (1981).

### Finite Strain

Finite strain was measured using center-to-center and object strain techniques (Ramsay, 1967; Dunnet, 1969; Fry, 1979) employed in Erslev's (1989) fabric analysis techniques. Oriented thin-sections were cut from paleomagnetic cores or hand samples collected at the outcrops. In each thin-section, the outlines of 120 to 250 quartz grains were digitized. From these digitized grain shapes, two least-squares, best-fit ellipses were calculated, one from the enhanced normalized Fry diagram (Erslev, 1988) and one from an Rf-ϕ plot based on the center-to-center distance between grain centers (Erslev, 1989). The normalization procedure reduces variations due to grain size and sorting by dividing the center-to-center distance between two grains by the sum of their average radii (Erslev, 1988). Enhancement eliminates the subjectivity of manual fit ellipses by calculating the least-squares best-fit ellipse from only those points within the rim of maximum point density (Erslev, 1989). In all the Bloomsburg samples, quartz-grain centers with a center-to-center distance less than 1.10 times the sum of their radii were used to calculate the enhanced normalized Fry ellipses. Object ellipses for each grain were generated by fitting a least-squares ellipse through the grain's ten neighboring grain centers. The ellipticity and orientation of these object ellipses were used to construct the Rf-ϕ diagrams. In order to test the assumption that the initial quartz grain distributions were random, the Rf-ϕ values for two samples from Delaware Water Gap were "unstrained" using Peach and Lisle's (1979) algorithm based on Lisle's (1977) Theta Curve method.

Thin sections were cut perpendicular to the fold axis. On the southern limb of the anticline, finite strain was measured in all three samples from each of five horizons across the high strain zone. A mean normalized Fry and Rf-ϕ strain ellipse was then calculated for the individual horizons from the three individual strain measurements. On the northern limb of the anticline, strain was measured in four samples and mean normalized Fry and Rf-ϕ strains calculate from those four measurements. In order to assess the strain variations within an individual sample, strains were also measured in three different regions of the same thin-section from two samples (NL1 and H14) and compared to the variations observed within a horizon and within the northern limb bed.

## PALEOMAGNETIC RESULTS

Paleomagnetic results were obtained from all ten of the northern limb samples and from fourteen of the fifteen horizons on the southern limb of the fold. Samples from horizon 15, just above the highly strained zone, were too weakly magnetized to yield meaningful results. As previous paleomagnetic studies of the Bloomsburg Formation have shown (Irving and Opdyke, 1965; Roy et al., 1967; Kent, 1988) thermal demagnetization uncovered two magnetic components; a secondary or B component with distributed unblocking temperatures up to 660$^{o}$C and a characteristic or C component with discrete unblocking temperatures between 660$^{o}$C and 690$^{o}$C. On the southern limb, directions from the coarser-grained samples near the center of the highly strained zone and from the horizon below this zone have significantly steeper inclinations than the rest of the samples, especially when compared to sample directions from horizons nearest the roof and floor of the zone. Incremental unfolding of the C component is best clustered at 45% unfolding, which is significantly different from both the prefolding and post-folding configurations at the 95% confidence level.

## FINITE STRAIN RESULTS

Normalized Fry diagrams show moderately to well-defined circular to elliptical vacancy fields (Figure 20b). Rf-ϕ plots are generally symmetrical about the maximum ϕ (Figure 20b). In samples from four horizons (H2, H9, H12 and H14) the Rf-ϕ fields are closed. The Rf-ϕ fields for samples from H5 and from the northern limb of the fold are open and indicate initial axial ratios of between 1.10 and 1.30. Comparison of the Rf-ϕ and normalized Fry results are in good agreement, although the Rf-ϕ results tend to yield slightly higher axial ratios and shallower ellipses relative to bedding. Both the between-sample and within-sample variations are small. Standard deviations for both the ellipticity and the ellipse orientation average less than 10% of the mean values. Both samples "unstrained" by the Peach and Lisle (1979) algorithm exhibited random pretectonic orientations at the 90% confidence level.

The reciprocal strain ratios which produced the lowest $\chi 2$ values were in agreement with the finite strains indicated by the normalized Fry results.

The finite strain results yielded strain ellipses with X/Z axial ratios as low as 1.25 for the northern limb of the anticline and as high as 3.20 for the horizon near the top of the high-strain zone margin (Figure 20a). On the southern limb of the anticline, the orientation of the long axis of each horizon's strain ellipse is roughly parallel to the trace of cleavage on the outcrop face. On the northern limb of the fold, the long axis of the strain ellipse is oriented at a high angle to bedding. Because the C component directions are toward the north the magnetization vectors lie within the XZ principal plane. When the C component inclination is plotted against the X/Z finite strain ratios, inclination varies as a function finite strain. The low strained horizons have shallow to moderately upward inclinations, the moderately strained samples have nearly flat inclinations, and the highest strained samples have shallowly downward inclinations.

## DISCUSSION

### Analysis of Finite Strain

The observed fabric elements suggest that the dominant deformation mechanisms were pressure solution and low temperature plasticity (microfolding of the phyllosilicate grains) associated with the development of cleavage. Pressure solution and microfolding result in a systematic redistribution of quartz-grain centers that is indicative of the added strain (Ramsay and Huber, 1983). This rearrangement of the quartz-grain centers relative to one is largely due to grain-shape modification but may also include grain rotation (e.g. Kerrich and Allison, 1978; Engelder and Marshak, 1985). However, the observation that the quartz c-axes appear to be aligned in the highest strained horizons at Delaware Water Gap (H2 and H14) suggests that dislocation creep and dislocation glide were also important deformation mechanisms in these horizons. Because these mechanisms do not result in a systematic redistribution of grain centers, strain associated with dislocation processes cannot be measured by center-to-center techniques (Ramsay and Huber, 1983).

At Delaware Water Gap, the Rf-$\phi$ and $\chi 2$ results indicate that the initial distribution of quartz-grain centers was uniform and that the initial ellipticity of the grains was small. The long axes of the measured strain ellipses are oriented nearly parallel to the trace of cleavage on the XZ principal plane and strain ratios are higher in samples with the most intense cleavage development. In addition, variations in the magnitude and orientation of the strain ellipse within an individual sample and within a horizon are small, which indicates that finite strain at the scale of the paleomagnetic sample is relatively homogeneous. These results suggest that the measured center-to-center values, barring significant dislocation creep and dislocation glide, are representative of the tectonic strain that deformed these rocks. However, strain at the grain-scale may have been partitioned between progressive shortening of the quartz grains and progressive shearing of the enveloping mica and matrix material (Bell, 1985). If the strain is partitioned in this way, the finite strains indicated by the redistribution of the quartz grain centers may record only a portion of the total strain. Strain of the hematite grains in the matrix and embedded within chlorite grains may have been greater than indicated by the center-to-center values.

### Development of Strain

Both the mesoscopic and microscopic structures at the Delaware Water Gap fold suggests a progressive strain history that includes components of layer-parallel shortening and bedding-parallel shear. A sequence of layer-parallel shortening overprinted by bedding-parallel shear is consistent with current models of the progressive development of Alleghanian structures in the central Appalachian Valley and Ridge (Nickelsen, 1979; Gray and Mitra, 1991). Penetrative bedding-parallel shear appears to be localized within the south-dipping limb. This probably reflects local development of bedding-parallel shear fabrics that were subsequently enhanced and isolated in the south-dipping limb. However this same strain pattern can developed in flexural slip/flow folds which are pinned near the inflection point of the north-dipping limb (Beutner and Diegel, 1985).

**Analysis of Paleomagnetic Directions**

The high unblocking temperatures, above 650$^{o}$C, indicate that C component magnetization is carried by hematite and probably resides in the larger 1-30 micron hematite crystals (Dodson and McClelland-Brown, 1980). The C component directions have north and up directions which is interpreted to be normal polarity (e.g. Van der Voo, 1989). The C component appears synfolding and yields a mean direction at 40% unfolding of D = 355.2$^{o}$, I = -25.2$^{o}$ (k = 14.8) which corresponds to a north paleopole position at 35.6$^{o}$N, 110.0$^{o}$E.

The apparent synfolding nature of the C component can be interpreted as a true synfolding magnetization, a combination of prefolding and postfolding magnetizations, or a strain modified prefolding magnetization. Given the strong correlation between inclination and the magnitude of finite strain (Figure 20c), the most likely explanation is that the C component has been systematically reoriented as a result of the deformation. Hematite's remanence lies within the basal plane (Nagata, 1961) so that any rotation of the basal plane will result in a similar rotation of the remanence vector. Because the C component directions are upward and to the north or downward and to the south, top-to-the-foreland (top-to the-north or northwest), bedding-parallel shear will rotate the basal planes in the same direction as the shear and shallow the inclinations. In the most strained horizons at Delaware Water Gap, the strain has rotated the remanent vectors through the bedding plane, which, in the case of bedding-parallel shear is the shear plane. Rotation of particles through the shear plane suggests that remanence rotation is analogous to rigid particle rotation in a viscously deforming matrix (Jeffrey, 1923).

At the Delaware Water Gap, strain is localized in the south-dipping limb. Unfolding the strained southern limb directions against the relatively undeformed northern limb directions results in an overcorrection of the magnetizations and a synfolding geometry develops from the original prefolding magnetization.

As a test of this interpretation, the fold test at Delaware Water Gap was repeated using the northern limb sample directions and only the sample directions from the southern limb horizons which have relatively low strains (H5 and H6). In this case, the distribution is best clustered at 80% unfolding. However, this distribution is not statistically different from the prefolding configuration at the 95% confidence level (McFadden and Jones, 1981).

Previous interpretations have suggested that the Bloomsburg Formation was remagnetized and partially folded during a phase of deformation in the Late Devonian (Miller and Kent, 1989). There are several observations that argue against Devonian folding and remagnetization. (1) Remagnetization and partial folding in the Devonian would require an angular unconformity between the Lower Devonian and Lower Carboniferous strata in the central Appalachians Valley and Ridge. For example, based on Montour Ridge results, this interpretation would predict an angular unconformity of approximately 12$^{o}$-14$^{o}$ above the Bloomsburg Formation at Montour Ridge. Yet, there is no published sedimentological or structural data to suggest that a Devonian angular unconformity exists in the central Appalachian Valley and Ridge. (2) There is no structural evidence to suggest that the Devonian and Silurian rocks in the Pennsylvanian Valley and Ridge suffered any deformation associated with the Acadian Orogeny. Rather, the Bloomsburg rocks appear to share a common Alleghanian structural history with younger Carboniferous rocks (Gray and Mitra, 1991). (3) The pervasive nature of orogenic remagnetizations (e.g. McCabe and Elmore, 1989; Miller and Kent; 1988a) suggests that if the Bloomsburg was remagnetized in the Devonian, then other pre-Devonian rocks in the region would also exhibit a Devonian remagnetization. However, other pre-Devonian redbeds like the Andreas redbeds (Miller and Kent, 1988b) and the Rose Hill Formation (French and Van der Voo, 1979) do not contain Late Devonian signals. (4) The dual polarity of the Bloomsburg Formation at a similar type fold at Round Top in Maryland appears to be stratigraphically controlled and suggests that the C component has an early diagenetic or detrital origin.

# EVENT AND SEQUENCE STRATIGRAPHY AND A NEW SYNTHESIS OF THE LOWER TO MIDDLE DEVONIAN, EASTERN PENNSYLVANIA AND ADJACENT AREAS

by
Charles A. Ver Straeten

## INTRODUCTION

The act of sedimentation, and the preservation of any given deposit over geologic time is, really, quite the exception. The sedimentary rock record is largely the result of only the most significant events; the continual day-to-day or year-to-year type of processes we see active during our lives merit almost no attention in the deposits that cover the earth's surface (Dott, 1983). Even many shale deposits, which we think of as formed by slow "background sedimentation," upon closer examination are shown to be the result of episodic deposition and erosion (Schieber, 1998).

The record of sedimentation and environmental change at any given locality is the result of the interaction of local, regional, and global processes. Proximity to a nearby river delta system, flexural subsidence of a foreland basin adjacent to a rising mountain belt, and a global eustatic sea level rise are examples of each. Even thin rock layers, from a single, shell-rich, storm-derived tempestite bed to a basin-wide and possibly globally correlatable black shale layer formed during a maximum sea-level highstand, mark different scales of processes acting locally through geologic time.

However, in spite of the complex interaction of a broad range of local to global processes, and the spotty record of time preserved in sedimentary rocks, certain layers are continuous and can be traced along great distances. Such marker beds, when recognized, can sometimes be very widely correlated and used to subdivide the rock record at a scale much finer than possible using classical lithostratigraphic and biostratigraphic methods (Kauffman, 1988). A thin bentonite clay, deposited as a volcanic ash layer over hours to days, forms an almost instantaneous slice of time through a stratigraphic succession. Bentonites and other time-significant marker beds of a sedimentologic, paleobiologic, or chemical origin, can be widely correlated and used to create a high-resolution subdivision of the sedimentary rock record (Kauffman, 1988). Using that data, we can more clearly delineate the relationships between different stratigraphic units in different parts of a sedimentary basin, and better understand the historical development of the region's geologic history.

Upper Lower and Middle Devonian strata of the Delaware Water Gap and adjacent areas record the interplay of the full range of local to global processes. And they contain within them a number of distinctive, time-significant marker beds that can be correlated widely across the Appalachian foreland basin. These permit the construction of a new basin-wide synthesis of strata from the Lower Devonian (Pragian) Oriskany Formation to the lower part of the Middle Devonian (Eifelian-Givetian) Hamilton Group. More significantly, the new data more tightly constrain the timing and distribution of events and processes active across eastern North America between 410 and 385 million years ago (dates from Tucker et al., [1998]).

In this paper we will examine the results of the new stratigraphic synthesis, describe the succession regionally, and discuss the sequence stratigraphy of upper Pragian to Givetian rocks of the region. One result of the new basinwide synthesis is the recognition of widespread synonomy of

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stratigraphic names—multiple names for what are the same strata. The paper will, in part, explore the existing stratigraphic scheme and informally present a preliminary revision of the regional stratigraphic nomenclature.

**Tectonic Setting**

The Early to early Middle Devonian was a time of significant change and reorganization of the Appalachian Basin. The collision of eastern North America with another, smaller continent and/or series of terranes (Rast and Skehan, 1993) resulted in four episodes of orogenesis through the Devonian and early Mississippian Acadian Orogeny (Ettensohn, 1985).

An initial carbonate shelf- or ramp-type basin geometry characterized the first approximate eight million years of the Devonian (based on Tucker et al., 1998) (Lochkovian-age Helderberg Group limestones, succeeded by Pragian-age Oriskany-Ridgeley quartz arenites and Glenerie-Shriver carbonates) deposited during a major episode of regression. The development of a significant unconformity around the margins of the basin (Wallbridge Unconformity of Sloss, 1963) is overlain by similar strata deposited during an initial, slow transgression.

Through the approximate 24 million years of the late Early to Middle Devonian (Emsian, Eifelian, and Givetian stages; duration from Tucker et al., 1998) the Appalachian foreland basin system underwent three major pulses of subsidence and foredeep development, the first two followed by a gradual return to basin geometries similar in character to the earlier ramp-like basin topography (Ettensohn, 1985; Ver Straeten, 1996a). Each of these cycles is reflected in deposition of initial basinal black shales, subsequent coarser clastics that become transitional to carbonate-dominated facies that cap the cycles. These patterns were discussed by Ettensohn (1985) in the context of tectonically active to quiescent tectophases (Tectophases I, II, and III of the Acadian Orogeny) that were superposed over a series of apparent global, eustatic sea level changes (Johnson et al., 1985; Ver Straeten, 1996a).

Reconstruction of the details of a basin's history is dependent upon basinwide, high-resolution stratigraphic correlations of the sedimentary succession. Recognition and correlation of widely distributed marker units, such as altered volcanic ash falls (K-bentonites), phosphate- and or glauconite-rich beds, maximum flooding surfaces, faunal epiboles, and other distinctive cyclic and event deposits, permit a fine-scale subdivision of the sedimentary fill at a resolution greater than achievable by classical litho- or biostratigraphic methodologies.

Ettensohn (1985) outlined his initial model of Acadian Tectophases on the overall sedimentary changes through the foreland basin fill, as seen in the upper Lower and lower Middle Devonian (Pragian, Emsian, and Eifelian stages) of eastern Pennsylvania. A more detailed interpretation of the basinal history during the 1980's was hampered by the lack of a high-resolution stratigraphy. Recent basinwide analysis of the Pragian to Eifelian, however, presents new data from which to examine the interplay of foreland basin flexure, sedimentation, and eustasy within the basin, and orogenesis along the eastern margin of the North American (Laurentian) continent.

The interval of study in Pennsylvania comprises two successions of quartz arenites and carbonates succeeded by dark gray to black shales. Strata between the two successions mark a gradual return to more carbonate-rich conditions. These comprise Acadian Tectophase I and the beginning of Tectophase II. The two cycles are represented by: 1) Oriskany-Ridgeley quartz arenites to Esopus shales-sandstones, succeeded by the mixed clastics and carbonates of the Schoharie Formation; and 2) Onondaga Limestone to the so-called "Marcellus Shale" (Bakoven black shale, Stony Hollow-Hurley-Cherry Valley mixed clastics and carbonates, and succeeding Oatka Creek-Brodhead Creek-"upper Marcellus" black shales).

## STRATIGRAPHIC OVERVIEW

### Stage-level Terminology, Past and Present

Various North American stage-level terminologies have been applied to the upper Lower to lower Middle Devonian strata examined in this study. Based on the New York succession, Rickard (1975) proposed a scheme of four stages: a) The Deerpark Stage (=Port Jervis and Oriskany Formations); b) the Sawkill Stage (Esopus and Schoharie Formations; c) the Southwood Stage (lower three members of the Onondaga Limestone); and d) the Cazenovian Stage (the Seneca Member of the Onondaga Limestone and overlying strata of the "Marcellus Shale," in New York now subdivided into the Union Springs (lower) and Oatka Creek Formations (Ver Straeten et al., 1994, in preparation; see below). The combined Sawkill and Southwood Stages originally was termed the "Onesquethaw Stage" by Cooper et al. (1942) and has been widely used by workers in the central and southern part of the Appalachian Basin, including eastern Pennsylvania (e.g., Dennison, 1960, 1961; Inners, 1975, 1979; Epstein, 1984).

With increasing global communication between Devonian workers, fostered by the international Subcommission on Devonian Stratigraphy (SDS), the internationally recognized Pragian, Emsian and Eifelian stages are recommended and applied here. The Deerpark and Sawkill Stages are approximately equivalent to the European Pragian and Emsian Stages; the Southwood and Cazenovian stages are in general equivalent to the Eifelian Stage of European terminology (Rickard, 1975). New conodont studies are presently underway in New York and Pennsylvania to better resolve correlations between the Appalachian Basin and Emsian-Eifelian successions worldwide.

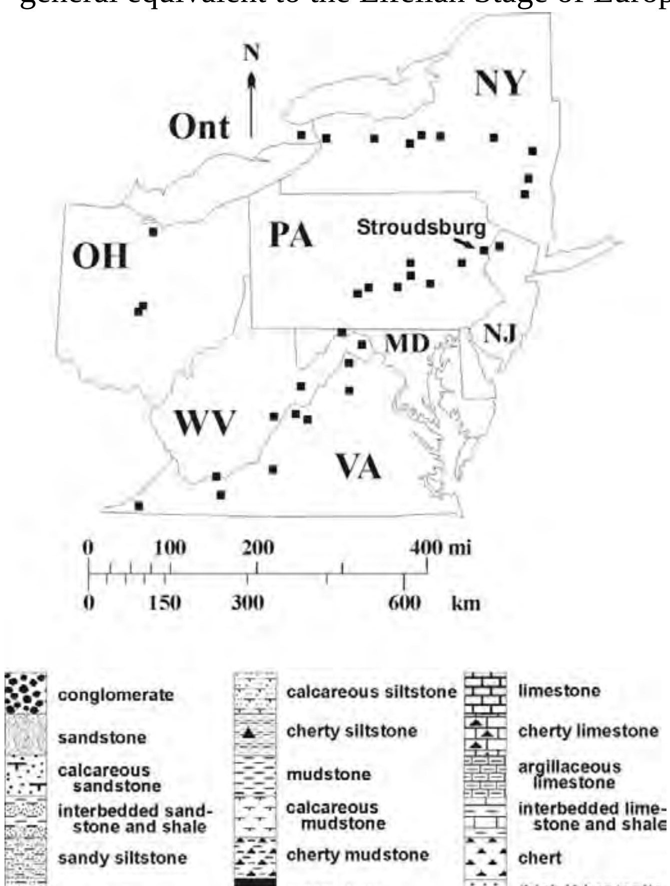


Figure 21. Key outcrops of upper Lower to lower Middle Devonian strata, Appalachian foreland basin. Map shows the location of key outcrops of upper Pragian to Eifelian strata across the Appalachian Basin, and represents >1/10 of outcrops utilized in this study.

### Stratigraphic Revisions of Pragian to Eifelian strata, Eastern Pennsylvania and New York

#### *New Stratigraphic Interpretations*

Extensive event and cyclic stratigraphic analysis of upper Lower and lower Middle Devonian outcrops in eight states (>350 outcrops, NY, NJ, PA, MD, VA, WV, TN, and OH; Figure 21) permits widespread, relatively fine-scale correlation of Oriskany-Ridgeley to lower Hamilton ("Marcellus Shale") strata along the entire Appalachian Basin outcrop belt. Figure 22 illustrates a new synthesis of formation- and member-level stratigraphic relationships for upper Pragian-, Emsian, and Eifelian-age strata across Pennsylvania and New York. Accompanying Figure 23 outlines a generalized facies overlay for the same strata.

The multiplicity of names for the same strata across a region creates a clutter of terms that mask stratigraphic relationships, as can be seen in different Appalachian Basin stratigraphy charts published over the years (e.g., Cooper et al., 1942; Rickard, 1975; Patchen et al, 1985; Berg et al., 1986). One result of

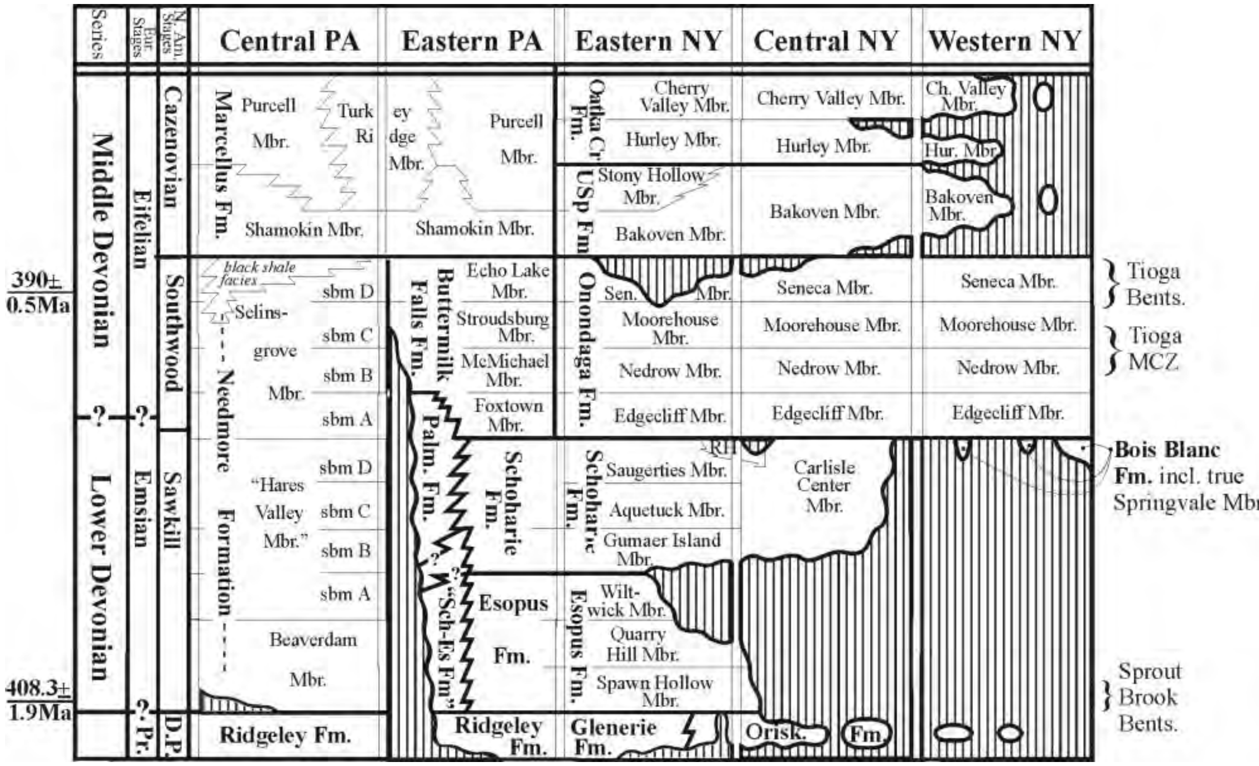


Figure 22. Stratigraphic synthesis, upper Lower to lower Middle Devonian, Pennsylvania and New York. Formation- to member-level stratigraphic synthesis of upper Pragian, Emsian and Eifelian strata along the outcrop belt in Pennsylvania and New York. Modified after Ver Straeten (1996a, 1996b). *Abbreviations:* D.P. = Deer Park; OatCr = Oatka Creek; Pr = Pragian; RH = Rickard Hill Mbr.; sbm = submember.

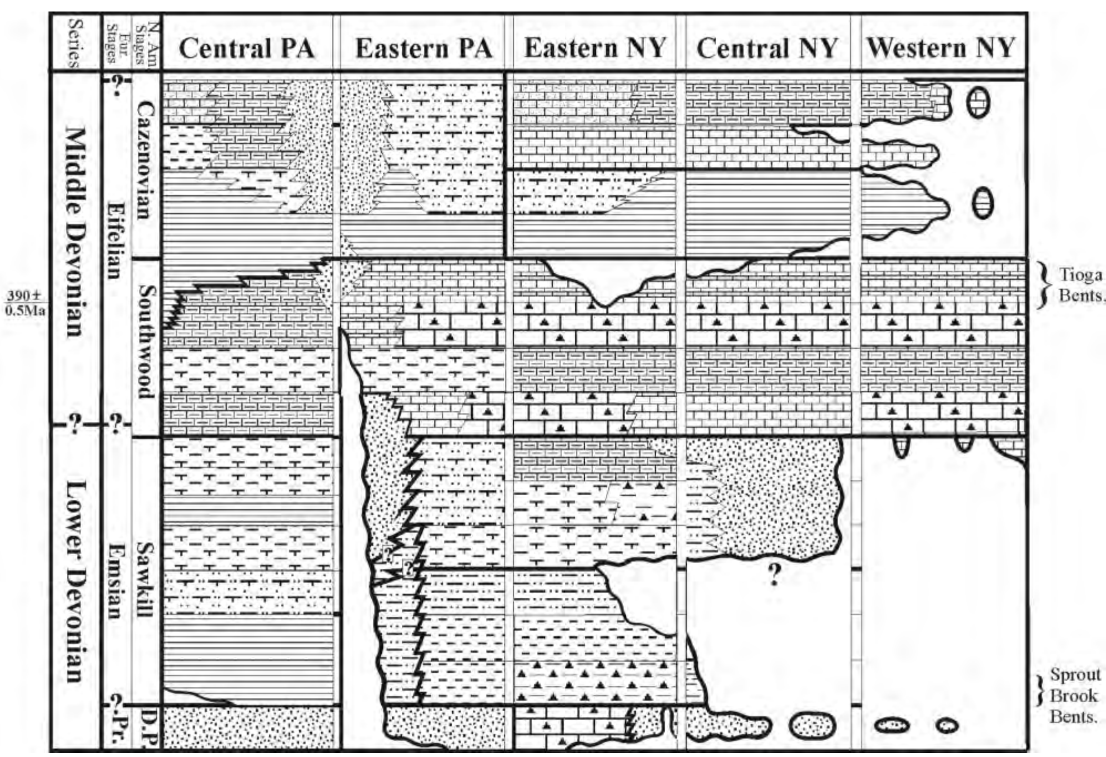


Figure 23. Lithofacies, upper Lower to lower Middle Devonian, Pennsylvania and New York. Overlay diagram of Figure 22, showing lithofacies of upper Pragian, Emsian, and Eifelian strata along the outcrop belt in Pennsylvania and New York.

the new correlations is the recognition of widespread synonomy in the stratigraphic nomenclature for the same strata across the basin. The International Stratigraphic Guide (ISG; Salvador, 1994) and the North American Stratigraphic Code (1983) state that the demonstration of synonomy is grounds for abandonment of formal stratigraphic terms. The IGS further states that the later name should be replaced by the earlier.

In a paper in preparation, Ver Straeten and Brett will propose nomenclatural changes for upper Pragian to Eifelian strata in the Appalachian Basin, based on priority of older names or, where appropriate, propose new names. A preview of parts of this stratigraphic revision is shown in Figure 24, and further discussed in the following text and Figures below.

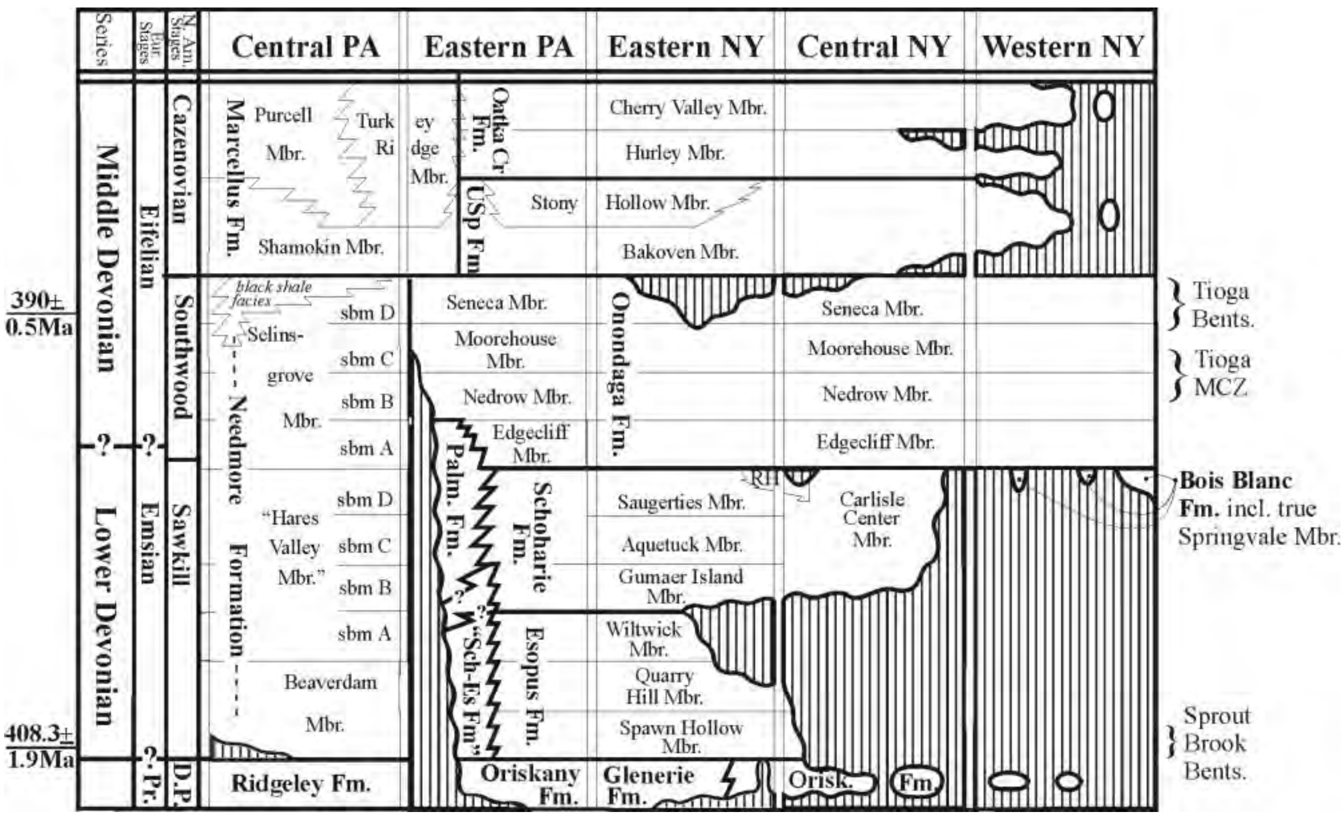


Figure 24. Preliminary stratigraphic revision of upper Lower to lower Middle Devonian nomenclature, Pennsylvania and New York. Overlay of Figure 22, showing proposed informal revisions of upper Pragian to Eifelian strata in New York and eastern Pennsylvania.

## THE FORMATIONS OF EASTERN PENNSYLVANIA

### Oriskany Formation

Previous studies and new preliminary work on the Oriskany Sandstone of New York (Vanuxem, 1839) and time equivalent conglomerates, quartz arenites, and cherty carbonates across the Appalachian Basin show a complexity of facies changes and unconformity development. Quartz-rich sandstones and conglomerates formerly assigned to the Ridgeley Formation (Schuchert et al., 1913) of Pennsylvania are correlative with the Oriskany Formation of and equivalent limestones and conglomerates (Glenerie and Connelly Formations, respectively) in New York State. Due to the synonomy of the two names, the older term Oriskany is herein applied to the strata. This is also consistent with the use of the name Oriskany in states to the south of Pennsylvania.

Throughout Pennsylvania sand-dominated facies of the formation features the distinctive Oriskany Sandstone fauna of large, robust brachiopods (e.g., *Costispirifer*, *Rensselaeria*, and

*Hipparionyx*), mixed with numerous other forms, especially in finer- grained, more calcareous strata. The formation typically appears as a white, bimodal quartz sandstone and quartz-pebble conglomerate with brachiopod molds. Non- to slightly-arenaceous, cherty limestone in the northeastern part of the state, dominantly in the subsurface (Rickard, 1989), represents a southwestward extension of the Oriskany-equivalent Glenerie Limestone of southeastern New York. In other regions of the state, sand-dominated strata may interfinger with chert-rich calcareous facies of the Shriver Formation, which is largely synonymous with the cherty Glenerie Limestone. The Shriver generally underlies Oriskany Sandstone facies, but increasingly dominates the succession in more distal, offshore facies.

In the Stroudsburg area, sandstones and cherty limestones of the Oriskany and underlying Shriver/Glenerie Formations total approximately 20-30 m in thickness. Northward through the Delaware Water Gap area, the sand content in the upper strata initially increases (Alvord and Drake, 1971), then decreases toward the New Jersey-New York border area where the entire Oriskany-equivalent interval are represented by relatively chert-poor, silty limestones (D. Monteverde, personal communication, 2001).

The Oriskany and equivalent Glenerie-Shriver Formations occur within the *Costispirifer arenosus* subzone of the *Rensselaeria* (brachiopod) Range Zone and the *Icriodus huddlei* conodont zone (Lower Devonian; Dutro, 1981; Klapper, 1981). Further discussion of the Oriskany Formation in Pennsylvania may be found in Cleaves (1939) and numerous Pennsylvania Geological Survey Atlases (e.g., Epstein et al., 1974). Much work remains to be done on this interval of Lower Devonian rocks.

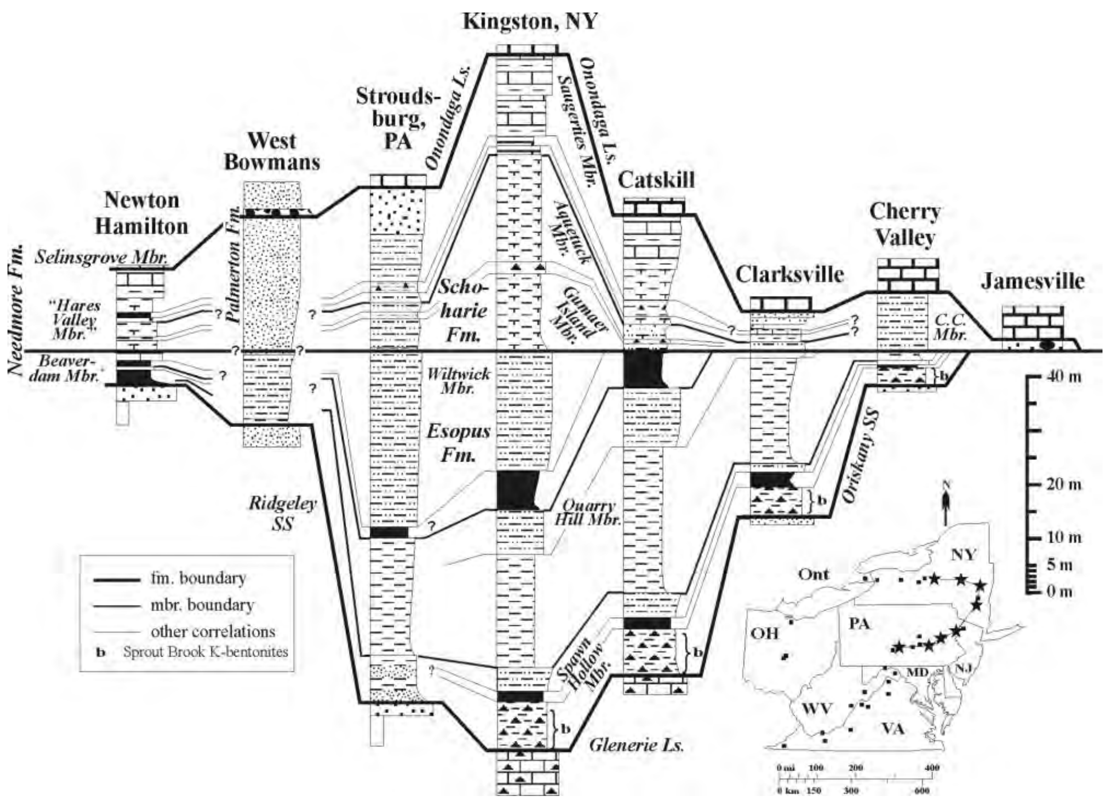

Figure 25. Correlation of Emsian-age Esopus and Schoharie Formations and equivalent strata, Pennsylvania and New York. Figure shows the correlation of time-significant markers in the Esopus and Schoharie Formations in New York and Pennsylvania. The Esopus Formation represents a "third order" depositional sequence, and is composed of three members (Spawn Hollow, Quarryville, and Wiltwick), each of which represents separate subsequences. The Schoharie Formation comprises a second depositional sequence; the Gumaer Island and Aquetuck-Saugerties Members comprise two subsequences within the formation. As shown (Newton Hamilton), the Beaverdam Member and basal thin limestones of the "Hares Valley Member" are the lateral equivalent of the Esopus Formation. Datum = base of Schoharie Formation.

**Esopus Formation**

Medium- to dark-gray shales and siltstones to fine-grained sandstones of the Esopus Formation (Darton, 1894) in eastern Pennsylvania are continuous with Esopus strata in the type area of eastern New York (see Figures 24 and 25). Up to 65 m of fine-grained siliciclastics of the Esopus Formation crop out in the vicinity of Stroudsburg (Inners, 1975; Rehmer, 1976). Rickard (1989) reports in excess of 120 m in the subsurface in northeastern Pennsylvania near the New York border (Pike County). The Esopus "Shale" is comprised dominantly of generally non-calcareous, dark-gray shales, siltstones, and fine-grained sandstones that commonly appear highly burrowed to bioturbated. Southwest of Stroudsburg the Esopus Formation thins and pinches out against the southeastern margin of the basin (Auburn Promontory of Swartz and Swartz, 1941) near Schuylkill Haven (90 km southwest of Stroudsburg). The Esopus and overlying Schoharie Formations have been combined by some workers into one undifferentiated unit in the region southwest of Stroudsburg (Schoharie-Esopus Formations; Epstein and Epstein, 1967, 1969; Inners, 1975; Epstein, 1984).

Few workers previously attempted to subdivide the Esopus Formation into member-level subdivisions (however, see Boucot et al., 1970; Rehmer, 1976). Detailed study of the formation, initially in New York, indicates a set of three distinct units comprise the Esopus Shale (Figure 25). The lower and upper of the three units are herein informally introduced as the Spawn Hollow and Wiltwick Members (Figure 22). The term Quarryville Member was previously proposed by Boucot et al. (1970) for middle strata of the formation. The subdivision is based on both lithologic differences and the recognition of three distinct cyclic packages of strata that comprise the formation in its type area. At least the upper two cycles, and possibly all three are correlatable across the Appalachian Basin from eastern New York through eastern and into central Pennsylvania, where they comprise strata of the Beaverdam and lowest part of the "Hares Valley Member" of the Needmore Formation (see Figure 25). The upper two, and apparently all three members, can be further correlated to southwestern Virginia, where they are distinguishable as the lower dark gray shales and cherts of the Huntersville Formation (Ver Straeten, 2001).

The Spawn Hollow Member at the base of the Esopus Formation is characterized by interbedded thin siltstones and shales, a black shale, and a capping siltstone to fine sandstone unit (in vertical succession). Throughout eastern New York the interbedded interval also features up to 15 thin K-bentonites, the Sprout Brook K-bentonites (Ver Straeten, 1996a). No evidence of the K-bentonites is yet noted in the Stroudsburg area. The succeeding Quarryville Member in the middle of the formation comprises a thick, homogenous, mudstones, topped by a pair of siltstone to fine sandstone beds. The rock generally appears highly bioturbated, predominantly by Chondrites trace fossils.

The base of the upper member of the Esopus (Wiltwick Member) is marked by a distinctly laminated black shale or interlaminated dark gray shale and siltstones. An approximate 2 m-thick unit near the base of the road cut along the west side of US 209, adjacent to Buttermilk Falls, appears to represent this unit. It is succeeded by a thick succession of siltstones to very fine and fine sandstones to the top of the formation. The silt- to sandstones generally appear massive; close examination shows a horizontal pinstriped, fully bioturbated texture of *Zoophycos* traces.

The lower contact of the Esopus Shale in the Delaware Water Gap area appears sharp above the Ridgeley Sandstone or, locally, the Glenerie Limestone (Shriver Chert). Megafossils are generally rare to uncommon in the Esopus Formation, and generally consist of small brachiopods (*Atlanticocoelia* and *Leptocoelia*). Abundant ichnofossils in the formation include *Zoophycos* and *Chondrites*. The Esopus Formation in New York is assigned to the Emsian-age *Etymothyris* (brachiopod) Zone (Lower Devonian; Boucot, 1959). Detailed discussions of the Esopus Formation in eastern Pennsylvania are given in Inners (1975), Rehmer (1976), Epstein (1984), and Ver Straeten (1996a, 1996b). Zircons from one of the Sprout Brook K-bentonites in the lower part of the formation was recently dated (Tucker et al., 1998) and yielded a geochronologic age of 408.3 + 1.9 Ma.

**Schoharie Formation**

The Schoharie Formation (Vanuxem, 1840) of eastern Pennsylvania consists of calcareous mudstones and siliceous siltstones to fine sandstones. In the Stroudsburg area (US 209 road cuts, Buttermilk Falls) the formation is 31 m in thickness (Ver Straeten, unpublished data; see Figure 25 and accompanying paper by Ver Straeten, this guidebook, p. 54). Two subdivisions are recognized in the Stroudsburg region (Inners, 1975); a lower massive, dark gray, pyritic, calcareous mudstone with common *Zoophycos* or *Chondrites* traces and an upper massive, dark gray siliceous to calcareous mudstone and siltstone to sandstone unit with vertical burrows and common phosphatic nodules.

The Schoharie Formation in eastern Pennsylvania is more fully discussed in the guidebook paper referred to above.

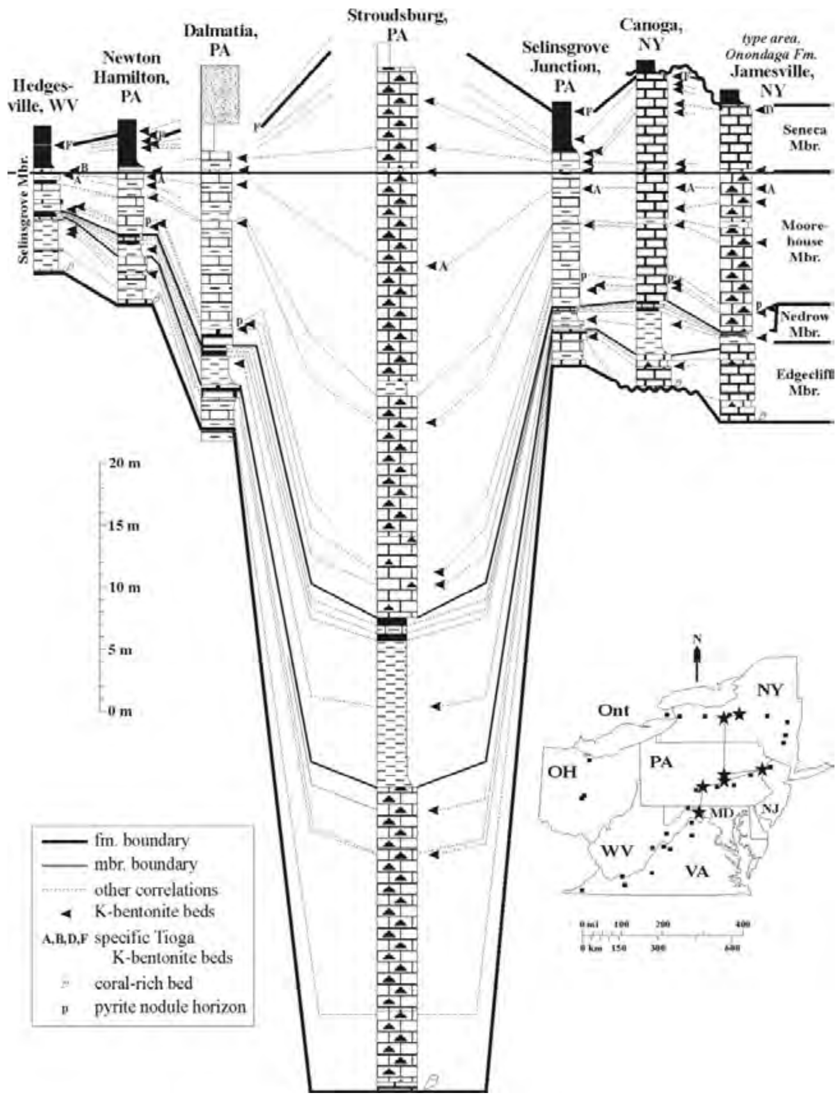


Figure 26. Correlation of Emsian- (?) and Eifelian-age Onondaga and Selinsgrove Limestones, Pennsylvania and New York. Figure shows the correlation of time significant marker beds in the Onondaga Limestone of New York and eastern Pennsylvania, and the lateral equivalent Selinsgrove Member (Needmore Formation) in central Pennsylvania to northern West Virginia. Note the consistent geometric relationships between marker beds, even through the thick and thin successions (Hedgesville, 10 m; Stroudsburg, 83 m). Note also the additional thin K-bentonite beds below the Tioga A-G interval of Smith and Way (1983) and Way et al. (1986). Datum = Tioga B K-bentonite of Smith and Way (1983; = Onondaga Indian Nation K-bentonite of Conkin and Conkin, 1984).

**Onondaga Limestone**

In the Stroudsburg area of eastern Pennsylvania post-Schoharie strata comprise a thick succession of two cherty, generally fine-grained limestone bodies, separated by an intervening calcareous shale-rich unit (Figure 26). The strata, formerly termed the Buttermilk Falls Formation of Willard (1939), is the direct lateral equivalent of the Onondaga Limestone. Therefore, it is herein informally abandoned and the strata are assigned to the Onondaga Limestone (or Formation).

At Stroudsburg, the Onondaga Limestone conformably overlies the Schoharie Formation. Three to four members previously recognized locally (Epstein, 1984; Inners, 1975) are the direct lateral equivalents of the four members of the Onondaga Limestone in New York State (Ver Straeten, 1996a, 1996b). The four members (formerly Foxtown, McMichael, Stroudsburg, and Echo Lake Members, designations abandoned) are visible in a nearly complete outcrop in the railroad cut at East Stroudsburg (STOP 7, Day 2). The members as herein proposed are (Figures 24 and 26): 1) the Edgecliff Member (25 m-thick), comprised of dark-gray-weathering, generally fine-grained, cherty limestone. Corals are abundant in the base, and large crinoid columns (holdfast fragments?) are abundant in the lower part of the member; 2) the Nedrow Member (13 m-thick), characterized by dark-gray, fossiliferous, calcareous shales and interbedded gray-weathering, fine-grained, fossiliferous, nodular to tabular-bedded limestone. Fossils are abundant in the Nedrow Member; 3) the Moorehouse Member (36.6 m-thick), a gray-weathering, fossiliferous, fine- to medium-grained limestone with tabular-bedded to nodular cherts and minor calcareous shales. Fossils commonly occur throughout the member; and 4) the Seneca Member (6.7 m-thick), relatively similar to the underlying Moorehouse Member, but lighter weathering and characterized by coarser, chonetid brachiopod (Hallinetes) coquinites and a lesser component of chert.

Multiple altered volcanic ashes, the Tioga A-G K-Bentonite beds, are found in upper Onondaga strata across eastern Pennsylvania (Epstein et al., 1974; Inners 1975; Way et al., 1986) and across the Appalachian Basin. A prominent K-bentonite bed, the Tioga B bed of Way et al. (1986; = Onondaga Indian Nation ash in New York [Conkin and Conkin, 1979, 1984; Conkin, 1987]) occurs in the upper part of the Moorehouse Member. The Tioga B bed marks the base of the upper member (Seneca Member) of the Onondaga Limestone in New York.

Southwest of Stroudsburg the Onondaga Limestone begins to thin and change character, becoming less to non-cherty and relatively fossiliferous. The four members reported from Stroudsburg were previously not recognized. However, recent study shows that all four members are recognizable at least as far southwest as Lehigh Gap, where Edgecliff equivalent strata comprise the upper 7 m of the Palmerton Sandstone, marked at its base by a distinctive, meter-thick conglomerate bed (Ver Straeten, 1996a, 1996b).

The Onondaga is reported to be absent in the vicinity of Susquehanna Gap near Harrisburg (Willard, 1939). However, northwest of Harrisburg (Lambs Gap), black Bakoven shales conformably overlie fossiliferous quartz sandstones and conglomerates and a 23 cm-thick greenish clay that appears to be a K-bentonite bed. The fauna of the sandstones has not been studied, but it is the author's suggestion that the strata represent sandy, nearshore Onondaga facies, and that the prominent clay bed represents one of the prominent Tioga K-bentonite beds.

The four members of the Onondaga Limestone, and numerous thin marker beds, can be directly correlated to four subdivisions in the Selinsgrove Limestone Member (Needmore Formation) in central Pennsylvania and southward into Virginia and West Virginia (Ver Straeten, 1996a, 1996b, 2001). In southwestern Virginia, the Edgecliff Member is represented by the Bobs Ridge Sandstone at the top of the Huntersville Chert (Ver Straeten, 2001).

The widely recognized Tioga B K-bentonite bed at the base of the Seneca Member has been dated by Roden et al. (1990) at 390 + 0.5 Ma. The position of the Emsian-Eifelian stage boundary in eastern North America (= base of the Polygnathus costatus patulus conodont zone), is presently not recognized, but occurs between the base and top of the Edgecliff Member. Conodonts diagnostic of the succeeding Polygnathus costatus partitus conodont zone are found within the Nedrow Member; the top of the Nedrow marks the first occurrence of forms diagnostic of the *Polygnathus costatus costatus* zone (Klapper, 1971, 1981). The Edgecliff, Nedrow, and lower part of the Moorehouse Members occur in the upper part of the *Amphigenia* brachiopod Assemblage Zone, in the upper of two subzones, the *Fimbrispirifer divaricatus* subzone (=large Amphigenia Zone; Dutro, 1981). Upper Moorehouse and Seneca strata are within the overlying *Paraspirifer acuminatus* zone. Oliver and Sorauf (1981) summarize the rugose coral biostratigraphy for the Onondaga Limestone, and report two Assemblage

Zones (*Acinophyllum segregatum* zone and an unnamed zone). The first is broken into two subzones which include the Edgecliff (*Synaptophyllum arundinaceum* subzone) and the Nedrow-Moorehouse Members (Eridophyllum seriale subzone).

**"Marcellus Shale" (lower part)**

The term "Marcellus Formation" (Hall, 1839) in Pennsylvania is applied to black shale-dominated strata in the lower part of the Middle Devonian Hamilton Group. Different from the traditional usage of "Marcellus" in New York State, where it represented a time-stratigraphic unit, the name in Pennsylvania has been applied to all lower Hamilton Group organic-rich black shale facies.

Detailed study of the Marcellus Shale across the Appalachian Basin indicates that it is comprised of two distinctive, major successions, comparable in scale to each of the three other formations (Skaneateles, Ludlowville, and Moscow) of the Hamilton Group in New York State. Therefore, in New York the Marcellus "Shale" has been raised to subgroup status, and formation- and member-level stratigraphy has been revised (Ver Straeten et al., 1994; Ver Straeten and Brett, in preparation). Two formations are recognized within the Marcellus Subgroup, both former members of the Marcellus that are raised to formation level (Union Springs and Oatka Creek Formations). Two members comprise the Union Springs Formation, the previously established Bakoven and Stony Hollow Members. The lower part of the overlying Oatka Creek Formation consists of three members, the Hurley, Cherry Valley, and Berne Members. (Note: the Hurley Member was previously placed in the Union Springs Formation by Ver Straeten et al., 1994.)

The Stony Hollow Member of eastern New York was originally correlated with the Cherry Valley Member, found west of Albany (Cooper, 1941). The latter, however, comprises a separate unit at the top of the Stony Hollow, as does an intermediate unit that comprises the Hurley Member (Ver Straeten 1996a). (Note: the Hurley Member was previously placed by Ver Straeten et al., 1994, in the Union Springs, but is here assigned to the base of the Oatka Creek Formation.)

Recent work in Pennsylvania shows that much of the same stratigraphic framework of the lower part of the Marcellus subgroup in New York can be recognized across Pennsylvania and areas to the south. In the Stroudsburg area, strata above the Onondaga Limestone have been termed the "Union Springs Member" and Stony Hollow Member by Epstein and Epstein (1969) and Alvord and Drake (1971). Following the revision outlined above, the Union Springs is raised to formation-level status. Initial black shales, assigned to the Bakoven Member, are overlain, as previously, by buff-weathering calcareous shales to siltstones of the Stony Hollow Member.

Poor exposure in the Stroudsburg area does not as yet permit recognition of the overlying Hurley and Cherry Valley Members locally. However, both are recognized to the southeast, at Palmerton, and in most outcrops throughout central Pennsylvania (Ver Straeten, 1996b). As across the basin, the Cherry Valley Member is succeeded by additional black shales, in the local area termed the Brodhead Creek Member.

Key references on the lower part of the Marcellus Formation and related strata in Pennsylvania include Willard (1935, 1939), Cate (1963), Faill et al. (1978), and Ver Straeten (1996a,1996b). Additional discussions of the Marcellus Shale in eastern Pennsylvania include Epstein et al. (1974).

Basinwide, the biostratigraphy of the lower part of the Hamilton Group is poorly resolved. Studies by Klapper (1971, 1981) indicate that the Bakoven Member occurs within the *Tortodus kockelianus australis* conodont zone and the overlying Hurley and Cherry Valley Members lie within the *Tortodus kockelianus kockelianus* zone. Conodont faunas of the overlying parts of the Oatka Creek Formation have received little if any attention. The Eifelian-Givetian stage boundary falls within the

Oatka Creek Formation, above the Cherry Valley Member. The lack of detailed biostratigraphic data, however, prevents identification of the precise position of the boundary.

***Union Springs Formation, Marcellus Subgroup***

Bakoven Member: Organic-rich black shale facies of the lower part of the Marcellus Formation in Pennsylvania, which variously underlie the Stony Hollow, Purcell, or Turkey Ridge Members, are assigned herein to the Bakoven Member of the Union Springs Formation. The shales are characteristically fissile, non-calcareous, and pyrite-rich. Small to large carbonate concretions may be found, especially in the lower part and locally in uppermost strata. The black shales are generally unfossiliferous or feature low diversity, styliolinid-diminutive brachiopod-cephalopod assemblages similar to those reported for equivalent strata by Brower and Nye (1991) in New York State. Thin to thick intervals of deformed, heavily-slickensided shale occur at different levels, but are especially abundant beneath overlying, more resistant strata (see STOP 8, Day 2). In more basinward areas of Pennsylvania, the black Bakoven shale facies interfinger with and laterally replace strata of the Stony Hollow, Purcell, and Turkey Ridge Members. Thickness of the black Bakoven Shale varies regionally from zero to >90 m across the state (Willard, 1939; Faill et al., 1978). The Bakoven Member is poorly exposed throughout much of eastern Pennsylvania and northwest New Jersey.

A thin, widely recognizable K-bentonite bed, generally 3-6 cm-thick, occurs in the middle part of the Union Springs Formation and its equivalents throughout the basin. First discovered in central Pennsylvania, it is now recognized from eastern New York to southwestern Virginia to central Ohio, where it is found within the middle of the Delaware Limestone. The bed typically appears as a honey-tan to light gray, soapy-feeling clay bed and forms a prominent, continuous recession along the outcrop. The position of the bed relative to the top of the Bakoven black shale varies along the outcrop belt, due to facies changes between distal and proximal facies in the Union Springs Formation. In distal areas of the basin, the K-bentonite is found within black shales assigned to the Bakoven Member. However, in more proximal areas, it occurs near the top of the Bakoven, closely to the base of the Stony Hollow Member, or in the Harrisburg (PA) area, the base of the Turkey Ridge Member (Mahantango Formation).

Stony Hollow Member: The Stony Hollow Member is a calcareous to dolomitic, fine- to medium-grained siliciclastic unit that characteristically weathers buff to dark gray. With the overlying Hurley and Cherry Valley Members, it commonly forms a prominent ridge between the less resistant dark shales of the underlying Bakoven and overlying Brodhead Creek Members. The Stony Hollow Member ranges from finely laminated shales and siltstones with a low degree of burrow mottling in the lower part of the unit to a bioturbated fine- to medium-grained sandstone near the top. Rare pelagic fossils (e.g., *dacryoconarids, Styliolina*) in the lower part of the unit are replaced by uncommon benthic trilobites, brachiopods, and small rugose corals near the top. These comprise a distinctive fauna that is unique to the Stony Hollow and overlying Hurley Members, one that differs from that of both the underlying Onondaga Limestone and the overlying remainder of the Hamilton Group. The brachiopods *Variatrypa arctica, Kayserella, Carinatrypa,* and *Pentamerella* cf. *P. winteri,* a small rugose coral (*Guerichiphyllum,* formerly *Nalivkinella*), and a trilobite (*Dechenella haldemanni*) are the most characteristic forms of this unique fauna. The *Variatrypa arctica* fauna was widespread throughout the Eastern Americas Realm at that time, including in Iowa, the Michigan Basin, and the James Bay region of northern Ontario (Day and Koch, 1994; Ehlers and Kesling, 1970; Sanford and Norris, 1975). Koch (1978) attributes the appearance of this fauna to a “limited migration” of Old World Realm brachiopods from Arctic North America during deposition of the Union Springs Formation. Koch (1988) states that the breakdown of paleobiogeographic boundaries may be due to continent-wide transgression at that time.  Boucot (1990) associates this migration with a "geologically sudden lowering of the global climatic gradient" (discussed below) that resulted in dispersal of warmer water faunas into the Appalachian Basin during Union Springs time.

***Lower part of the Oatka Creek Formation, Marcellus Subgroup***

Hurley and Cherry Valley Members: The Hurley Member comprises a lower thin, richly fossiliferous limestone and overlying shale-dominated strata (Ver Straeten et al., 1994).  The unit is found throughout New York to central Pennsylvania.  The Hurley in central to western New York and Pennsylvania is generally relatively thin (ca. 30-60 cm).  As noted above, the unique fauna found predominantly in the lower limestones ("Chestnut Street beds" of Griffing and Ver Straeten, 1991) is characterized by the proetid trilobite *Dechenella haldemanni* and the small rugose coral *Guerichiphyllum echoense*, originally described by Oliver (1964) from samples of the Stony Hollow Member near Echo Lake, PA (close to STOP 8 of Day 2).  The Hurley Member represents the basal unit of the Oatka Creek Formation throughout its outcrop belt.

The overlying Cherry Valley Member at the base of the Mount Marion Formation in eastern New York is characterized by a cephalopod-rich limestone lithology that extends west of the Albany area, and terrigenous sand-rich strata south of Albany through the Hudson Valley outcrop belt.  The term Cherry Valley Member was expanded by Ver Straeten et al. (1994) to include equivalent strata in eastern New York formerly placed in the upper part of the Stony Hollow Member.  The revised Cherry Valley Member is composed of two lithosomes—an eastern sand-dominated facies and a central to western carbonate-dominated facies.

Strata correlative with the Hurley Member (Union Springs Formation) are recognizable across most of the Pennsylvania outcrop, most notably the proetid trilobite-bearing Chestnut Street Submember (Figure 27).  These beds can presently be correlated all across eastern to south-central Pennsylvania to the region along the Maryland border, where they have not as yet been positively identified.  Recognition of the Hurley Member also permits positive identification of strata coeval with the Cherry Valley Member.  The Cherry Valley equivalent in Pennsylvania may, as in New York, be represented by limestone- or sandstone-dominated facies.  The classic cephalopod fauna, dominated by *Agoniatites vanuxemi* and *Striacoceras typum*, is commonly found in more carbonate-dominated exposures.

At Palmerton, 12 m of black shales of the Bakoven Member overlie a covered interval above the Onondaga Limestone.  A thin (ca. 2 cm-thick), light-weathering clay near the base of the exposure may represent one of the Tioga bentonite beds (Tioga G? of Way et al., 1986).  Small to medium-sized carbonate concretions occur in the upper part of the Bakoven Member.  The black shales are directly overlain by a deeply leached limestone bed (ca. 40 cm-thick) that features proetid trilobites (*Dechenella haldemanni*) and other forms; the fossiliferous bed and a thin overlying shale represent the Hurley Member (Figure 27).  Overlying calcareous shales and fine-grained limestones (3.5 m-thick) are the lateral equivalent of the Cherry Valley Member in New York State.

At Swatara Gap northeast of Harrisburg (Figure 27), farther to the southwest, 21 m of black shale, capped by 2-3 m of transitional, increasingly silty black to dark gray strata, comprise the Bakoven Member.  A thin, iron-rich clay bed with biotite 0.8 m below the top of the unit is the widespread mid-Union Springs K-bentonite found throughout the Appalachian foreland basin.  A very thick section of relatively coarse, quartz-rich sandstones (Turkey Ridge Member, Mahantango Formation) overlie the Bakoven Member at Swatara Gap.  Apparent plant root traces characterize the middle to upper parts of the Turkey Ridge sandstones, indicate at least periodic subaerial exposure within the unit.  It is not presently known whether the entire thickness of the sandstone at Swatara Gap is correlative with the

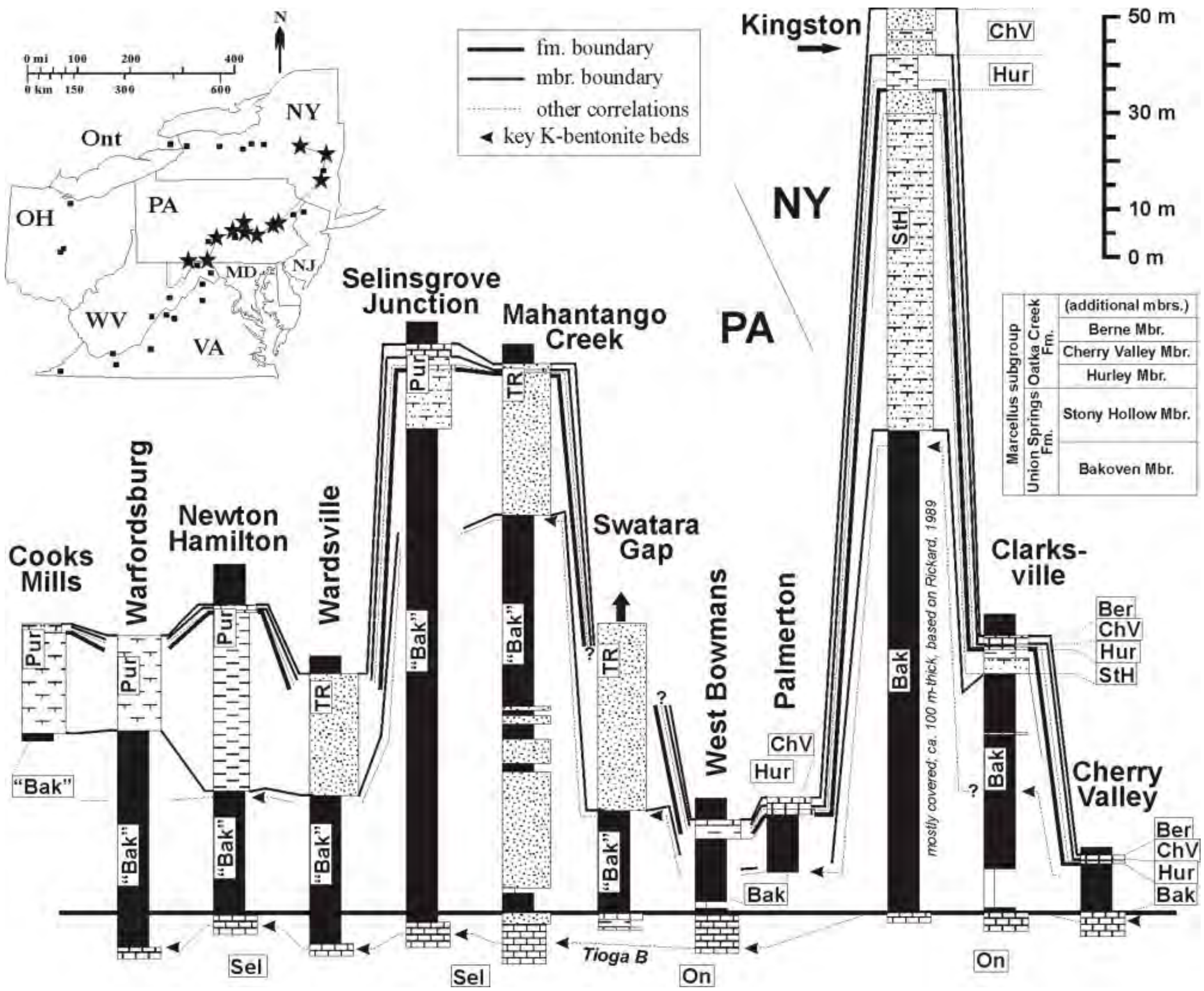


Figure 27. Correlation of Eifelian-age Union Springs and lower part of Oatka Creek Formations and equivalent strata, Pennsylvania and New York. Figure shows the correlation of time-significant markers in strata of the Union Springs (with Bakoven and Stony Hollow Members) and the lower part of the Oatka Creek Formations Hurley, Cherry Valley, and Berne Members) and equivalent strata. The Turkey Ridge Member (Mahantango Formation) represents sand-dominated facies equivalent to the Stony Hollow, Hurley and Cherry Valley Members. In central Pennsylvania the Purcell Member ("Marcellus Shale") is variously correlative with the Stony Hollow (all or part), Hurley and Cherry Valley Members, or only the latter two units. Note the thin, widely correlatable K-bentonite in the middle of the Union Springs Formation; it can be correlated in many sections between eastern New York to southwest Virginia to central Ohio. Abbreviations: Bak = Bakoven Member; Ber = Berne Member ChV = Cherry Valley Member; Hur = Hurley Member; Pur = Purcell Member; StH = Stony Hollow Member; TR = Turkey Ridge Member. Datum = base of Tioga F K-bentonite of Smith and Way (1983) or, in its absence, the top of the Onondaga Limestone.

Turkey Ridge Member, or if the upper part of the very thick section is continuous into stratigraphically higher sandstones of Mahantango Formation.

Post-Cherry Valley strata of the Oatka Creek Formation. The Cherry Valley Member and equivalent strata across the Appalachian foreland basin is succeeded by a generally thick succession of black to dark gray shales and siltstones to sandstones that are assigned to various units. In New York State, immediately overlying strata are termed the Berne Member (Ver Straeten, 1996a). Across Pennsylvania, the strata are generally assigned to undifferentiated upper "Marcellus Shale" or, in the Delaware Water Gap area, the Brodhead Creek Member. In areas proximal to Harrisburg, post-Turkey Ridge strata are termed the Gander Run or Dalmatia Members of the Mahantango Formation.

These strata have received little recent attention as yet, with a few exceptions (e.g., Ver Straeten, 1994). Much work remains to better document the relationships of the post-Cherry Valley succession across the basin.

### Lower part of the "Marcellus Shale" in central Pennsylvania

Into central Pennsylvania, the correlatives of the Union Springs and Oatka Creek Formations are readily distinguishable at least as far as the Maryland border (Figure 27). Throughout the area, time-equivalent facies characteristic of both the Turkey Ridge and Stony Hollow-Hurley-Cherry Valley lithosomes are found. In areas proximal to Harrisburg, Turkey Ridge sandstones are found, closely underlain by the thin, mid-Union Springs K-bentonite. The lateral equivalents of the Stony Hollow, Hurley, and Cherry Valley members can be discerned in some Turkey Ridge outcrops (e.g., Mahantango Creek section, west side of Susquehanna River); in other sections they have not been recognized.

North and west of the area where Turkey Ridge facies are developed, a transition to calcareous shales and bedded to nodular, argillaceous limestones of the Purcell Member (Cate, 1963) is visible (Figure 27). In transitional outcrops, the sub-Hurley part of the succession features partial development of the Stony Hollow facies (e.g., Selinsgrove Junction section) overlying black shales equivalent to the Bakoven Member. In more distal outcrops (e.g., Washingtonville, Newton Hamilton) strata assigned to the Purcell Member consist only of the Hurley and Cherry Valley Members, just as seen in New York outcrops west of Albany (e.g., Cherry Valley). In these areas, the buff-weathering calcareous shales to siltstones of the Stony Hollow, which overlie the widely recognized mid-Union Springs K-bentonite, have been laterally replaced by Bakoven black shale facies. The Purcell Member and equivalent strata across central Pennsylvania and areas south feature numerous, small, golf-ball sized nodules of barite (Way and Smith, 1983; Nuelle and Shelton, 1986; Way, 1993).

South of Pennsylvania, along the Virginia-West Virginia border area, the lower part of the "Marcellus" shows the typical lower black shale and the distal (Hurley and Cherry Valley Members only) development of the Purcell Member. However, toward the southern part of the basin, Stony Hollow facies again reappear (e.g., north of Roanoke, VA), and submember-level units of the Stony Hollow seen in eastern New York are recognizable.

## SEQUENCE STRATIGRAPHY, UPPER LOWER AND MIDDLE DEVONIAN, PENNSYLVANIA AND NEW YORK

### Introduction

Sedimentary geology has been revolutionized over the last two decades, associated with the rise of the sequence stratigraphic paradigm. Sequence stratigraphy is powerful tool in the study of time-rock relationships, dividing the rock record into packages of cyclic, genetically related strata (Van Wagoner et al., 1988). The fundamental unit of sequence stratigraphy is a "depositional sequence, a coherent succession of strata that is bound at bottom and top by unconformities or their correlative conformities" (Mitchum et al., 1977). A sequence is formed through cyclic changes of relative sea level, a result of the interaction of eustasy, tectonics, and sedimentation.

Sequences are subdivided into "systems tracts," composed of smaller scale cycles ("parasequences"), and are deposited during different stages of a transgressive-regressive cycle. Three systems tracts are generally recognized within a sequence: 1) A "Lowstand Systems Tract" (LST) at the base of a cycle, which consists of progradational to aggradational strata deposited during a fall to early rise in relative sea level. The lowstand systems tract of a sequence is not always well preserved. 2) a "Transgressive Systems Tract" (TST), which is deposited during a rapid rise in relative sea level. This results in onlap of sea level and sedimentation onto the basal unconformity of a sequence; depositional is retrogradational. Condensed sections are common within the transgressive systems tract and the lower part of the overlying highstand systems tract. 3) "Highstand Systems Tract" (HST) forms during the late stage of a rise to the early stage of a fall in relative sea level, which results in deposition of aggradational, sediment-starved to progradational strata. The base of highstand deposits occur associated with a "surface of maximum flooding" that may be marked by a smaller scale, submarine unconformity during a period of extreme sediment starvation. Some workers (e.g., Brett and Baird, 1996) recognize "early" and "late" components of highstand systems tracts (EHS, LHS). These correspond to a shift from aggradational to progradational deposition within the later part of a depositional sequence.

Detailed field studies of upper Lower and Middle Devonian (upper Pragian-Givetian) strata in Pennsylvania and New York indicate that the succession is composed of at least 10 meso-scale ("third order") depositional sequences (Figure 28). Each sequence is bounded by basal unconformities or their correlative conformities at the base of the transgressive systems tract; as noted by Brett and Baird (1996), the lowstand systems tract of the sequences is rarely preserved. Lesser unconformities may form below and near to the base of the highstand systems tract.

Strata discussed in this paper comprise the lower part of Sloss's (1963) Kaskaskia Supersequence. The basal unconformity of DS2, developed in shallower parts of the basin, represents the Wallbridge Unconformity-supersequence boundary. The position of the Wallbridge Unconformity has been a source of debate over recent years, with some workers placing it at the top of the Oriskany-Ridgeley Sandstone. However, the top-Oriskany unconformity throughout the basin, from New York to southwestern Virginia, represents a surface of rapid transgression above the top of the sandstone, and not the position of a significant regression (Ver Straeten, 2001). The Wallbridge becomes amalgamated to an increasing number of younger, overlying sequence-bounding unconformities upon approaching the southern to northwestern margins of the Appalachian Basin.

**Upper Lower to Middle Devonian Sequence Stratigraphy**

***Depositional Sequence 1.*** As noted above, the margins of the basin are marked by development of the widespread Wallbridge Unconformity at the base of the Kaskaskia Supersequence. However, in deeper areas of the Appalachian Basin, deposition was continuous through the Pragian Stage of the Lower Devonian, and no unconformity is present. In those areas, time is represented by a mosaic of quartz arenites and generally cherty limestones (Oriskany and Glenerie-Shriver). As yet, there has been no systematic, basin-wide stratigraphic analysis of Pragian-age rocks, in part due to the difficulty of finding distinctive, time-significant marker beds in widespread, nearshore sandstone facies. A few studies, however, provide a basic outline of the development of what is herein termed Depositional Sequence 1.

Barrett and Isaacson (1981), in a study of the Oriskany Sandstone of western Maryland and West Virginia, reported a distinctive, depth-related cyclicity to changes in brachiopod assemblages. In that area of the basin, similar to parts of central and eastern Pennsylvania, the Oriskany Sandstone conformably overlies older strata. Building on previous analysis by Schuchert et al. (1913) and Woodward (1943), Barrett and Isaacson (1981) report an overall vertical shift from a relatively deeper subtidal *Acrospirifer* assemblage to a *Costispirifer*-rich assemblage of a shallower water aspect through the lower to upper Oriskany, reflecting a shallowing event during the early to late highstand systems tract. The faunal trends then undergo a reversal indicative of an additional deepening event that continues upward into overlying dark shales of the lower part of the Needmore Formation (=TST of Depositional Sequence 2). The initial deep-water environments, followed by the mid-Oriskany shallowing, is interpreted here to reflect the development of the post-Helderberg Depositional Sequence

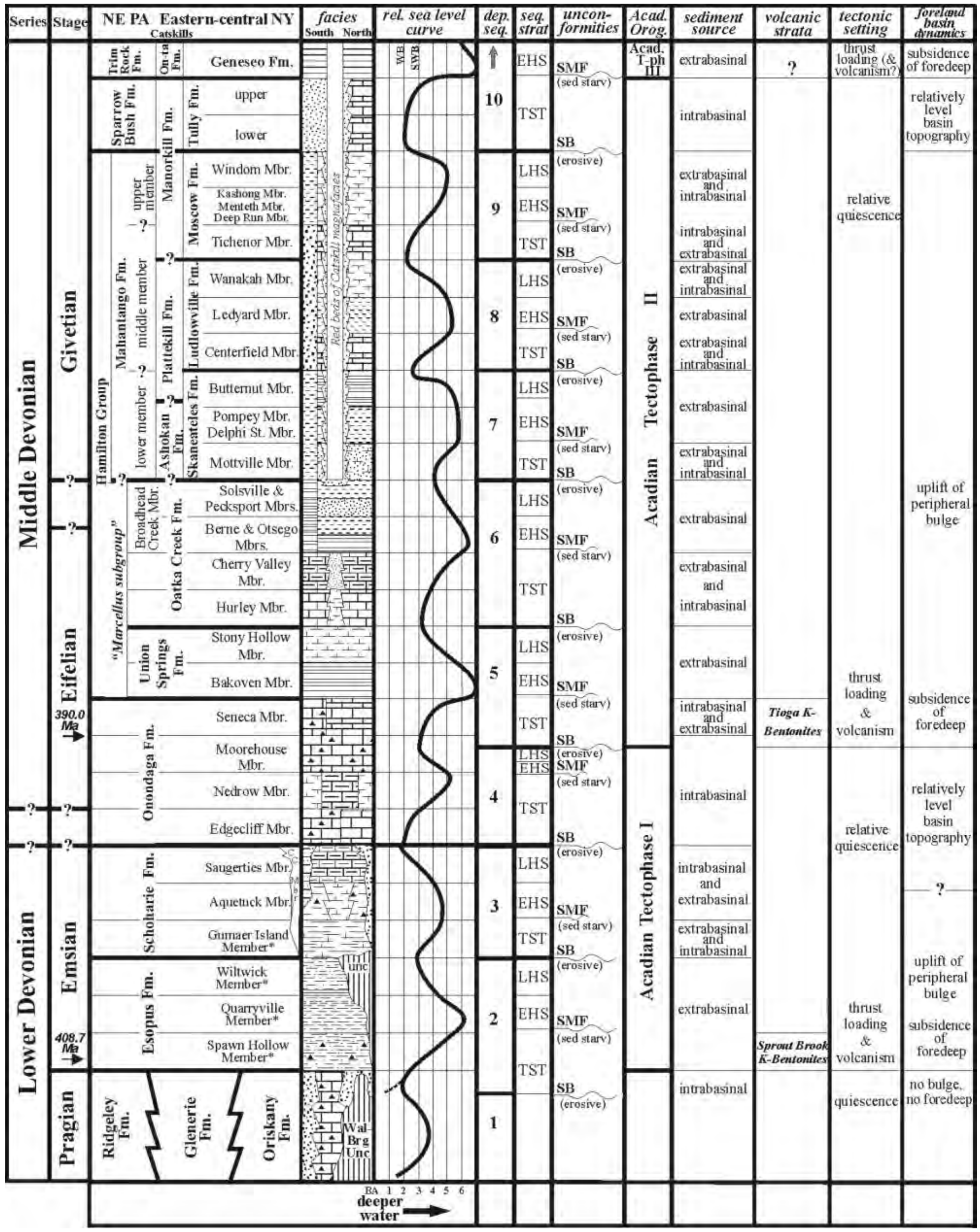


Figure 28. Stratigraphy, Sequence Stratigraphy, and the Acadian Orogeny, upper Pragian-Givetian, Eastern Pennsylvania and New York. Figure from left to right summarizes the age, stratigraphic nomenclature, lithofacies, relative sea level history, sequence stratigraphy, and effects of the Acadian orogeny relative to regional upper Lower and Middle Devonian strata (eastern PA and eastern to central NY). Note that correlations of the Mahantango Formation and Middle to Upper Hamilton Group strata between New York and Pennsylvania are very poorly understood at present. Asterisks mark informal, new member terms introduced in this paper. *Abbreviations:* Acad. Orog. = Tectophases of the Acadian orogeny; BA = Benthic assemblage zones 1-6 of Boucot (1975; 1=sea level, deepening to 6); CC = Carlisle Center Member; dep. seq. = depositional sequences; EHS = early part of highstand systems tract; LHS = late part of highstand systems tract; On-ta = Oneonta; SB = sequence boundary; seq. strat = systems tracts of depositional sequences; SL = sea level; SMS = surface of maximum flooding; SWB = storm wave base; Trim Rock = Trimmers Rock; TST = transgressive systems tract; WB = normal wave base. Modified after Ver Straeten, (1996a).

1. This follows with Johnson et al.'s (1985) interpretation of a significant T-R (transgressive-regressive) cycle through the bulk of the basinally deposited Oriskany Formation.

Again, the lack of detailed data at present hampers a fuller delineation of the development of Depositional Sequence 1. Much further work is needed in the strata of the Oriskany Formation and equivalents.

***Depositional Sequence 2.*** DS2 is composed of strata of the upper part of the Ridgeley-Oriskany (and equivalent) formations and the overlying Esopus Formation. Toward the margins of the basin, the Wallbridge Unconformity forms the basal bounding surface of the sequence. In deeper portions of the basin, including the Stroudsburg to Port Jervis area, the lowstand systems tract of DS1 is preserved. The transgressive systems tract consists of the uppermost Oriskany-Ridgeley Sandstones and possibly the Spawn Hollow Member (lower member) of the Esopus Formation. The surface of maximum flooding may occur at the base of the middle (Quarryville) Member; the exact position is unclear at this time. In a general sense, the middle, relatively fine-grained Quarryville Member of the Esopus Formation comprises the early highstand of DS1; an overall coarsening-upward trend through the upper part of the Quarryville and succeeding Wiltwick Members is associated with late highstand progradation of silt- to fine sand-sized siliciclastics.

The Esopus Formation is also divisible into three major coarsening-upward successions (=Spawn Hollow, Quarryville, and Wiltwick Members). Each commences with dark gray to black shale and culminates in bioturbated argillaceous siltstone or fine-grained sandstone. The tops of the Spawn Hollow and Quarryville members are capped by a sharply defined transgressive surface.

***Depositional Sequence 3.*** DS3 comprises strata of the Schoharie Formation (Gumaer Island, Aquetuck, and Saugerties Members). A basal sequence-bounding unconformity of DS3 erosionally truncates the underlying Esopus Formation in eastern to east-central New York (Ver Straeten, 1996). Again, a lowstand systems tract is not recognized at the base of the sequence. The Gumaer Island Member composes the transgressive systems tract of DS3. A common, if subtle, discontinuity at the base of the Aquetuck Member marks a transgressive surface closely below the surface of maximum flooding of DS2. Early highstand conditions characterize the Aquetuck Member; the maximum highstand of sea level appears to be represented by the widespread interval of dark shaly strata in the lower part of the member. A general shallowing upward trend through the upper part of the Aquetuck and Saugerties Members is indicative of late highstand conditions. Two medial scale cycles within DS3 are composed, respectively, of the Gumaer Island and Aquetuck-Saugerties Members; smaller-scale cycles are also a prominent feature of the Schoharie Formation in some areas.

***Depositional Sequence 4.*** The fourth post-Wallbridge sequence consists of the Edgecliff, Nedrow and lower to middle parts of the Moorehouse Members of the Onondaga Limestone. Throughout much of Pennsylvania and eastern New York the base of the sequence is conformable; a laterally equivalent erosive unconformity occurs across central to western New York. Lowstand conditions are not recognized, but may be found in the lower part of the Edgecliff Member, associated with initial growth of coral bioherms. The Edgecliff Member comprises the transgressive systems tract; a significant transgressive surface at the base of the Nedrow Member marks a sharp deepening that continues to a position of maximum flooding in the "black beds" at the top of the Nedrow Member. Lower to middle strata of the Moorehouse Member represent early to late highstand deposits of DS4.

Smaller-scale cycles in DS4 are widely recognizable now throughout the Pennsylvania and New York outcrop belts. Three medial-scale cycles in the Edgecliff Member are widely correlatable in the Appalachian Basin, as are two apparent cycles each in the Nedrow and Moorehouse Members. Finer, parasequence-scale cycles (~1 m-thick) reported in the Edgecliff Member in western to central New

York (Brett and Ver Straeten, 1994) are difficult to distinguish in the coarser, less differentiated, chert-dominated facies of the member of eastern Pennsylvania and eastern New York. Thin rhythmic cherty and non-cherty couplets in eastern Pennsylvania could be a manifestation of smaller scale cycles similar alternating limestone and calcareous shale cycles seen the in Schoharie Formation and in the Onondaga-equivalent Selinsgrove Limestone of central Pennsylvania.

***Depositional Sequence 5.*** DS5 is marked at its base by a gradational change from shallowing- to deepening-up lithologic and faunal trends; the succession is conformable along the Pennsylvania and New York outcrops, associated with an overall shallow ramp geometry of Onondaga strata across eastern North America. The laterally equivalent basal unconformity occurs in correlative shallower water deposits of the Columbus Limestone of central Ohio. Upper Moorehouse strata represent widely preserved lowstand deposits of DS5; fining-upward trends through the overlying Seneca Member indicate a rise in relative sea level (transgressive systems tract). The prominent unconformity at the Onondaga-Union Springs contact in New York, not seen but presumed present in the Stroudsburg area, represents a prominent transgressive surface, probably closely underlying the base of early highstand. This major surface, significantly, youngs to the west across New York, associated with progressive westward, collapse of the upper Onondaga platform driven by subsidence of the Appalachian foredeep during the onset of Acadian Tectophase II.

Early highstand deposits of DS5 are composed of black shales of the overlying Bakoven Member (Union Springs Fm.). Late highstand deposits, which are represented by calcareous shales to siltstones and sandstones of the Stony Hollow Member, first appear above a thin, mid-Union Springs Formation K-bentonite that is found basinwide. Different scales of shallowing-upward cycles are displayed through the Stony Hollow Member and equivalents across the Appalachian Basin.

***Depositional Sequence 6.*** Development of overlying DS6 is presently little known in Pennsylvania. In New York it comprises the Oatka Creek Formation. Fossiliferous thin limestones at the base of the sequence (Hurley Member) may alternatively be interpreted to be lowstand or lowest transgressive systems tracts (not the Cherry Valley Member, as reported by Ver Straeten, 1996a). In eastern New York DS6 is conformable, but the multiple unconformities low in the sequence erosionally truncate upper DS5 and lower DS6 strata across central to west-central New York.

The overlying part of the sequence is very poorly known outside of New York State at present. Black to dark gray shales of the Berne Member (Oatka Creek Formation) in eastern New York represent early highstand deposits of DS5. Thick overlying deposits of the Otsego Member and undefined upper Oatka Creek strata represent progradational infilling of the preserved eastern foredeep of the basin; uppermost strata of DS5 in eastern New York may or may not be are by fluvial-dominated strata of the Ashokan Formation (Rickard, 1975).

Three prominent medial-scale subsequences mark Depositional Sequence 6 in the New York succession. There comprise the Berne, Otsego, and the Solsville-Pecksport Members, respectively. Finer scale parasequences are represented in the upper Berne and lower Otsego Members by 3-8 m-thick successions of dark gray mudstones capped by thin shell beds as reported by Ver Straeten (1994). The cycle-capping shell beds represent sediment-starved flooding surfaces at the base of the parasequences.

***Depositional Sequences 7, 8, 9 and 10.*** Succeeding sequences 7-10 of the Middle Devonian Hamilton Group and Tully Limestone in New York have been discussed in detail by Brett and Baird (1996). The four sequences comprise, in New York, the Skaneateles, Ludlowville, Moscow, and combined Tully and Geneseo Formations of central to western New York. These depositional sequences are characterized by a basal limestone-sandstone unit that overlies a sequence-bounding unconformity (e.g., Stafford-Mottville, Centerfield, and Tichenor-Menteth Members, DS 7-8, respectively; Tully Formation, base of

DS 10). Flooding surfaces that cap the limestones are succeeded by dark, shale-dominated strata that in general coarsen upward to the base of the overlying sequence.

A high-resolution stratigraphy of the post-Oatka Creek Hamilton Group in New York has been the focus of many studies by Brett and Baird and co-workers. However, correlations between the New York and Pennsylvania Middle Devonian successions, even at the member-level, are still relatively poorly defined. Many of the key marker beds of the New York Hamilton are not as yet identified in Pennsylvania, with the exception of local recognition of the Centerfield coral bed, at the base of Sequence 8. A glance at the overall lithologic trends in Pennsylvania and New York, however, indicate overall similarities. As in central Pennsylvania, the Mahantango Formation of the Delaware Water Gap area features an overall coarsening up through the lower member to the sand-dominated middle Member. Similarly, in New York an overall coarsening up sequence marks the strata of the Skaneateles and Ludlowville Formations in the middle to upper middle Hamilton Group. In both states the coarsest sand-rich facies are then succeeded by a more shale-dominated strata (in New York, the Moscow Formation), overlain by the Tully Formation, or in northeastern Pennsylvania, the Tully-equivalent Sparrow Bush Sandstone.

Application of the sequence stratigraphic model and the high resolution stratigraphic methods should assist in refining correlations through the Pennsylvania succession and tying it to equivalent strata in both the northern and southern parts of the Appalachian Basin. Prave and Duke (1991) and Slattery (1995), working in strata of the Mahantango Sandstone in central Pennsylvania present some of the first sequence stratigraphic analyses of the Hamilton Group in Pennsylvania. Additional work by Brett and Baird continue the work, and have begun to link the Pennsylvania and New York Middle Devonian (e.g., Brett and Baird, 1996). Much work remains to be done, however.

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# THE SCHOHARIE FORMATION IN EASTERN PENNSYLVANIA

by
Charles A. Ver Straeten

## INTRODUCTION

Upper Emsian (upper Lower Devonian) rocks of the Schoharie Formation and its equivalents along the Appalachian Basin comprise a mixed clastic and carbonate succession, intermediate between underlying mudstones to fine terrigenous sandstones (Esopus Formation) derived from early uplift in the adjacent Acadian orogenic belt and overlying limestones (Onondaga Limestone) formed during a time of relative quiescence in the Acadian mountains. The formation was first named by Vanuxem (1840) for calcareous, fossiliferous, fine-grained sandstones in the hills above Schoharie, in eastern New York. The term was restricted to the Schoharie-Helderbergs area west of Albany, New York until Chadwick (1927) identified fine-grained argillaceous limestones and calcareous mudrocks as Schoharie strata in the Hudson Valley of eastern New York. The formation has since been recognized throughout eastern New York, northwestern New Jersey, and eastern Pennsylvania (Goldring and Flower, 1942; Johnsen, 1957; Inners, 1975; Epstein, 1984; Epstein et al., 1974; Ver Straeten, 1996a,b) and correlated into the Skunnemunk-Green Pond outlier in southern New York and northern New Jersey (Boucot et al. 1970; Ver Straeten, 1996a).

## BIOSTRATIGRAPHY AND AGE

The Schoharie Formation is assigned to the Lower Devonian *Eodevonaria arcuata* subzone of the *Amphigenia* brachiopod assemblage zone, the *Aemulophyllum exigum* coral assemblage zone (Dutro, 1981; Oliver and Sorauf, 1981). Conodont control on the Schoharie Formation is poorly constrained at present, and has been questionably placed within the upper Emsian *Polygnathus serotinus* conodont zone (Klapper, 1981). The problem is primarily due to the concentration of previous studies in relatively shallower water facies in New York State, which yield non-diagnostic *Icriodus* conodonts. A search for globally correlatable *Polygnathus* conodonts is presently underway by the author in deeper water facies from Emsian-age strata in the Needmore Formation in central Pennsylvania.

No datable K-bentonites are presently recognized within Schoharie and correlative strata in the Appalachian Basin. Zircons from one of the Sprout Brook K-bentonites low in the Esopus Formation in New York yielded an age of 408.3 + 1.9 Ma (Tucker et al., 1998); the Tioga B K-bentonite bed at the base of the Seneca Member of the Onondaga Limestone has been dated at 390 + 0.5 Ma (Roden et al., 1990). The Schoharie Formation lies midway through that succession.

## SCHOHARIE FORMATION

In eastern Pennsylvania the Schoharie Formation is comprised largely of calcareous mudstones and siliceous siltstones to fine sandstones. At the chief reference section in the Stroudsburg area (U.S. 209 roadcuts, Buttermilk Falls), the formation totals 30.3 m in thickness (Ver Straeten, unpublished data; see Figure 29). Two subdivisions are recognized in the Stroudsburg region (Inners, 1975); a lower massive, dark gray, pyritic, calcareous mudstone with common *Zoophycos* or *Chondrites* traces and an

Ver Straeten, C. A., 2001, The Schoharie Formation in eastern 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. 54 - 60.

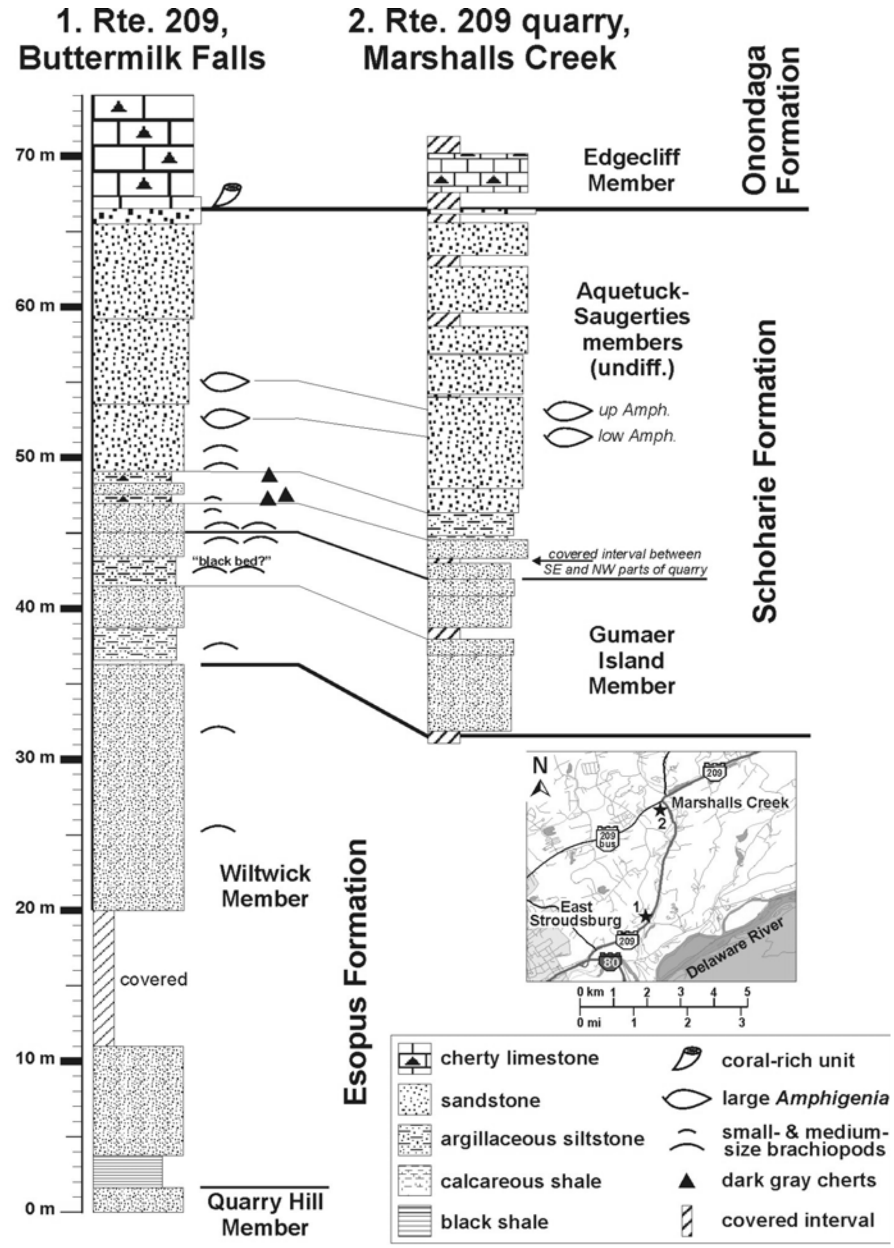



Figure 29. Stratigraphy of the Schoharie Formation along US 209 near Stroudsburg, PA. Correlated sections of the Schoharie Formation at US 209 cuts adjacent to Buttermilk Falls and quarry along US 209, 3 miles to the north, at Marshalls Creek (see Inners and Ver Straeten, this guidebook, p. 61. For lithologic key, see Figure 21 of additional paper by Ver Straeten, this guidebook, p. 35.

upper massive, dark gray, predominantly siliceous mudstone and siltstone to sandstone unit with vertical burrows and abundant small nodules that are at least in part of a phosphatic composition. The lower subdivision contains a low-diversity brachiopod fauna of *Atlanticocoelia*, *Coelospira*, *Acrospirifer*, and *Eodevonaria* (Inners, 1975). The upper subdivision grades upward through a distinctive finer-grained, more argillaceous interval with scattered chert and a fauna of small brachiopods to increasingly coarser strata above. Two thin calcareous beds in the middle of the upper unit (11.25 m and 12.85 m below top of Schoharie) feature abundant brachiopods (including large *Amphigenia*) and small rugose corals; additional fossils are found through upper strata of the formation.

In the Hudson Valley of eastern New York, post-Esopus and pre-Onondaga strata were variously identified as the Schoharie Formation (Johnsen, 1957; Inners, 1975) or the Carlisle Center (lower) and Schoharie (upper) Formations (Rickard, 1975). Recent study shows that strata of the type Carlisle Center in its type area (east-central New York) represent undifferentiated Schoharie strata, and do not underlie it as previously interpreted. Therefore the term "Carlisle Center Member" will be restricted to undifferentiated Schoharie-equivalent sandy facies in central New York. The lower of three post-Esopus units in the Hudson Valley is here informally assigned to Gumaer Island Member, which is overlain by the previously established Aquetuck and Saugerties Members.

The upper part of Figure 30 shows the correlation of the Schoharie Formation and equivalent strata in Pennsylvania and New York. The lower unit of the Schoharie in the Stroudsburg area is the lateral equivalent of the Gumaer Island Member, overlain by the undifferentiated Aquetuck-Saugerties Members (see further discussion below)

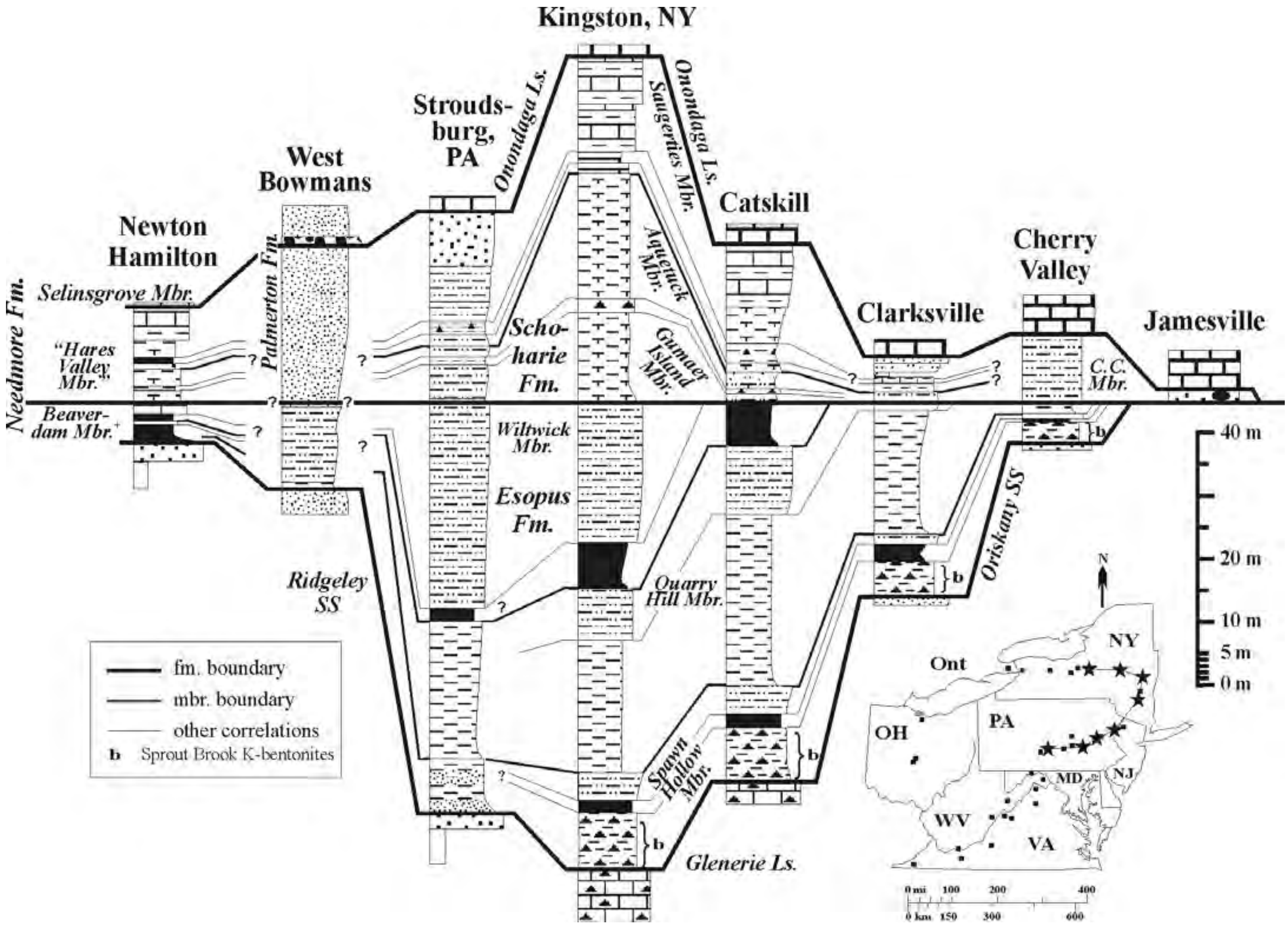



Figure 30. Correlation of Emsian-age Schoharie and Esopus formations and equivalent strata, Pennsylvania and New York. Figure shows the correlation of time-significant markers in the Esopus and Schoharie formations in New York and Pennsylvania. The Esopus Formation represents a "third order" depositional sequence, and is composed of three members (Spawn Hollow, Quarryville, and Wiltwick), each of which represents separate subsequences. The Schoharie Formation comprises a second depositional sequence; the Gumaer Island and Aquetuck-Saugerties members comprise two subsequences within the formation. As shown (Newton Hamilton), the Beaverdam Member and basal thin limestones of the "Hares Valley Member" are the lateral equivalent of the Esopus Formation. Datum = base of Schoharie Formation.

Southwest of Stroudsburg the Schoharie Formation is reported to grade into a lower unit of interbedded, fossiliferous, light-weathering, thin shales and cherty siltstones ("Schoharie Fm." of Epstein et al., 1974, formerly "Bowmansville Chert") and overlying lower to middle strata of the Palmerton Sandstone (Inners, 1975; Ver Straeten, 1996b). Precise relationships between Stroudsburg and the Palmerton-Lehigh Gap area are presently unclear to the author, with the exception of the position of the Schoharie-Onondaga contact (see below). Preliminary work indicates that the so-called Schoharie in the Palmerton area is probably not time-correlative with Schoharie-age rocks, but represents upper strata of the Esopus Formation (Figure 30).

In central Pennsylvania, the Gumaer Island and combined Aquetuck-Saugerties Members, along with other distinctive units, can be directly correlated into the second and third to fourth subdivisions of the "Hares Valley Member" of the Needmore Formation, respectively (Figure 30; also see Figure 22 of additional paper by Ver Straeten, this guidebook, p. 38). These units can be further correlated into the Virginia and West Virginian outcrop belt, where they pass into upper cherty strata of the Huntersville Formation (Ver Straeten, 2001).

## PALMERTON FORMATION

Not exposed in the Stroudsburg area, but wedging in to the southwest (southern Monroe, Carbon, and Schuylkill Counties), are sandstones of the Palmerton Formation, or Sandstone (Swartz and Swartz, 1941). The unit comprises a localized body (ca. 65 km long x 8-11 km wide, E-W by N-S; Inners, 1975) of mature, quartz sand-rich strata. The Palmerton Sandstone consists of generally massive bedded, dark to light gray, medium- to coarse-grained quartz arenite with minor quartz pebble conglomerate (Epstein et al., 1974). Thickness of the Palmerton Sandstone ranges from a feather's edge south of Stroudsburg (northeast) and in central Schuylkill County (southwest) to a maximum of approximately 43 meters at Little Gap, northeast of Palmerton (Swartz and Swartz, 1941; Inners, 1975). The unit is relatively unfossiliferous. The lower contact of the Palmerton Sandstone with the underlying “Schoharie Formation” (sensu Epstein et al., 1974) is variably gradational to sharp; Epstein (1984) reports an erosional relief of up to 0.8 meter at the contact in south-central Monroe County.

The Palmerton Formation is overlain all along the outcrop belt by strata assigned to the Onondaga Limestone (Epstein, 1984). However, study of both formations in the Palmerton-Lehigh Gap area shows that the upper part of the Palmerton, from the base of a prominent meter-thick conglomerate bed, is the lateral equivalent of the Edgecliff Member of the Onondaga Limestone (Figure 30). Argillaceous strata overlying the Palmerton Sandstone represent the shaly Nedrow Member of the Onondaga, succeeded by more carbonate-rich strata of the Moorehouse and Seneca Members.

For further discussion of the Palmerton Formation see Sevon (1968, 1970), Epstein et al., (1974), Inners (1975), Epstein (1984), Ver Straeten (1996a,b).

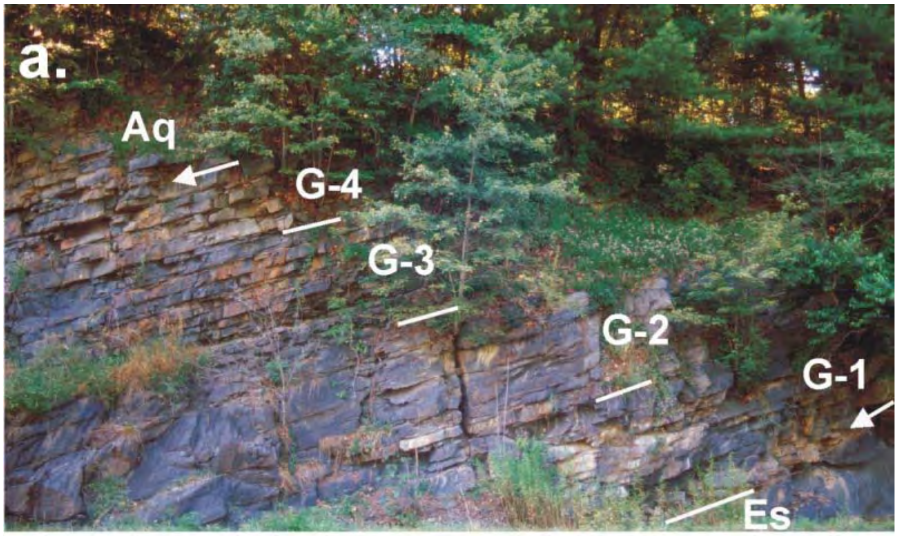
## SCHOHARIE FORMATION ALONG US 209 AT BUTTERMILK FALLS

New detailed examination of a complete section of the Schoharie Formation in roadcuts along US 209 at Buttermilk Falls on Marshalls Creek permit correlation of several distinct, widespread, time-significant marker beds into the eastern Pennsylvania outcrop belt (Figures 29 and 30). At Buttermilk Falls, the Schoharie Formation is comprised of 30.3 meters of predominantly very fine to fine-grained sandstones and siltstones with lesser amounts of clastic mudstones. It overlies dark gray, *Zoophycos*-burrowed very fine- to fine-grained sandstones and siltstones of the upper member of the Esopus Formation. The lower part of the Schoharie Formation is slightly calcareous, becoming increasingly siliceous up section. Elongate dark gray chert nodules occur only in one interval; other small nodules previously interpreted to be chert scattered though the middle to upper part of the section (Epstein, 1984) appear, at least in part, phosphatic in composition (black to dark gray weathering through light blue to white).

Subtle, but significant differences in lithology, bedding thickness, and macro- and trace fossil content permit subdivision of Schoharie strata at the US 209 exposure at Buttermilk Falls. Some of the units within the formation can now be correlated out of the area into the classic Schoharie of eastern New York and equivalent strata in central Pennsylvania.

Dark gray, non-calcareous, thick to massive bedded siltstones to fine sandstones of the Wiltwick Member (top of the Esopus Formation) show prominent *Zoophycos* pin-striping throughout. The high degree of bioturbation probably contributes to the massive appearance of the strata. Scattered uncommon *Atlanticocoelia* are found through the member. The Esopus-Schoharie contact, at the base of the Gumaer Island Member, is marked by a thin, recessive-weathering interval of olive-gray calcareous shale with *Chondrites* trace fossils.

Four subdivisions can be discerned within the approximately 8.7 m-thick Gumaer Island Member along US 209 (see Figure 31a): 1) a lower, 2 m-thick interval of thin- to medium-bedded argillaceous siltstones, with a few thin shaly crevices. A number of *Atlanticocoelia* brachiopods occur approximately 1.4 meters above the base of the unit. 2) More resistant, medium- to thick-bedded, dark gray siltstones with *Zoophycos* traces and scattered *Atlanticocoelia*s. 3) A more recessive-weathering interval of finer-grained, argillaceous siltstone beds. A concentration of *Atlanticocoelia* shells occurs about 6 m above the Gumaer Island base. Bedding shows an initial thinning upward to a prominent crevice approximately 6.5 m above the base of the member, followed by a gradual upward thickening of the beds to; 4) a more resistant ledge of medium-bedded siltstones that continue to thicken upward to a 50 cm-thick bed. Units 1-2 and 3-4 represent two separate small-scale transgressive-regressive cycles superposed over the medial-scale cycle that comprises the Gumaer Island. The third unit is correlative with the key, cherty "black bed" marker of the member in the Hudson Valley (eastern NY; Figure 30). The "black bed" interval represents the deepest water facies of the member all across the northern and central part of the Appalachian Basin. Throughout the Hudson Valley, the lower and upper contacts of the Gumaer Island Member feature scattered white quartz pebbles. None were found at US 209 after extensive search; one or two have been found at the top of the Gumaer Island equivalent in central PA outcrops (Newton Hamilton), however.



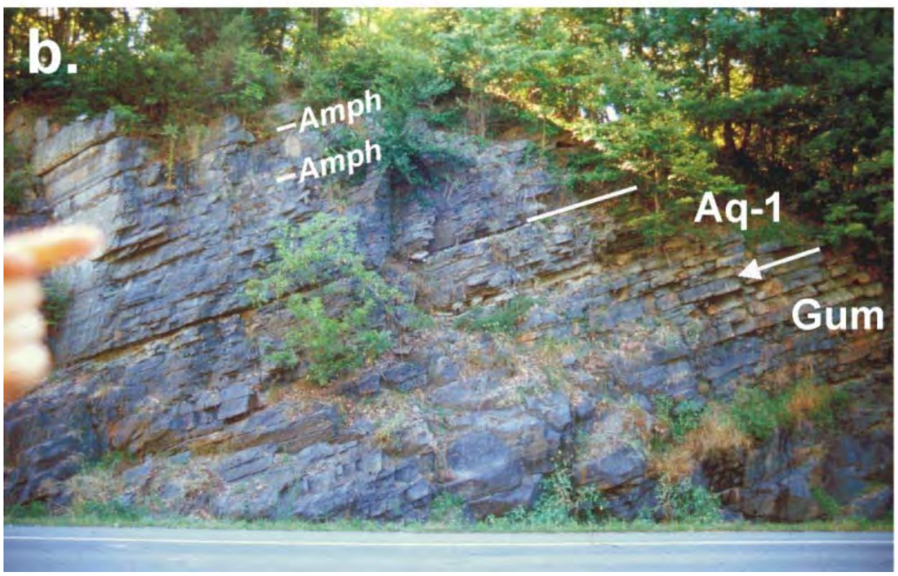




Figure 31. Lower to middle parts of the Schoharie Formation, US 209 at Buttermilk Falls.

a. Photograph of uppermost Esopus Formation, and the Gumaer Island and basal Aquetuck-Saugerties Members of the Schoharie Formation. Arrows point to formation/member contacts. Units G-1 to G-4 discussed in text. Total thickness of Gumaer Island Member = 8.7 meters.

b. Photograph of upper Gumaer Island and lower Aquetuck-Saugerties members. Arrow points to the members contact. Finger points to lower of two locally correlatable beds with large *Amphigenia* brachiopods in the section (marked *Amph*). Thickness of Aq-1 unit, between the white lines = 2.3 m.

The transition into strata of the undifferentiated Aquetuck and Saugerties Members (Figure 31b) is marked by another turnaround in proxies related to relative water depth (bedding thickness, grain size, and macro- and trace fossil trends) (ca. 8.7 m above the base of the Schoharie Formation). This shift corresponds to similar trends at the Gumaer Island-Aquetuck contact throughout eastern New York and in correlative strata in central Pennsylvania. Decreasing grade in all of these indicators (< bed thickness, < grain-size/ > argillaceous content, < size in macrofaunal brachiopods, a gradational change from *Zoophycos* to small *Chondrites* ichnofacies) 8.7 m above the formational contact indicates a second major deepening event at the base of an Aquetuck-Saugerties transgressive-regressive cycle. The position of a more rapid transition occurs approximately 11 m above the base of the formation, where *Atlanticocoelia* is replaced by small chonetid (?) brachiopods, and *Zoophycos* disappears. The succeeding two meters mark the position of lowest grade in the four factors noted above for the entire Aquetuck-Saugerties interval. Elongate dark gray chert nodules (Figure 32b) occur at three positions through the interval, which is capped by a prominent crevice approximately 12.9 m above the Esopus Formation. The entire finer-grained interval is correlatable to a distinctive argillaceous interval in the Aquetuck Member at Kingston, NY and to a darker, cherty interval in the member at Catskill. In central Pennsylvania it is represented by black shales in submember C of the "Hares Valley Member" of the Needmore Formation (e.g., Newton Hamilton; Figure 30).

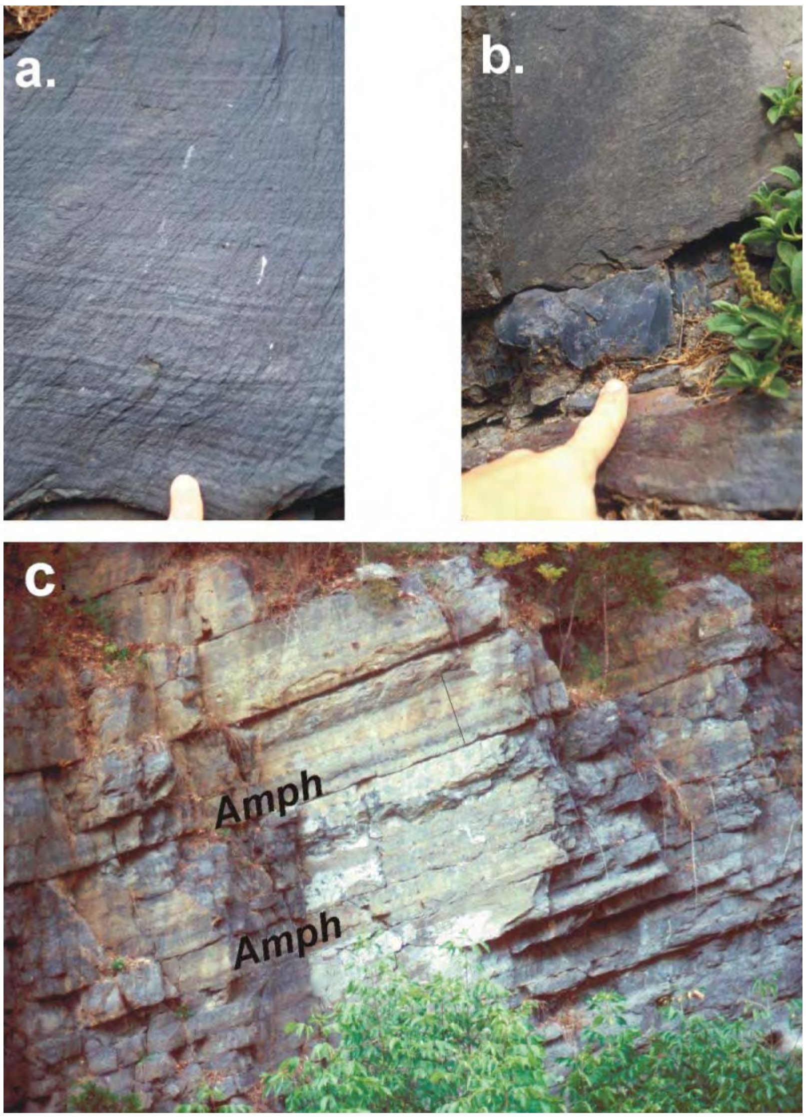



Figure 32. Strata of the Esopus and Schoharie Formations, US 209 at Buttermilk Falls

a. Photograph of horizontal *Zoophycos*-"pinstripes" from the upper part of the Esopus Formation. Finger for scale;

b. Closeup photograph of fine-grained unit of the lower part of the Aquetuck-Saugerties Members, showing chert nodule in front of finger;

c. Photograph of middle part of the Aquetuck-Saugerties Members, showing position of the two *Amphigenia* (*Amph*) beds and characteristic thick bedding.

All four environmental proxies show a transitional shift to increasing grade above the 12.9-m crevice (Figure 32c), marked in part by an initial return to *Atlanticocoelia* and *Zoophycos* facies, transitional to more diverse medium to large brachiopod-dominated facies. Two beds, at approximately 17 and 18.1 m feature large *Amphigenia* brachiopods, diagnostic of Schoharie-age strata across eastern North America. Macrofaunal changes are correlated with changes in the trace fossils assemblages, reflected in increasing vertical burrows upward through the upper Schoharie.

Generally increasing grain-size and bedding thickness again correlate with the paleobiologic proxies, all indicative of overall shallowing upward through the combined Aquetuck-Saugerties members. In the type area of the two members (Hudson Valley, eastern New York) a shift from nodular to bedded limestones marks the member contact; no known event bed marks the facies transition between the two members. Therefore, no attempt is made to distinguish the two in the Stroudsburg area. As the strata become increasingly siliceous up through the section, the two calcareous *Amphigenia* beds make good marker beds locally; it is not known if they have a greater regional significance, or if they have any association with the members contact in eastern New York.

Uppermost strata of the Schoharie Formation along US 209 at Buttermilk Falls are characterized by meter-plus thick massive bedding, fine-grained sandstones, and vertical trace fossils. A shift from sandstones to overlying limestones of the Edgecliff Member (Onondaga Limestone) is readily apparent along the outcrop.

See Inners and Ver Straeten, this guidebook, p. 61 for discussion of Schoharie stratigraphy at Riccobono's quarry on US 209, about 3 miles to the north (see Figure 29).

# RICCOBONO'S "QUARRY IN THE SCHOHARIE": STRATIGRAPHY, OPERATIONS, AND MASS MOVEMENT

by
Jon D. Inners and Charles A. Ver Straeten.

## INTRODUCTION

The Route 209 Enterprises ("Riccobono's") quarry along US 209 in Marshalls Creek (Figure 33) exposes a nearly complete section of the Lower Devonian Schoharie Formation, the main bedrock aggregate source of the Stroudsburg area (Figure 34; see Ver Straeten, this guidebook, p. 54, Figure 29). Stratigraphically, the section extends from the dark, intensely cleaved, slightly calcareous siltstones of the Gumaer Island Member in the small inactive pit at the south end up through the lighter-gray, well-jointed, highly calcareous siltstones of the undivided Saugerties/Aquetuck Members in the main pit. The site also provides an instructive example of hillslope mass-movement that may help to explain an observation of I. C. White (1882) concerning the prevalence of huge limestone blocks in the glacial till of this part of Monroe County. From a DEP permit standpoint, this pit at the present time is not technically a "quarry"—rather it is a site-development operation. Removal of rock is incidental to modification of the site to erect a commercial operation, in this case "Alaska Pete's Trading Post." About six months ago, a formal application for a DEP mining permit was submitted to cover plans to expand the pit beyond the original site-development area. For simplicity's sake, the operation will from here on be referred to as a "quarry" even though it will be such only after the mining permit is approved.

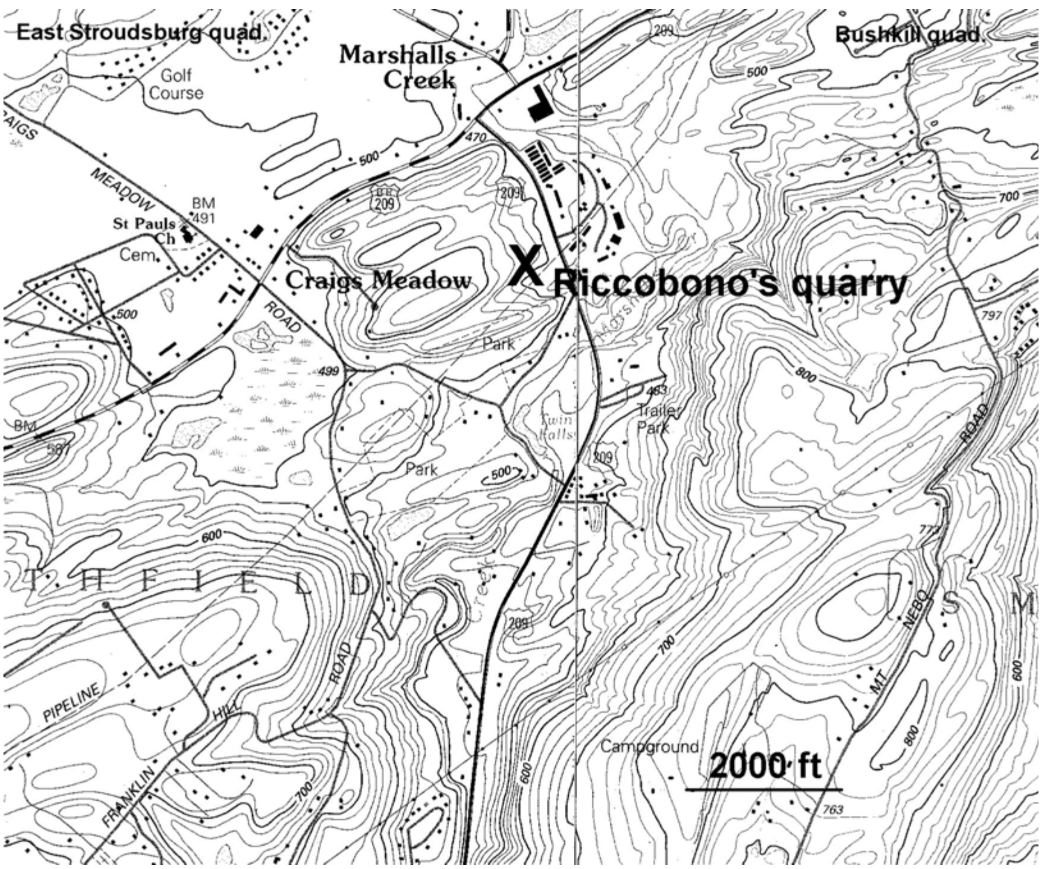



Figure 33. Location map of Riccobono's "Quarry in the Schoharie" along US 209 in Marshalls Creek.

Route 209 Enterprises is a subsidiary of Haines & Kibblehouse (H&K) Materials, Skippack, PA, whose other operations include the Locust Ridge quarry (Monroe Co., Catskill sandstone), the West Mountain quarry of Scranton Materials (Lackawanna Co., Spechty Kopf sandstone), and the

Inners, J. D. and C. A. Ver Straeten, 2001, Riccobono's "quarry in the Schoharie": stratigraphy, operations, and mass movement, *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. 61 - 67.

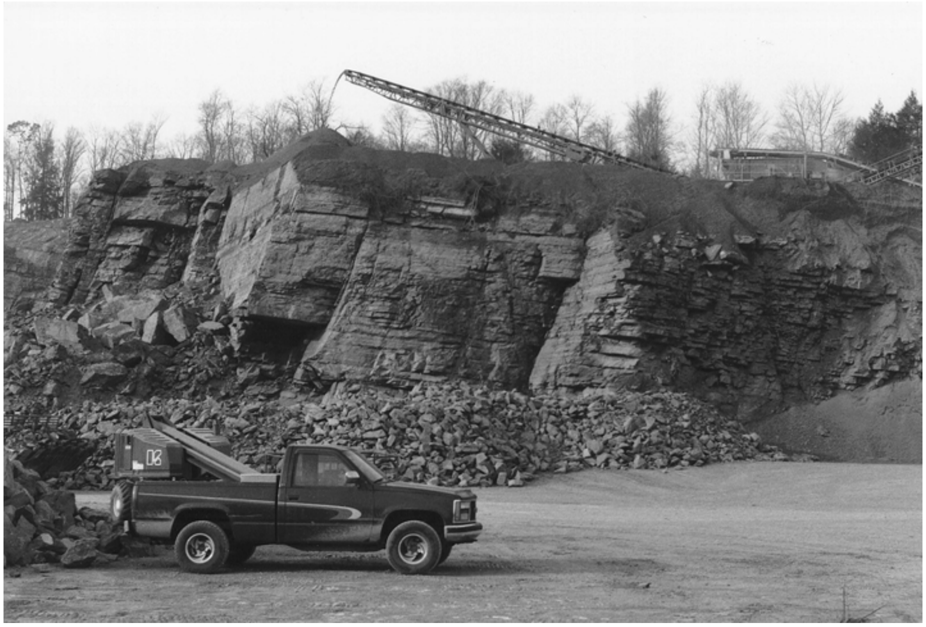


Figure 34. View of Riccobono's quarry looking west, showing well jointed siltstone of the Saugerties/Aquetuck Members, undivided, of the Schoharie Formation. Attitude of joints is approximately N65°E/75°SE.

Blooming Glen Materials quarry (Bucks Co., Lockatong argillite and Brunswick sandstone) (Barnes, 1997). The landowner and developer is Frank Riccobono of Northpark Development Corporation, Marshalls Creek, PA. Mr. Riccobono and Kirk Knecht, the foreman, provided the information on quarry operations discussed below.

## STRATIGRAPHY

This working quarry exposes much of the Schoharie Formation as it is developed in the Stroudsburg-Delaware Water Gap area. Lower cherty limestones of the Edgecliff Member of the Onondaga Limestone are also visible at the top of the section, above a thin covered interval that hides the basal coral-rich bed of the Onondaga Limestone. The quarry exposure, on the west side of US 209, is located 2 miles north of the road cut near Buttermilk Falls (see Ver Straeten, this guidebook, p. 54, Figure 29).

A total of approximately 108 feet (33 m) of the Schoharie Formation is exposed in two areas of the quarry (see Ver Straeten, this guidebook, p. 54, Figure 29). Near US 209, southeast of the main pit, over 36 feet (11 m) of dark-gray, slightly calcareous siltstones to fine-grained sandstones of the lower Schoharie Formation (Gumaer Island and lower part of Saugerties/Aquetuck Members) are exposed (Figure 35a). The outcrop is less weathered than the outcrop adjacent to Buttermilk Falls. The strata appear highly bioturbated (including the trace fossil *Zoophycos caudigalli*), and feature scattered *Atlanticocoelia* brachiopods and some small rugose corals. Two more resistant, very fine-grained sandstones occur 15.7-20.0 feet (4.8-6.1 m) and 30.0-32.8 feet (9.15-10.0 m) above the base; the top of the lower bed forms a prominent platform in the southeastern quarry. These represent the caps of the two intra-Gumaer Island cycles noted two miles to the south at the US 209-Buttermilk Falls exposure. The interval immediately above the lower sandstone, largely covered by talus at the platform level, correlates with the Gumaer Island "black bed" of eastern New York.

Almost the entire Saugerties/Aquetuck Members, undivided, (76.1 feet, or 23.2 m, thick) is well exposed in the long-weathered, south-facing wall in the northwest part of the quarry (Figure 35b). Despite the fact that a broad area of cover separates the southeastern and northwestern parts of the quarry, comparison of the sections with that of the cut near Buttermilk Falls indicates that little of the strata are missing. The widely correlatable fine-grained unit of the Aquetuck Member occurs low in the section, 4.1-9.8 feet (1.25-3.0 m) above the base. The lower and upper *Amphigenia* marker beds seen three miles south (US 209 cuts at Buttermilk Falls) were located approximately at 26.2 feet (8.0 m) and 32.5 feet (9.9 m) above the base, respectively. The general increasing grade of the four environmental proxies (lithology/grain-size, bedding thickness, macrofauna and trace fossils) reported for the middle to upper Saugerties/Aquetuck at Buttermilk Falls display the same trends in the quarry, reflecting overall shallowing through the members to the top sandstone bed of the Schoharie and/or the basal coral-rich limestone bed of the Onondaga Limestone.

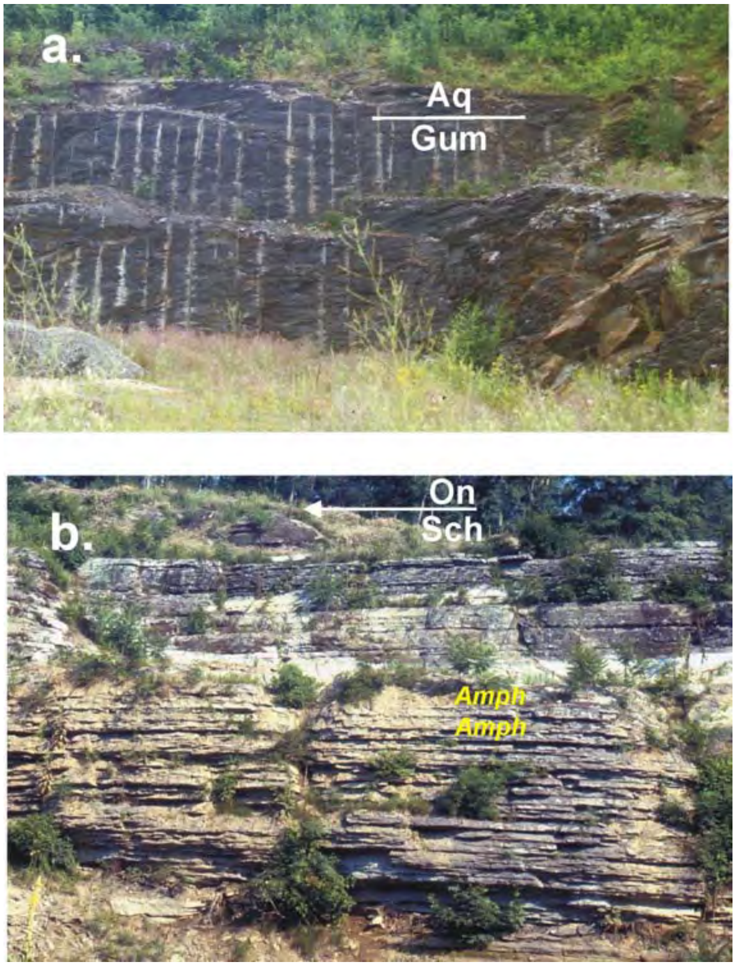



Figure 35. The Schoharie Formation in Riccobono's quarry.

a. Photograph of the flat-lying Gumaer Island and lower part of the Saugerties/Aquetuck Members in the southeastern area of the quarry. White line, at 33 feet (10 m) above base of section, separates the members.

b. Photograph of the Saugerties/Aquetuck Members in the northwestern area of the quarry, showing the position of the two *Amphigenia* (*Amph*) marker beds. Lower strata of the Onondaga Limestone (not seen) overlie the Schoharie Formation at top. Schoharie section totals 76.1 feet (23.2 m).

## QUARRY OPERATION

Site modification and removal of stone began in 1999, the intent being to clear and level about 20 acres. The final developed area will extend back in an east-west direction about 600 feet from the edge of US 209, a 2:1 slope formed by blasting down of the highwall reaching back to the final limit of the pit (approximately at the present tree line). The blasting contractor on the job is Explo-Tech out of Stewartsville, PA, near Reading. About eight acres are now being worked. It will take about a year and a half to mine out the currently active area (a big east-west, partially joint-bounded block along the north side). By the time this block is removed, the DEP mining permit should have been approved—and the quarry can then be expanded beyond its original limits. With enlargement of the quarry, the crushing plant will be moved to the north, and site development (i.e., the construction of "Alaska Pete's") will proceed in the old quarry area. Total cost of the project is about $20 million, much of which will be recouped by aggregate sales.

The Route 209 Enterprises operation has three crushers: the **primary** breaks material down to 8 in. or less, the **secondary** to 3 to 4 in. or less, and the **tertiary** to 1/4 in. or less (Figure 36). Oversize blocks are broken down by jackhammers to provide feed for the primary crusher, which processes about 300 tons/hr. The bulk of production goes to township, county, and state road construction. Aggregate sizes produced are 2A (road material and driveways), 2B (concrete and septic systems), and 1B (blacktop). Grit from screenings also goes into blacktop, and six- to eight-inch surge material is used for road-base and drainage. The aggregate passes all PennDOT requirements for base course, subbase, and wearing course, and has an excellent skid-resistance rating.

The siltstone feedstock is very tough material. As a result, the die-plates on the primary crusher have to be changed every 3 or 4 months. While this is a high rate of wear, stone at some other H & K quarries is much more abrasive. At the Silver Hill quarry (Lancaster Co., Hammer Creek argillite), die-plates must be changed every month.

When the quarry first opened, it was thought that the planned relocation of US 209 in the near future (due to begin in early 2002) would provide a big market for the stone. Current plans for the new highway, however, call for on-site crushing of excavated rock (mostly "Helderberg" limestone) for road aggregate rather than bringing in rock from outside sources. PennDOT conducted an intensive test-drilling program in late 2000-early 2001 to prove the feasibility of this approach.

Figure 36. Operations at Riccobono's quarry, showing the various crushers and numerous piles of processed stone. The primary crusher is in the center of the photo.

## MASS MOVEMENT

On the south side of the main ridge that is here being quarried away is a splendid example of joint-controlled, bedrock mass movement (Figure 37). Several blocks of Schoharie siltstone (Saugerties/Aquetuck Members)—the largest of which is 14 feet wide (north-south), 68 feet long (east-west), and about 10 feet high—have moved out along one of the prominent N65$^{o}$E-striking joint up to 11 feet from the solid ledge of the ridge (Figure 38a, b.). The crevasse thus formed by movement of the largest block was originally filled with blocky rubble containing some rounded clasts, probably a mixture of colluvium and colluviated glacial till (Figure 39). Some of this material has been removed during the present quarry operation, exposing the walls of the crevasse. To the west of the well-exposed blocks are at least three more partially concealed by vegetation; these have moved successively farther (up to 30 ft) to the south (Figure 38a). It is likely that the entire slope for several hundred feet to the west contains such detached blocks.

Figure 37. Mass movement of large joint blocks of Schoharie siltstone at southwest end of main quarry. The open crevasse is the result of removal of original fill material during quarry operations (see Figure 39).

The whitish-weathered siltstone ledge directly above the crevasse is smoothed and polished and exhibits glacial striae trending S60$^{o}$W, parallel to the ridge (Figure 40). Striae on top of the ledge higher on the hill trend S28$^{o}$W, the dominant ice-flow direction in the late Wisconsinan. The walls of the crevasse, however, are not striated. In fact, the north wall—formed by the in-place ledge—is covered with quartz-mineralized patches and was clearly not scoured by the glacier (Figure 41).

The absence of striae and the presence of undisturbed mineralization in the crevasse indicate that it is mainly a postglacial feature, though slight outward shove along the joint may have been initiated by late-stage glacial movement. Gravity and periglacial frost action combined to cause most of the outward movement, as well as a slight downslope rotation of the main displaced block. As the block shifted away from the bedrock ledge, the widening crevasse was filled with debris colluviating down the hillslope.

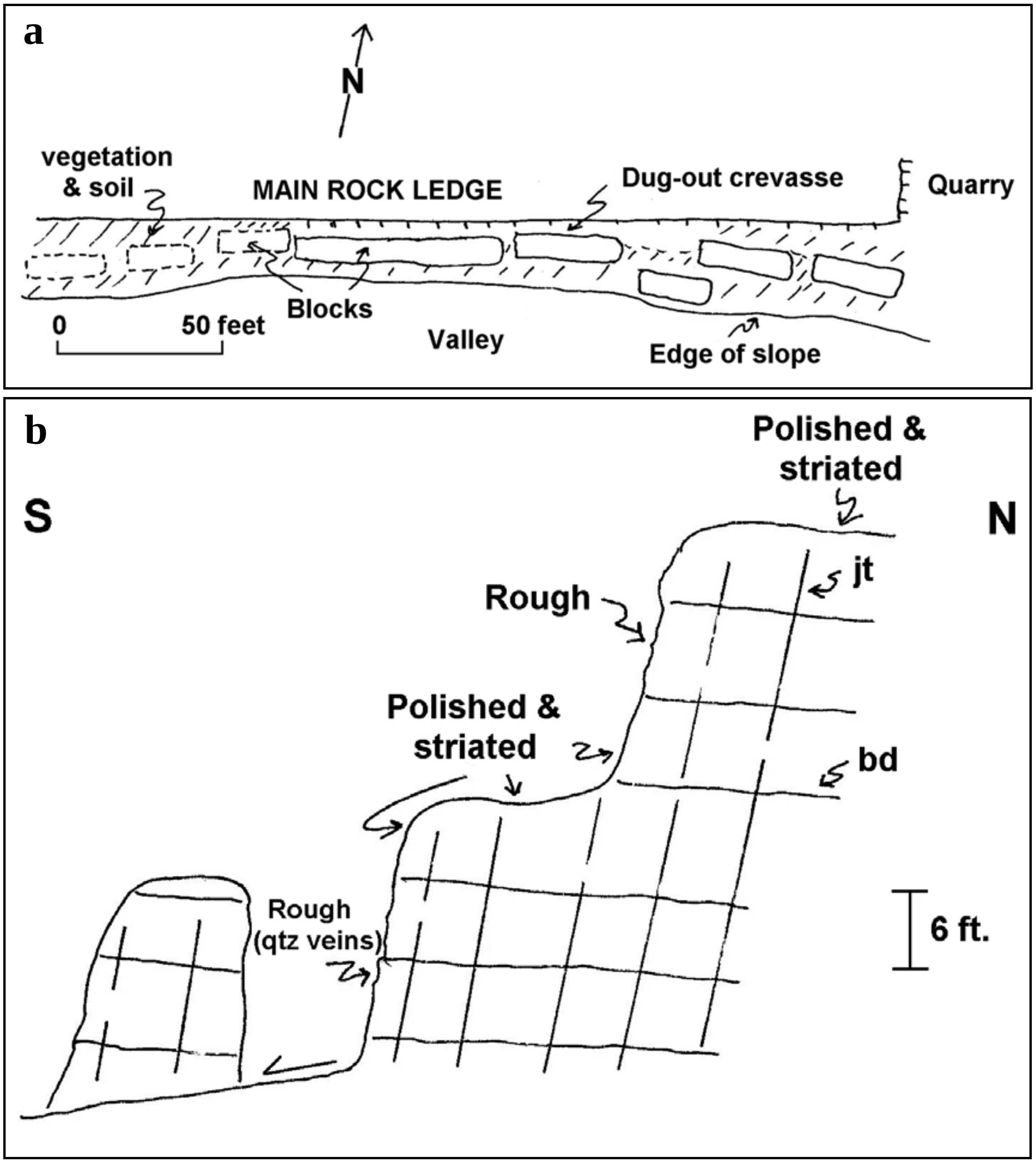



Figure 38. Generalized diagrams of mass movement in Schoharie siltstone at Riccobono's quarry:
**a**. map view; **b**. cross section.

In describing glacial erosion in the area, I. C. White (1882, p. 44-48) notes the abundance of "immense" ("many…as large as a good sized house") joint-blocks of "Cauda-Galli Grit" (Schoharie-Esopus) and "Corniferous Limestone" (=Onondaga) that have been glacially quarried from the slopes of Wallpack Ridge. One of these blocks is noted on the Day-2 road log at mile 9.1, less than a mile and a half southwest of Riccobono's quarry (Figure 42). It is probable that erosion of these large blocks was facilitated in many cases by "pre-glacial" mass movement similar to the post-glacial movement observed here.

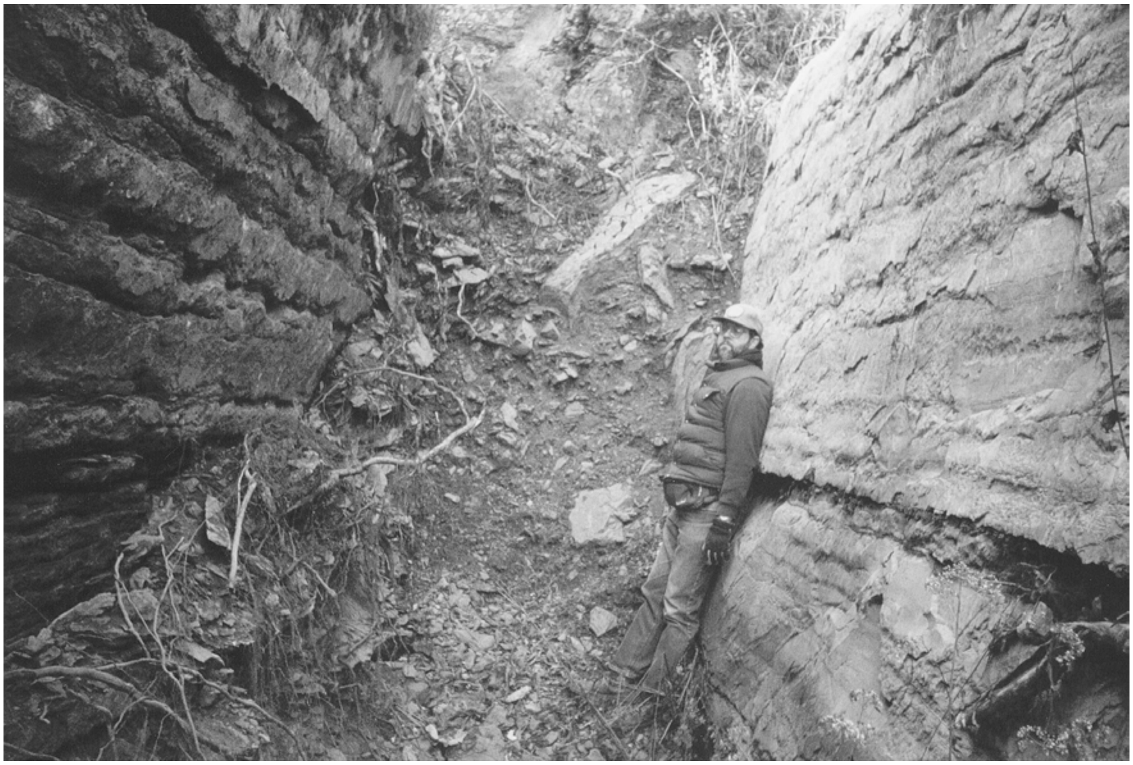


Figure 39. Blocky rubble (colluvium and colluviated glacial till) exposed at west end of crevasse. This material was removed from the crevasse for a considerable distance east of this point during quarry operations.

Figure 40. Glacially polished and striated siltstone ledge on north wall of crevasse. Striae trend about S60°W, parallel to the quarry ridge.

Figure 41. Glacially polished and striated surface just above crevasse (top of photo) and mineralized, unglaciated surface within crevasse (bottom of photo)—main siltstone ledge on north wall of crevasse.

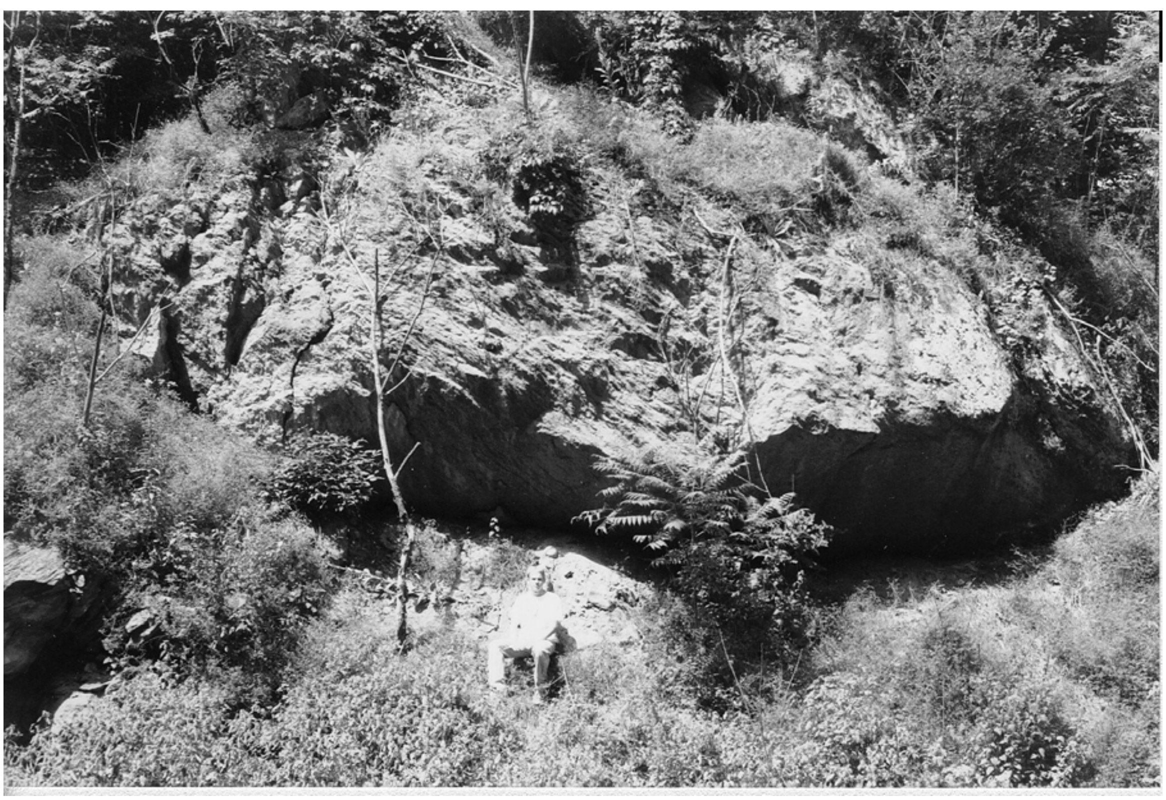


Figure 42. “Immense” erratic boulder of Onondaga Limestone (Edgecliff Member) on west side of US 209, 1.35 mile south of Riccobono’s quarry (see further discussion at mile 9.1 of the Day-2 road log).

# BIOSTRATIGRAPHIC DETERMINATION OF THE BASAL MARTINSBURG FORMATION IN THE DELAWARE WATER GAP REGION

by

David C. Parris, Louise F. Miller, and Richard Dalton

ABSTRACT

The Jacksonburg-Martinsburg contact, a gradational boundary, may now be determined with greater biostratigraphic precision. It now is correlated to the *Corynoides americanus* Subzone by comparisons with a shelly fauna obtained in association with graptolites from two localities. So far as can be determined, there is no age overlap between the Martinsburg Formation and any portion of the Jutland Sequence, the latter consisting entirely of older Ordovician rocks.

## INTRODUCTION

Considerable progress has been made during the past two decades in the graptolitic facies of the Ordovician System in New Jersey and adjacent areas. Despite the challenges of collecting in such metamorphosed and tectonized rocks, a significant number of graptolite localities have been found (Parris and Cruikshank, 1992). Age determinations for a number of localities and sequences have gradually accumulated (Parris et al., 1993, 1998). These discoveries contribute to a framework from which the stratigraphic and tectonic development of the region may be interpreted.

Here we report a previously unpublished locality, of which the fauna provides a precise age for the Jacksonburg-Martinsburg boundary. The nature of the lithostratigraphic contact is gradational, from the predominantly carbonate Jacksonburg Formation upwards to the predominantly clastic Martinsburg Formation. The venerable work of Weller (1903) established a detailed faunal sequence for the Jacksonburg Formation which was little improved upon in the following century, and which has proven its value once again in the current study. Considering the limitations of time and transportation available to Weller, his work was of excellent quality. He discovered the first significant graptolite localities in New Jersey and would doubtless have found many more if time had been sufficient. He had only limited success in the Hudson River Slates (now the Martinsburg Formation), but established that graptolite facies predominated there and enabled basic correlations.

Precise correlations delimiting the base of the Martinsburg Formation have been difficult to establish because of the predominantly graptolitic facies. The underlying Jacksonburg Formation has always produced substantial shelly fossils of course, which did enable a maximum age and zonation to be determined. What has been lacking is a site with shelly and graptolitic material together, the uncommon kind of locality that enables cross correlations, as does the Swatara Gap site higher in the Martinsburg Formation (Wright et al., 1977). The site we report here, however, begins to fulfill that need. Discovered some years ago, but never before comprehensively described, the site has produced new collections that effectively provide the desired correlations.

We gratefully acknowledge the field assistance of Donald Monteverde and Richard Volkert, and the kind permission of the J. Fritz family, owners of the site. Shirley S. Albright and Robert Ramsdell assisted with our identifications of taxa, greatly improving them. E. J. Reimer assisted with cataloguing and data entry. We thank Dr. Heyo Van Iten for precise determination of the conulariid specimen.

Parris, D.C., et al., 2001, Biostratigraphic determination of the basal Martinsburg Formation in the Delaware Water Gap region, *in*, Inners, J. D. and Fleeger, G. M., eds., 2001—a Delaware River odyssey, Guidebook, 66$^{th}$ Annual Field Conference of Pennsylvania Geologists, Shawnee-on-Delaware, PA, p. 68 - 71.

## SITE DESCRIPTION AND METHODS

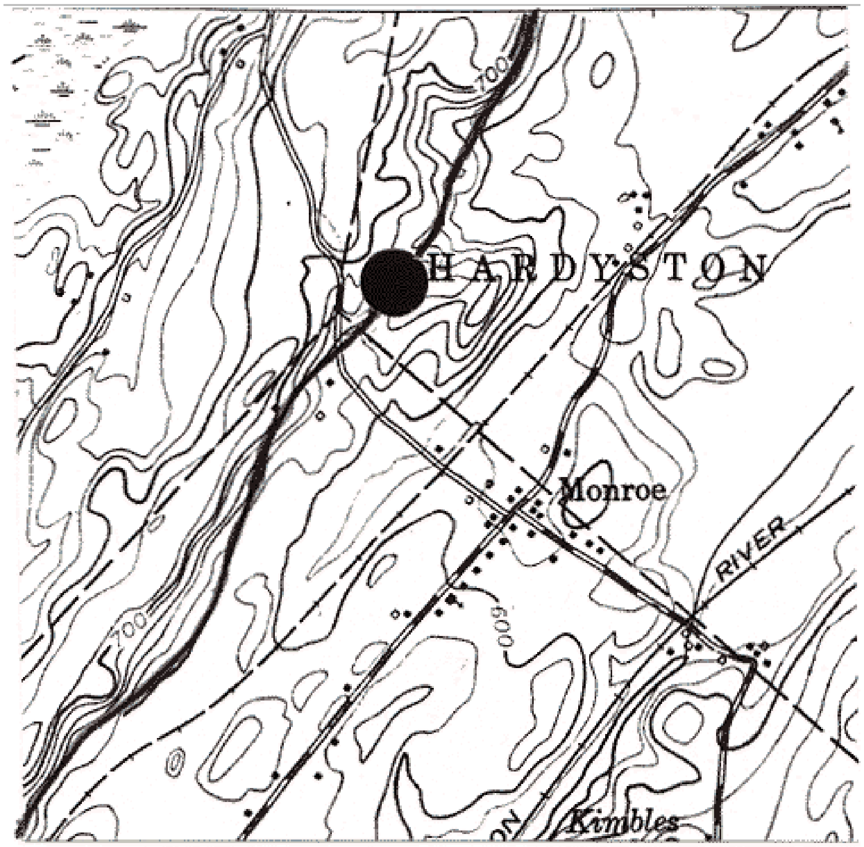



Figure 43. Portion of Newton East 7.5' quadrangle, indicating Monroe Locality.

The site is here designated the Monroe Locality, Om-200 of the New Jersey State Museum survey notes. Although located near the common corner of Hardyston, Lafayette, and Sparta Townships in Sussex County, New Jersey, the outcrop from which all fossils were retrieved is entirely within Hardyston Township (Figure 43). Although known for some time to one of us (R.D.), previous collecting efforts had not resulted in any comprehensive study. Renewed collecting efforts were begun in 1998, and continue to the present. All collecting has been by surface examination of weathered material derived from the outcrop. The stratum is a fine-grained meta-sandstone less than 10 meters above the underlying Jacksonburg Formation. The slaty cleavage is at a low angle to the bedding planes (>30$^{o}$); thus the weathered specimens are quite easily found and are satisfactory for identification. The approximately 200 identifiable specimens thus far recovered include both shelly and graptolitic material, and may be designated as the Monroe Faunule. All specimens are deposited with the New Jersey State Museum and catalogued as shown (prefix NJSM). Identifications were carried out using the Museum's comparative collections, with the exception of the one conulariid specimen, which also was sent to a specialist for refined identification.

## RESULTS AND DISCUSSION

The resulting faunal list (Table 2) includes a number of taxa that are particularly useful for correlation. In particular, *Orthograptus amplexicaulus* (Hall) is the name-giver of the Zone to which we correlate this site (Figure 44). It has large rhabdosomes with distinctive thecal spacing and overlap, and thus is easily identified. *Glyptograptus euglyphis* Lapworth also is a graptolite species of considerable stratigraphic utility. The genus is readily recognized by its gently sigmoid thecae (Bulman, 1970), and the species by its long narrow rhabdosome and thecal spacing (Ruedemann, 1947). It ranges from the *Glyptograptus teretiusculus* Zone up to the *Corynoides americanus* Subzone (biostratigraphy of Berry, 1960, 1968, 1970, 1971; Finney, 1982, 1986). This taxon is also present at the Middleville (Om-22) and Port Murray (Om-50) Localities, both of them low in the Bushkill Member of the Martinsburg Formation. The Port Murray Locality has been the most precisely dated site in all the graptolite-dominated facies of the area (Parris et al., 1993), and is correlated to the *Corynoides americanus* Subzone. It has a graptolite fauna that also includes *Climacograptus*, and thus is essentially identical to the Monroe Faunule in its graptolite taxa. This would appear to give a very sound correlation of the Monroe Faunule (and thus the site and the basal Martinsburg Formation) to the *Corynoides americanus*

GRAPTOLITE ZONE CORRELATIONS
(After Ross et al, 1982)

| Correlations: British | Correlations: American | Zones | Subzones |
|---|---|---|---|
| Caradoc | Mohawkian | 13. *Orthograptus amplexicaulus* | *Climacograptus spiniferus*<br>*Orthograptus ruedemanni*<br>*Corynoides americanus* |
| | | 12. *Climacograptus bicornis* | |
| | Whiterockian | 11. *Nemograptus gracilis* | |
| Llandeilo | | 10. *Glyptograptus cf. G.teretiusculus* | |
| Llanvirn | | 9. *Paraglossograptus etheridgei* | |
| | | 8. *Isograptus* | |
| | | 7. *Didymograptus bifidus* | |
| Arenig | Ibexian | 6. *Didymograptus protobifidus* | |
| | | 5. *Tetragraptus fruticosus* (3 and 4-branched) | |
| | | 4. *Tetragraptus fruticosus* (4-branched) | |
| | | 3. *Tetragraptus approximatus* | |
| Tremadoc | | 2. *Clonograptus* | |

Figure 44. Summary of graptolite zonation applicable to Ordovician of New Jersey, originally published in Parris et al., (1998).

Subzone. In the absence of information to the contrary, the dominance of *Glyptograptus euglyphus* in a basal Martinsburg Formation site now seems to signify correlation to the *Corynoides americanus* Subzone, rather than the *Climacograptus bicornis* Zone, the expectation of our previous investigations (Parris et al., 1998). Neither of the two name-givers of these zones has been found in our area, however.

The trilobite genus *Flexicalymene* is considered a guide fossil lineage for the Middle and Upper Ordovician. The genus, with its characteristic lateral lobes on the glabella, is readily recognized. The species *Flexicalymene senaria* (Conrad) has been recorded from the Jacksonburg Formation (Weller, 1903) and is widespread in North American faunas of the Mohawkian Stage (Trenton and Shermanian correlatives of previous usage). The degree to which it has evolved toward the characteristic morphology of its hypothetical descendant, *Flexicalymene meeki* (Foerste) should be a reliable stratigraphic marker. Alternatively, evolutionary progression toward the morphology of the species *Flexicalymene granulosa* (Foerste), identified in the Swatara Gap fauna (Wright et al., 1977), may be utilized, as it would appear from Foerste's original descriptions/discussions that it too is a descendant of *Flexicalymene senaria* (Foerste, 1909, 1919). Our specimens are essentially identical to morphology of *Flexicalymene senaria* as described by Foerste (1910) and to the specimens cited by Weller (1903) in the Jacksonburg Formation faunas, the latter directly compared by us.

The conulariid *Conularia* cf. *trentonensis* (Hall) is a particularly interesting element of the fauna, conulariids being uncommon in any fauna (Albright, 1995). However, the same species has been found in the Swatara Gap Fauna (Wright et al., 1977), which is high in the Martinsburg Formation. It therefore is not especially useful for detailed biostratigraphy.

Gastropods of the Monroe Faunule are not well preserved and are only tentatively identified here, but may be more useful for correlation when more and better specimens are found. The few specimens found at the Middleville Locality Om-22 (Parris and Cruikshank, 1992) may now be referred to cf. *Lophospira* sp. (NJSM 12595) and cf. *Sinuites* sp. (NJSM 12749), identifications which may also be refined with further collecting and examination. The other taxa thus far collected from the Monroe locality are also found within the immediately underlying Jacksonburg Formation and are either very long-ranging forms or not precisely identified, or both. It is to be hoped that further collecting will bring additional correlation taxa, notably of graptolites and trilobites.

The correlation of the basal Martinsburg Formation to the *Corynoides americanus* Subzone gives a somewhat further emphasis to the lack of chronostratigraphic overlap between rocks of the Jutland Klippe and the Martinsburg Formation. Any common interval would presumably have been within the *Climacograptus bicornis* Zone, but no portion of the Martinsburg Formation seems to correlate to that Zone.

Table 2. Fauna of Monroe Locality (Om-200), Hardyston Township, Sussex County, New Jersey.

| Taxon | NJSM Catalogue Number |
|---|---|
| *Conularia* cf. *trentonensis* (Hall) | 19577 |
| cf. *Prasopora* sp. | 20153 |
| *Plectorthis plicatella* (Hall) | 20154 |
| *Orbiculoidea sp.* | 20156 |
| cf. *Sinuites* sp. | 20168 |
| cf. *Hormotoma* sp. | 20167 |
| *Michelinoceras* sp. | 20157 |
| *Flexicalymene* cf. *senaria* (Conrad) | 20158 |
| cf. *Pterygometopus* sp. | 20159 |
| Crinoidea, indeterminate | 20160 |
| *Glyptograptus euglyphus* Lapworth | 20161 |
| *Orthograptus amplexicaulus* (Hall) | 20169 |
| *Climacograptus* cf. *mohawkensis* (Ruedemann) | 20155 |

# GEM OF THE MIDDLE DEVONIAN:
# THE "CENTERFIELD FOSSIL ZONE" AT BRODHEAD CREEK

by
Denise Wilt

***ABSTRACT***

The "Centerfield fossil zone" of the Mahantango Formation (Middle Devonian) in northeastern Pennsylvania has long been recognized as a faunally rich zone. It was first correlated to the Centerfield Member of the Ludlowville Formation in New York by Willard (1936) on the basis of its position, physical character, and fauna. The Centerfield Member in New York is recognized as a time-stratigraphic unit laterally continuous across New York. The Centerfield fossil zone as currently recognized in Pennsylvania does not correlate physically nor lithologically to the Centerfield Member in New York because of its varied stratigraphy and the discontinuous nature of its occurrence. The outcrop on PA 191 near Brodhead Creek offers a full exposure of the Centerfield fossil zone and its fauna. The fauna of the Centerfield fossil zone is similar to that of the Centerfield Member. As in New York, a faunal cyclicity also exists in the outcrops of the Delaware Water Gap area. The cyclicity represents a transgressive sequence in which rocks that are very lithologically similar (fine to medium siltstone) change faunally from a brachiopod-dominated assemblage to a series of coral-dominated assemblages.

## INTRODUCTION

Fossils of the "Centerfield fossil zone" in Pennsylvania have been collected by many, but there has been limited research to interpret the paleoecology and distribution of faunal associations within the unit. The Centerfield Limestone of western New York contains cyclic faunal associations that have been interpreted as a regressive-transgressive cycle (Savarese, 1984; Savarese et al., 1986). If the Centerfield fossil zone of Pennsylvania is indeed equivalent to the Centerfield of New York, it can be inferred that the sea level fluctuations were taking place contemporaneously throughout the northern and eastern Appalachian Basin. This might indicate that regional, perhaps tectonic, driving forces were behind the sea level fluctuations.

## BACKGROUND

### Stratigraphic Setting

The Mahantango Formation along with the Marcellus Formation constitutes the Middle Devonian Hamilton Group in Pennsylvania. The formation represents some of the earliest siliciclastic deposits of the west-northwestward prograding clastic wedge of the Acadian orogenic York Highlands (Faill, 1985). The Mahantango Formation (Givetian) represents four million years of deposition (Harland et al., 1989).

The Mahantango Formation was first recognized as the Cadent Shales by Rogers (1858). The Second Geologic Survey of Pennsylvania replaced Cadent with Hamilton for the sequence of rocks between the Marcellus shale and the Tully limestone (White, 1882). Willard (1939) introduced the name Mahantango for the previous "Hamilton" and placed it with a Hamilton Group along with the Marcellus Formation. He also attempted to correlate the Mahantango in eastern Pennsylvania with the units of the Hamilton Group recognized in New York. He proposed that the three formations of the New York Hamilton Group, the Moscow, Ludlowville, and Skaneateles, were discernible faunally in northeastern Pennsylvania as facies within the Mahantango Formation.

Wilt, Denise, 2001, Gem of the Middle Devonian, the "Centerfield fossil zone" at Brodhead Creek, *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. 72 - 80.

The Mahantango in northeastern Pennsylvania is composed primarily of massive, nonbedded silty shales. The unit coarsens to the northeast, containing subordinate siltstones and sandstones. To the south, in the Lehighton and Palmerton quadrangles there is only one mapped unit of siltstone, the Nis Hollow Siltstone. Four fossil zones have been mapped by Epstein et al. (1974), the Little Gap, Kunkletown, Centerfield, and Tully Fossil Zones. The first three fossil zones contain brachiopods, tabulate and rugose corals, bryozoans, and crinoid columnals.

## The Centerfield Fossil Zone

White (1882), the first to identify the Centerfield fossil zone, incorrectly identified it as Tully. Later it was recognized as a separate fossil-rich zone by Prosser (1895), Willard and Cleaves (1933), and Willard (1936). The Centerfield fossil zone as it is currently recognized was first correlated to the Centerfield Member in New York on the basis of its position, physical character, and fauna. According to Willard, the zone is found as a narrow band in the lower Ludlowville Facies, extending westward from Pike County through Monroe, Carbon, Schuylkill, Lebanon, and Dauphin Counties and finally disappearing entirely in Perry County. In eastern Pennsylvania it reaches a maximum thickness of 20 feet (6.1 meters) and thins westward to 3 or 4 feet (0.9-1.2 meters). Along with a decrease in thickness there is also an accompanying loss in coral abundance. Willard estimated that the shoreline at the time of formation was located 30 miles (48 km) east and southeast of the Stroudsburg area.

Beerbower (1957) studied the Brodhead Creek outcrop near Stroudsburg and found three faunal associations, the *Zonophyllum, Douvillina,* and *Protoleptostrophia* associations. He attributed changes in these to possible changes in substrate or rates of sedimentation. Beerbower and McDowell (1960) further analyzed random bulk samples and stratigraphically controlled samples from Pike, Monroe, and Carbon counties. They found two associations, one dominated by corals and bryozoans and the other dominated by brachiopods.

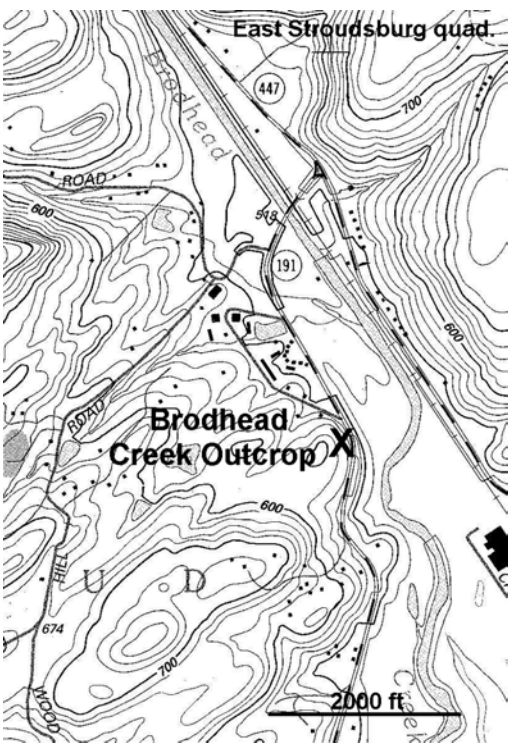


Figure 45. Location map of the Brodhead Creek outcrop of the Centerfield fossil zone in Stroud Township, Monroe County.

Caramanica (1968) carried out a detailed analysis of the coral paleontology and paleoecology of the Centerfield fossil zone in northeastern Pennsylvania. He found it contains three stratigraphically and laterally restricted biostromes that contain coral faunas similar to those of the Traverse Group of Michigan and the Hamilton Group of New York. Maximum faunal diversity was seen in the southwest near Stroudsburg where there is the least amount of coarse sediment in the unit. In the more coarsely clastic (northeastern) part of his study area, branching colonial and solitary corals are dominant types, whereas no particular form dominates the southwest area near Stroudsburg.

## BRODHEAD CREEK OUTCROP DESCRIPTION

The Centerfield fossil zone is exposed on PA 191, 0.7 miles (1.1 km) south of the intersection of PA 191 and PA 447 (41° 02' N, 75° 12' 30" E, East Stroudsburg quad; Figure 45). According to Caramanica (1968), the Centerfield fossil zone at Brodhead Creek is inferred to be 920 feet (280 meters) above the Mahantango-Marcellus contact and 1770 feet (540 meters) below the Mahantango-Trimmer's Rock contact.

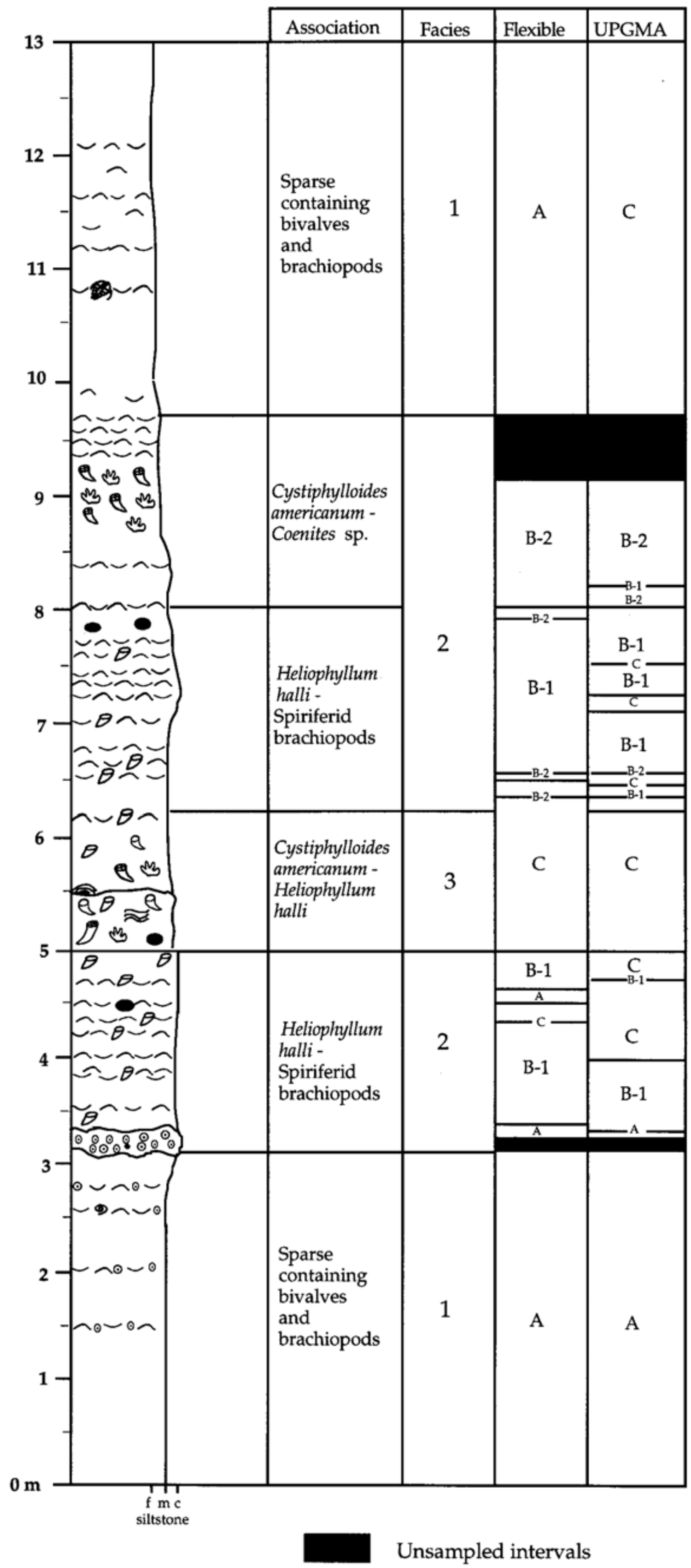



Figure 46. Brodhead Creek outcrop with associations, facies, and cluster analysis results.

The Centerfield fossil zone is exposed completely in 13 meters of this outcrop (Figure 46). There is lithological variation from fine to medium siltstone, and there are few sedimentary structures. The main variation within the fossil zone is in the composition of the fauna found within these siltstones. The top and bottom of the fossil zone are both composed of similar lithologies and contain similar faunas. There is a transition to a coarser and more fossiliferous siltstone at 3.1 meters. At 5.0 meters there is an abrupt increase in the density of fossil material, which then decreases over the following four meters. The fossil zone has been divided into three Facies representing these changes in lithology and faunal content.

In the following description of the Centerfield fossil zone the Facies are numbered basally from one upwards. The described Facies reoccur at other outcrops in northeastern Pennsylvania (see Wilt, 1999) in varying stratigraphic sequences.

**Facies 1**

Facies 1 is exposed twice in the Brodhead Creek outcrop, the lower Facies occurrence is exposed from the base of the outcrop to 3.1 meters and the upper Facies occurrence is exposed from 9.7 to 13.0 meters. It comprises fine shaly siltstone devoid of sedimentary structures except for concretions and layers of shell debris. The exposed rock weathers reddish brown. Pyrite films cover casts and molds of fossils. The lower and upper occurrences of Facies 1 vary in faunal composition, the lower occurrence being less fossiliferous and containing more concretions. In the lower occurrence (below 3.1 meters) the following species are found: *Lingula* sp., *Tropidoleptus carinatus, Mucrospirifer mucronatus, Nuculoidea* sp., *Paleoneilo* sp., *Modiomorpha* sp., *Platystoma* sp., *Fenestella* sp., and unidentified crinoid columnals. The brachiopods and bivalves are disarticulated but valves are generally intact. They occur mainly on single bedding planes in both convex and concave positions with no apparent orientation of the valves. Disarticulated crinoid columnals occur on bedding planes associated with brachiopods and bivalves. Articulated columnals that are several centimeters long are present in the interval from 2.9 to 3.0 meters.

The upper occurrence (above 9.7 meters) contains a similar, but more robust, fauna composed of *Coenites* sp. (in lower part of the subunit), *Rhipidomella* sp., *Tropidoleptus carinatus, Mucrospirifer mucronatus, Mediospirifer audaculus, Fimbrispirifer* sp., *Elita* sp., *Atrypa* sp., *Cypricardella* sp., *Nuculoidea* sp., *Fenestella* sp., *Anastomopora* sp., *Phacops rana*, and *Greenops boothi*. Brachiopods and bivalves are disarticulated and occur both in convex and concave positions. They occur mainly on single bedding planes with no distinct orientation. One individual of the trilobite *Greenops boothi* was found in an enrolled position. The crinoids occur both articulated and disarticulated. Disarticulated specimens are mainly found on single bedding planes along with disarticulated shell material. The majority of corals are small and mainly oriented parallel to bedding planes.

**Facies 2**

Facies 2 occurs from 3.1 to 5.0 meters and from 6.2 to 9.7 meters. This Facies is distinguished by shell concentrations that occur throughout the medium siltstone. The transition from Facies 1 to 2 is marked by an increase from fine to medium siltstone that begins with a 10-cm-thick bed of crinoid-rich shell material. This crinoid-rich bed contains small (1 cm diameter) horizontal burrows. It also contains rare small dark gray clasts that range from 3 to 5 mm in diameter. From 3.2 to 3.8 meters shell concentrations with crinoid debris throughout the matrix are found in discrete layers. Fauna is sparse, with only a few corals and brachiopods seen. This includes indeterminate rugose corals and brachiopods as well as *Rhipidomella* sp., *Pentamerella* sp., *Megastrophia* sp., *Mucrospirifer mucronatus*, *Elita* sp., *Nucloidea*, sp., *Paleoneilo* sp., *Pterinopectin* sp., and *Fenestella* sp. At 3.8 meters a 2-cm-thick shell concentration occurs. From 3.8 to 4.0 meters bedding is massive and contains *Heliophyllum halli, Spinocyrtia granulosa, Mediospirifer audaculus*, *Athyris* sp., and an indeterminate brachiopod. In this interval disarticulated crinoid columnals are abundant and dispersed throughout the sediment. Brachiopods are disarticulated and corals oriented parallel to bedding.

The interval between 4.0 and 4.35 meters is lithologically similar to previous intervals. Shell concentrations as well as corals (*Heliophyllum halli*) are found from 4.1 to 4.2 meters. The fauna of this interval includes *Heliophyllum halli, Leiorhynchus* sp., *Protoleptostrophia* sp., *Mucrospirifer mucronatus*, and *Cypricardella* sp. The brachiopods are mostly disarticulated, and the corals are oriented parallel to the bedding plane.

From 4.35 to 4.9 meters more corals are present with coral diversity increasing from 4.5 to 4.9 meters. The medium-gray siltstone is massive and has a honeycombed appearance due to the weathering of corals along the outcrop face. The corals are primarily found in discrete layers and appear disturbed, altered from their life position.

The fauna from 4.35 to 4.70 meters is composed of *Stereolasma rectum*, *Heterophrentis* sp., *Heliophyllum halli, Favosites* sp. aff. *Favosites placenta, Favosites nitella, Pleurodictyum dividua, Syringopora maclurei, Rhipidomella* sp., *Mucrospirifer mucronatus, Mediospirifer audaculus, Elita* sp., *Athyris* sp., *Atrypa* sp., indeterminate bivalve species, *Pterinopectin* sp., *Phacops rana*, and *Fenestella* sp. From 4.7 to 5.0 meters the corals are present in discrete layers in the massive, medium-grained siltstone. From 4.85 to 5.0 meters the corals are also found dispersed within the sediment and appear to be relatively undisturbed from their original life position. The fauna is composed of *Heliophyllum halli, Favosites* sp. aff. *Favosites placenta, Pleurodictyum dividua, Syringopora maclurei.*

The upper occurrence of Facies 2 (from 6.2 to 9.7 meters) is composed of gray fine siltstone that contains increasing shell concentrations upward from 6.2 meters. The upper occurrence of Facies 2 differs from the lower in that it contains a relatively greater coral fauna. The transition from Facies 3 to 2 is marked by an increase in the occurrence of shell concentrations. The fauna of this unit is composed of *Stromatoporella* sp., *Stereolasma rectum, Heliophyllum halli, Cystiphylloides americanum, Favosites clausus, Favosites nitella, Favosites* sp. aff. *Favosites hamiltoniae, Coenites* sp., *Syringopora maclurei, Mucrospirifer mucronatus, Mediospirifer audac*ulus, indeterminate brachiopods, *Phacops rana, Fenestella* sp., *Anastomopora* sp., *Sulcoretepora* sp., *Thamnotrypa* sp., and *Taeniopora* sp.

From 6.2 to 6.5 meters *Cystiphylloides americanum* and *Heliophyllum halli* are dominant. There are several shell concentrations that contain coral, brachiopod, and crinoid fragments. The concentration at 6.2 meters contains one visible stromatoporoid (approximately 2 cm high.) At approximately 6.35 meters a shell concentration of transported or biogenically mixed material is present. It includes *Cystiphylloides americanum* (one of which is encrusted by undetermined bryozoan species), *Favosites clausus, Pleurodictyum dividua, Emmonsia* sp., *Coenites* sp., *Phacops rana*, fenestrate bryozoans, along with brachiopod and crinoid fragments.

From 6.5 to 6.7 meters there is a sparse coral fauna that is composed of *Cystiphylloides americanum* and *Coenites* sp. From 6.7 to 8.7 meters the number of corals decreases and shell concentrations increase. The shell concentrations are mainly composed of brachiopod and crinoid fragments with a few of these beds containing small corals—*Favosites* sp. aff. *Favosites nitella*. Larger specimens of aff. *Favosites placenta* (4-5 cm or larger radius as compared to 2-3 cm radius in the previously mentioned) are present at approximately 6.9 and 8.0 meters.

From 8.7 to 9.1 meters *Coenites* sp. and *Cystiphylloides americanum* are extremely abundant and completely dominant the enclosed fauna. From 9.3 to 9.7 meters there are numerous small centimeter-scale shell concentrations that are composed of brachiopod and crinoid fragments.

**Facies 3**

Facies 3 (5.0 to 6.2 meters) is composed of gray fine siltstone. Within this fossiliferous unit two different subFacies can be recognized. The first, from 5.0 to 5.6 meters, is a dense bed of corals and the second, from 5.6 to 6.2 meters, is less dense and contains beds of shell debris.

Figure 47. Coral-rich zone (mostly *Cystiphylloides americanum*) of Facies 3, approximately 5 meters above base of Brodhead Creek outcrop (see Figure 46). (Photo by D. Monteverde.)

The first of these intervals is composed of massive gray siltstone that contains the densest concentration of fossil material, as well as the richest and most diverse fauna of all the intervals at this outcrop (Figure 47). It contains the following fauna: *Stromatoporella* sp., *Stereolasma rectum, Breviphrentis* sp., *Breviphrentis pumilla, Heterophrentis* sp., *Heliophyllum halli*, an indeterminate branching form of *H. halli, Eridophyllum archiaci, Cystiphylloides americanum, Favosites* sp. aff. *Favosites placenta, Favosites* *clausus, Favosites hamiltoniae, Favosites placenta, Favosites radiatus, Emmonsia arbuscula, Thamnopora* sp., *Trachypora elegantula, Alveolites* sp., *Coenites* sp., *Pleurodictyum dividua, Syringopora maclurei, Mucrospirifer mucronatus, Mediospirifer* sp., *Phacops rana, Fenestella* sp., and *Anastomopora* sp. The dominant species is *Cystiphylloides americanum*. A greater number of coral species was identified than brachiopod species. More brachiopod species may be present than identified in this study because identification of brachiopod species was hampered by the massive nature of the rock and poor preservation of brachiopod valves.

The majority of corals in this interval appear to be in situ. Those that are not found in situ are seen to occur in discrete layers accompanied by disarticulated shell debris. There is evidence that coral growth occurred while steady sedimentation was ongoing. For example, an approximately 20-cm *Cystiphylloides americanum* was recovered from a vertical position within the surrounding rock (5.0 to 5.1 meter interval.) The lowermost portion of the coral is bent indicating that the coral was previously disturbed from its original life position.

At 5.4 meters there are several thin layers of stromatoporoid that are separated by approximately one cm of sediment and shell material. They extend laterally approximately 50 cm. Some of these layers appear to be encrusted by an unknown species of bryozoan. Above these thin layers two massive stromatoporoids (*Stromatoporella* sp.) are present at 5.5 meters. One is disturbed from its original life position and found in a bed with other transported fauna. Slightly above this another stromatoporoid is present in life position.

The second interval, from 5.6 to 6.2 m, also is composed of fine gray siltstone. It differs from the underlying subFacies in the fauna it contains and in the distribution of that fauna. The number of corals slowly decreases upwards from 5.6 meters. The fauna of this subFacies includes: *Stromatoporella* sp., *Stereolasma rectum, Breviphrentis* sp., *Heterophrentis* sp., *Heliophyllum halli, Eridophyllum archiaci, Cystiphylloides americanum, Favosites clausus, Favosites* sp. aff. *Favosites placenta*, *Emmonsia arbuscula*, *Thamnopora* sp., *Coenites* sp., *Pleurodictyum dividua*, *Syringopora maclurei*, indeterminate brachiopods, *Phacops rana*, *Fenestella* sp., and *Anastomopora* sp. *Cystiphylloides americanum*, *Heliophyllum halli*, and *Coenites* sp. are the most dominant taxa of this fauna. The corals are found both in situ and disturbed from life position.

The corals disturbed from their original life position are usually found associated with other fauna such as brachiopods and crinoids in small concentrations of shell material. There are examples of corals being encrusted by other corals and bryozoans (5.8 m *Cystiphylloides* sp. with attached *Eridophyllum archaici*, 5.9 m *Heliophyllum halli* encrusted by undetermined bryozoan species, 6.1 m *Heliophyllum halli* encrusted by *Syringopora* sp., and 6.3 m *Cystiphylloides* sp. encrusted by *Favosites* sp.) The trilobites are disarticulated, indicating that they were probably molts. Fenestrate bryozoans are mainly preserved as whole specimens.

## BRODHEAD CREEK FAUNAL ASSOCIATIONS

Faunal associations were interpreted with the aid of both field observations and the results of cluster analysis. Cluster analysis assesses similarity levels between individual elements, such as fauna within a sampling interval, and then groups these elements based on similarity levels. Flexible and UPGMA cluster analysis was performed using the computer program NTSYS-PC (Numerical Taxonomy and Multivariate Analysis System) version 1.8 written by Rohlf (1994.) Generally the field observations of faunal and lithological zones within the Facies were confirmed by the results of cluster analysis.

**Association A: Sparse Containing Bivalves and Brachiopods**

**Faunal Composition**

This association has no dominant species occurring throughout any of the related intervals. Rather, single bedding planes tend to be dominated by a species or a low diversity of species. In the lower unit (below 3.1 meters and in the 3.3 and 3.4 meter intervals) the following species are found: indeterminate rugose corals, indeterminate brachiopods, *Lingula* sp., *Tropidoleptus carinatus, Mucrospirifer mucronatus, Elita* sp., *Megastrophia* sp., *Nucloidea* sp., *Paleoneilo* sp., *Modiomorpha* sp., *Platystoma* sp., *Pterinopectin* sp., *Fenestrella* sp., and crinoid debris. The upper unit (above 9.7 meters) contains a similar fauna composed of *Coenites* sp. (below 10 meters), *Rhipidomella* sp., *T. carinatus, M. mucronatus, Mediospirifer audaculus, Fimibrispirifer* sp., *Elita* sp., *Atrypa* sp., *Cypricardella* sp., *Nucloidea* sp., *Phacops rana*, *Greenops boothi*, *Fenestella* sp., *Anastomopora* sp., and crinoid debris.

**Lithology and Stratigraphic Position**

This association is found below 3.4 meters and above 9.7 meters. It is composed of brittle fine shaly siltstones that weather to a reddish color. There are also small oblong to round iron concretions (approximately 4-5 cm) present, especially in the lower unit. It corresponds to Facies 1 and flexible cluster analysis Biofacies A.

**Inferred Paleoecology**

The overall uniform nature of the nonfossiliferous shaly siltstone suggests a slow steady rate of sedimentation. The lack of shell orientation would indicate that was little or no current influence. The accumulation of a majority of fossils on discrete bedding planes indicates that these concentrations were a result of repeated single events.

Oxygen levels within the sediment appear to have fluctuated between anoxic and aerobic. The lack of bioturbation and trace fossils within the sediment indicates that oxygen levels were low. The occurrences of *Cypricardella* sp., a fully buried filter-feeder, *Paleoneilo* sp., a shallow infaunal deposit feeder, and *Modiomorpha* sp, an infaunal endobyssate bivalve, all indicate that the oxygen interface fluctuated within the sediment at times. Association A is interpreted as being generally below storm wave base, but perhaps affected by occasional storms. The interval above 9.7 meters contains a relatively higher diversity of species that the lower unit of Association A. This may indicate that conditions such as oxygen levels were more favorable to habitation.

**Association B: *Heliophyllum halli* – Spiriferid Brachiopods**

**Faunal Composition**

This association is marked by the occurrence of *Heliophyllum halli* and spiriferid brachiopods such as *Mucrospirifer mucronatus, Mediospirifer audaculus*, and *Spinocyrtia granulosa*. Other corals present include *Stereolasma rectum, Heterophrentis* sp., *Cystiphylloides americanum, Favosites* sp. aff. *F. placenta*, *Favosites nitella*, *Pleurodictyum dividua, Syringopora maclurei*, and *Coenites* sp. Additional less common brachiopod species include *Pentamerella* sp., *Rhipidomella* sp., *Leiorhynchus* sp., *Megastrophia* sp., *Elita* sp., *Athyris* sp., *Atrypa* sp., *Protoleptostrophia* sp., and indeterminate brachiopods. The association also includes rare *Phacops rana* and *Fenestella* sp. Disarticulated crinoid columnals are found throughout, sometimes comprising a majority of the surrounding matrix. Corals and brachiopods are commonly associated with concentrations of shell debris.

**Lithology and Stratigraphic Position**

This association is found from 3.6 to 5.0 meters, from 6.8 to 7.9 meters, and at 8.3 meters in the stratigraphic sequence at Brodhead Creek. This association is composed of a fine gray siltstone that contains numerous centimeter-scale layers of shell concentration. The association occurs in Facies 2 and corresponds to flexible cluster analysis Biofacies B-1.

**Inferred Paleoecology**

Fossils occur mostly in discrete layers within this association. This probably indicates that the association represents either brief times of in-situ colonization or layers of storm-transported shell debris. Since a majority of the fauna is not found in-situ, it is more likely that it has been transported and hence represents storm event horizons. The layers of shell concentrations are thicker and more numerous that those found in Association A. This implies that in comparison to Association A, this association occurred in a shallow environment below normal wave base, but within the reaches of storm wave base. Sedimentation rates were probably higher and constant except for sporadic storm disruptions.

**Association C: *Cystiphylloides americanum –Heliophyllum halli***

**Faunal Composition**

This, the most fossiliferous association, is dominated by the rugose corals *Cystiphylloides americanum* and *Heliophyllum halli*. It also contains rare stromatoporoids (*Stromatoporella* sp.) occurring both as massive forms and thin millimeter-scale layers. *Favosites hamilton*ae and large *Favosites* sp. occur from 5 to 5.1 meters. Other common corals include *Breviphrentis pumilla, Breviphrentis* sp., *Eridophyllum arachiaci, Favosites* sp.*, Favosites clausus, Emmonsia arbuscula, Coenites* sp., *Pleurodictyum dividua*, and *Syringopora maclurei*. Rare corals include an indeterminate branching form of *H. halli, Stereolasma rectum, Heterophrentis* sp., *Favosites placenta, Favosites radiatus, Trachypora elegantula,* and *Alveolites* sp. *Mucrospirifer mucronatus* and *Phacops rana* are present but rare. The bryozoans *Fenestella* sp., and *Anastomopora* sp. are common and occur throughout the association. Crinoid columnals are also common and occur throughout but are more common as part of the supporting matrix in layers of shell concentrations.

**Lithology and Stratigraphic Position**

Association C is found from 5.0 to 6.4 meters. The association occurs in fine gray siltstone. The interval from 5.0 to 5.6 meters is densely fossiliferous, but from 5.7 to 6.4 meters faunal density decreases. It corresponds to Facies 3 and flexible cluster analysis Biofacies C.

**Inferred Paleoecology**

Association C can be divided into two distinct intervals, from 5.0 to 5.5 meters and from 5.5 to 6.2 meters. The first interval represents an environment with shallow clear waters below normal wave base that is infrequently disturbed by storm wave base. Its composition of mainly turbidity intolerant corals and stromatoporoids suggests that rates of sedimentation were very low or sediment was nonexistent. The abundance of corals also suggest that waters were clear, well oxygenated, warm, and of normal marine salinity.

From 5.1 to 5.2 meters there is evidence for slow, steady sedimentation. Several species of *Cystiphylloides americanum* were found in-situ in an upward position. One specimen shows evidence of early disruption from normal orientation and continued growth upward from the new orientation.

Above this interval of slow, steady sedimentation from approximately 5.3 to 5.6 meters, there are two separate taxa that provide evidence of influxes of sediment. The first is the occurrence of an indeterminate branching form of *H. halli* in the 5.3-meter interval. The budding of new corals on this specimen suggests that the coral was under an environmental stress. A similar species, *H. delicatum*, has been recognized as being well adapted to muddy, carbonate-poor conditions (Oliver and Sorauf, 1994). Sediment smothered layers of stromatoporoids occur from 5.4 to 5.5 meters. Some of the stromatoporoids, which are encrusted by a bryozoan species, show evidence of post mortem exposure prior to burial. The sediment that buried these stromatoporoids contains small corals, shell fragments, and crinoid debris.

Above 5.6 meters, corals decrease in abundance becoming less common and layers of shell concentrations become common. Intervals of low sedimentation occur periodically within the interval from 5.6 to 6.2 meters. This is evident in the post mortem exposure of rugose coral species and their subsequent encrustation by an indeterminate byrozoan species. Overall, the lack of a robust in-situ coral fauna suggests that sedimentation rates were higher in this upper interval of Association C. Periods of steady sedimentation punctuated by storm disruption are apparent in the common occurrence of layers of shell concentrations. This would place the interval below normal wave base, but within the reaches of storm wave base.

### Association D: *Coenites – Cystiphylloides*

#### Faunal Composition

This association is composed almost exclusively of *Coenites* sp. and *Cystiphylloides* sp. with small *Favosites* sp. occurring in a few intervals. The *Cystiphylloides* sp. has smaller diameter (1-2 cm) than those occurring in Association C.

#### Lithology and Stratigraphic Position

This association occurs at 6.6 meters and from 8.0 to 9.1 meters. It is found within Facies 2 and corresponds to flexible cluster analysis Biofacies B-2.

#### Inferred Paleoecology

The dominance of the sediment tolerant ramose form of *Coenites* sp. and the smaller diameter forms of *Cystiphylloides* sp. indicates that the sedimentation rates were higher than in previous associations. The occurrence of large, round *Favosites* sp. and layers of shell concentrations indicate that this association was within the reaches of storm wave base.

## SUMMARY

The fauna of the Centerfield fossil zone is similar to that of the Centerfield Member of New York. In northeastern Pennsylvania, a faunal cyclicity has been documented at the Brodhead Creek outcrop. This cyclicity is represented in a symmetrical pattern with transitions from Facies 1 to Facies 2 to Facies 3 to Facies 2 to Facies 1. The cyclicity represents a transgressive sequence in which rocks that are very lithologically similar (fine to medium siltstone) change faunally from a brachiopod-dominated assemblage to a series of coral-dominated assemblages. Similar cyclicity is found in the Centerfield Member of New York (Savarese, 1984) indicating that sea level fluctuations at this time were occurring throughout the Appalachian Basin.

# LATE WISCONSINAN END MORAINES IN NORTHWESTERN NEW JERSEY: OBSERVATIONS ON THEIR DISTRIBUTION, MORPHOLOGY AND COMPOSITION

by

Ron W. Witte

***ABSTRACT***

End moraines in northwestern New Jersey are prominent, segmented, arcuate belts of hummocky till that cross Kittatinny and Minisink Valleys, Kittatinny Mountain, and the New Jersey Highlands. 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. 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. 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, morainal 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.

## INTRODUCTION

*If, instead of remaining stationary, the margin of the ice moved alternately backward and forward within narrow limits, the effect would have been to spread the moraine by widening the zone of submarginal accumulation. If during the oscillation of the margin it remained stationary either during or after its minor recessions or advances, or both, subordinate ridges would be developed, marking the position of several halts. If the edge of the ice remained parallel to itself as it advanced and receded, these subordinate ridges would be parallel, and each a miniature terminal moraine."*

Rollin D. Salisbury, *in* Salisbury, 1902, p. 96
(On the origin of moraine-parallel ridges.)

Even the casual student of New Jersey's geology knows about the Terminal Moraine, a low, uneven ridge of boulders and soil that sweeps across the northern part of the state from Perth Amboy to Belvidere. The moraine's course divides the state into two contrasting landscapes. North of the moraine there are many fresh to lightly weathered rock outcrops, thick, stony soils, valleys filled with thick deposits of stratified sand and gravel, silt, and clay, and numerous wetlands and lakes. South of the moraine rock outcrops are sparse and weathered, soils are typically more clayey, and wetlands are sparse. Because the terminal moraine was a readily distinguishable feature of New Jersey's landscape, it was well studied around the turn of the 20$^{th}$ century. R. D. Salisbury, in his magnum opus, *The Glacial Geology of New Jersey*, devoted thirty-eight pages to its origin, composition, and topography, as well as several additional pages on recessional moraines. The moraine was tangible evidence that continental glaciation was a very real geologic event and that it had left an indelible imprint on the landscape. Only 50 years earlier diluvialist views were accepted as fact in the scientific community. As a sign of the changing times, the Terminal Moraine and the Ogdensburg-Culvers Gap moraine found their way on New Jersey's first bedrock map (Lewis and Kummel, 1912). This surely caused consternation

Witte, R.W., 2001, Late Wisconsinan end moraines in northwestern New Jersey: observations on their distribution, morphology and composition, *in*, Inners, J. D. and Fleeger, G. M., eds., 2001—a Delaware River odyssey, Guidebook, 66$^{th}$ Annual Field Conference of Pennsylvania Geologists, Shawnee-on-Delaware, PA, p. 81 - 98.

among the day's geologic elite, whom viewed the study of surficial deposits as a lowly endeavor and not the proper field of study for serious scientists.

## GEOLOGIC SETTING

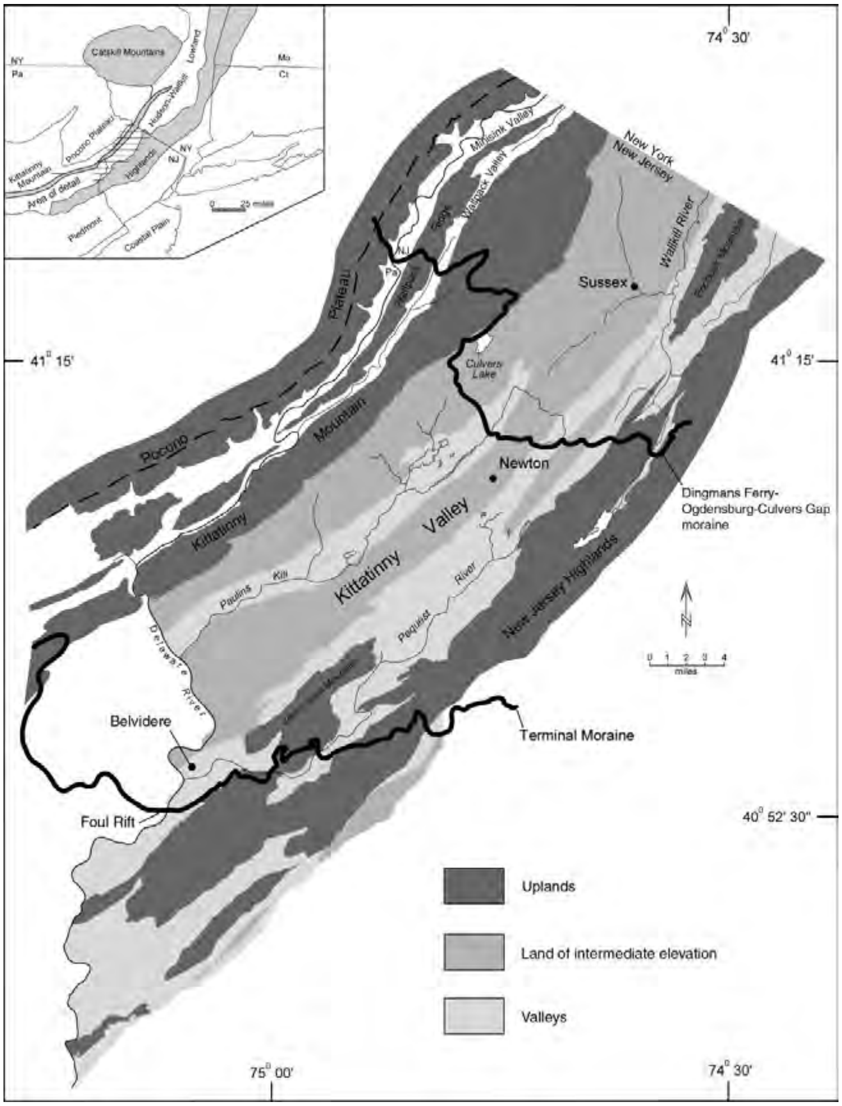



Figure 48. Physiography of northwestern New Jersey and location of geographic features named in text. Dark-gray areas represent major uplands, medium-gray areas represent lands of intermediate elevation, and light-gray areas represent valleys. Kittatinny Valley forms the broad lowland between Kittatinny Mountain and New Jersey Highlands. On the Figure, it includes areas shown as valleys, which are chiefly river valleys, and lands of intermediate elevation.

Kittatinny Valley is a broad lowland in northwestern New Jersey (Figure 48) that lies in a glaciated part of the Great Valley section of the Appalachian Valley and Ridge province. It is underlain by folded and thrust-faulted belts of dolomite, limestone, slate, and sandstone of Lower Paleozoic age (Figure 49). The valley is further cut by the Pequest River and Paulins Kill, which flow southwest toward the Delaware River, and the Wallkill River, which drains most of the upper part of Kittatinny Valley and flows northeast toward the Hudson River in New York. These rivers chiefly flow along strike-controlled belts of carbonate rock that are mostly dolomite of the Kittatinny Supergroup (Drake et al., 1996). Relief rarely exceeds 300 feet (90 m), rock outcrops are very abundant, and knobby topography is commonplace. Where the underlying rock contains more chert, narrow, strike-parallel ridges have formed. In other places fluvial erosion, dissolution, and glacial erosion have greatly lowered the valley floor by as much as 200 feet (60 m). Most of these areas are underlain by thick deposits of glaciofluvial and glaciolacustrine sediments, laid down during the late Wisconsinan glaciation. Slate, siltstone, and sandstone of the Martinsburg Formation (Drake et al., 1996) underlie interfluves in Kittatinny Valley and the area between Paulins Kill valley and Kittatinny Mountain. Overall, the average elevation here is about 300 feet (90 m) higher than the carbonate-floored valleys, and relief may be as much as 500 feet (150 m). Topography consists of rolling hills of moderate to steep slopes, and many strike-parallel ridges streamlined by glacial erosion. In some places bedrock is buried beneath drumlins and thick ground moraine.

The New Jersey Highlands, part of the southern extension of the New England physiographic province, borders the valley on its southeast side. Included with the Highlands is a large outlier in the southern part of the valley that includes Jenny Jump Mountain, Danville Mountain, High Rock Mountain, and Mount Mohepinoke. These uplands have rugged relief; their rough lands underlain by metasedimentary and intrusive rocks of Proterozoic age (Figure 49) that rise as much as 1000 feet (300 m) above the floor of Kittatinny Valley. Ridge lines chiefly follow layering in the country rock. However, discordant trends are common, and in places deep gaps cut across the southwest-trending topographic grain. Glacially scoured outcrops are common.

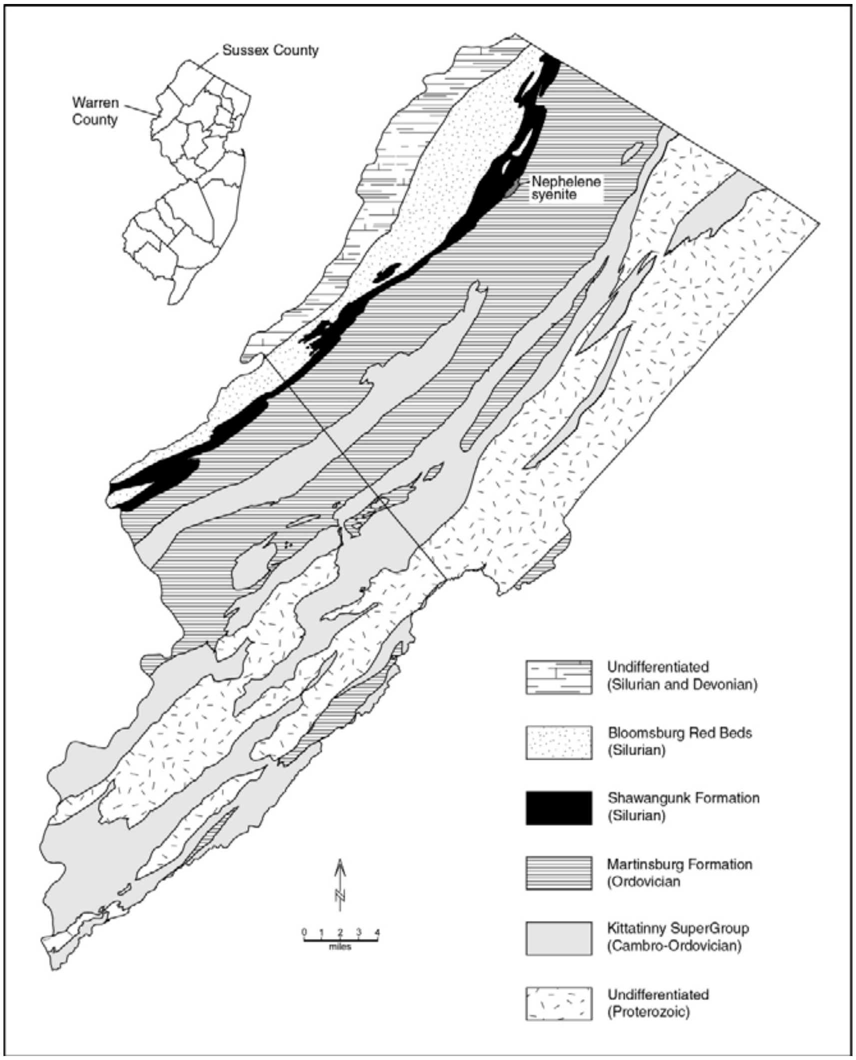



Figure 49. Simplified geologic map of Sussex and Warren Counties, northwestern New Jersey. Data modified from Drake et al. (1996). Lithologic Key: Proterozoic formations - chiefly metasedimentary and intrusive rocks with minor marble; Kittatinny Supergroup - dolomite and limestone; Martinsburg Formation - slate, shale, siltstone and graywacke; Shawangunk Formation - quartzite and quartz-pebble conglomerate; Bloomsburg Red Beds - shale and sandstone; Undifferentiated Silurian and Devonian Formations - shale, siltstone, sandstone, and limestone. Due to limited outcrop area the Jacksonburg Limestone is included with the Kittatinny Supergroup, and the Hardyston Quartzite is included with the Kittatinny Supergroup.

Kittatinny Mountain bounds Kittatinny Valley on its northwest side. The mountain is held up by the Shawangunk Formation, a tough, resistant quartzite, and quartz-pebble conglomerate, and the Bloomsburg Red Beds, which consists of interlayered red sandstone and red shale (Figure 49). Its nearly level summit forms a continuous ridge that extends from the Shawangunk Mountains in New York southwestward through New Jersey into Pennsylvania. Its main crest rises as much as 1000 feet (300 m) above the valley floor. In places its continuity is broken by large gaps, such as Culvers Gap, and Delaware Water Gap, and several smaller gaps and sags. Topography is rugged and consists of narrow- to broad-crested ridges that trend southwestward paralleling the main trend of the mountain. The mountain's steep southeast face also forms a nearly continuous escarpment in New Jersey. Rock outcrops are very abundant, and many have been shaped by glacial erosion. The piedmont that lies to the northwest of the mountain's main ridge here is also included as part of Kittatinny Mountain. Bedrock exposures are sparse there because the rock surface is covered by thick ground moraine and drumlins. In the Culvers Gap area, a series of repeating low amplitude folds and an overall decrease in the northwest dip of the outcrop belt nearly triples the mountain's width.

Wallpack Valley, Minisink Valley, and Wallpack Ridge lie northwest of Kittatinny Mountain (Figure 48). The valleys are narrow, deep, and trend southwest following belts of weaker rock, chiefly limestone, limy shale, and of Silurian and Devonian age (Figure 49). The valley floors are covered by thick deposits of glacial outwash and postglacial alluvium. The name *Minisink Valley* does not appear on U.S. Geological Survey topographical maps, but it is defined in Heilprin and Heilprin (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 translation of the word *Minisink* may be "the land from which water is gone" (Happ, 1938) or it may mean "stony country" (cited in Grumet, 1991, p. 176, as a personal communication from James Rementer, 1989).

## PREVIOUS INVESTIGATIONS

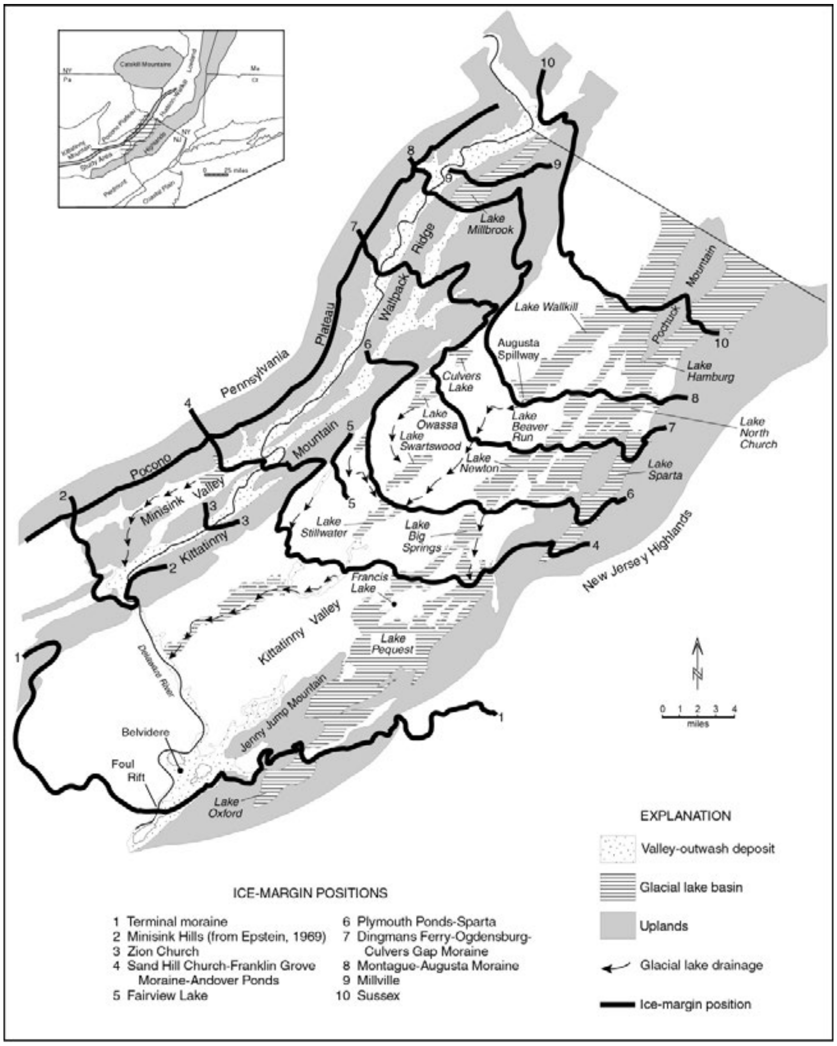



Figure 50. Late Wisconsinan ice-marginal positions of the Kittatinny and Minisink Valley ice lobes, and location of large glacial lakes, extensive valley-outwash deposits in northwestern New Jersey, and northeastern Pennsylvania. Modified from data by Crowl (1971), Epstein (1969), Minard (1961), Ridge (1983), and Witte (1988 and 1997a).

The surficial geology of northwestern New Jersey was first discussed by Cook (1877, 1878, and 1880) in a series of Annual Reports to the State Geologist. He included detailed observations on the age, distribution, and kinds of glacial drift, and evidence for glacial lakes. Lewis (1884) traced a terminal moraine westward from Delaware Valley to Salamanca, New York and considered it the same age along its length. A voluminous report by Salisbury (1902) detailed the glacial geology of New Jersey region by region. The Terminal Moraine (Figure 50) and all glacial drift north of it were interpreted to be of Wisconsinan age deposited during a single glaciation. Kames, kame terraces, deltas, recessional moraines, and glacial lakes were also described in Kittatinny Valley, where Salisbury also noted "in the northwestern part of the state, several halting places of ice can be distinguished by the study of successive aggradation plains in the valleys." Although it was recognized that some of these deposits marked ice-retreat positions, they were not documented within a larger chronostratigraphic framework. In Pennsylvania, Leverett (1934) slightly modified the trace of the Terminal Moraine and also assigned a Wisconsinan age to it and the glacial drift north of it. Cotter et al. (1986) also showed that the Terminal Moraine and the youngest glacial deposits in Pennsylvania and New Jersey are late Wisconsinan age, and are correlative with the Olean drift in Pennsylvania. Braun (1989), Witte and Stanford (1995), Stone et al. (in press) demonstrated that the youngest glacial deposits in eastern Pennsylvania and New Jersey are late Wisconsinan age,

Recessional moraines in Kittatinny Valley were originally identified by Salisbury (1902), and later remapped by Herpers (1961), Ridge (1983), and Witte (1988). Both the Ogdensburg-Culvers Gap and Augusta moraines were traced over Kittatinny Mountain by Herpers (1961), Minard (1961), and Witte (1997a). In Kittatinny Valley, Connally and Sirkin (1973) suggested the "Culvers Gap" moraine, not the Terminal Moraine, marked the southern limit of the late Wisconsinan ice sheet. Later, Connally et al. (1989) suggested that the limit was several kilometers south of the "Culvers Gap" moraine, based on the location of a "dead-ice sink" in the head of Paulins Kill valley. They further added that the moraines at Ogdensburg, Augusta, and Sussex are inversion ridges (meltwater deposits laid down in large cross-valley crevasses in stagnant ice), and therefore, they do not delineate ice-retreat positions. They also proposed that deglaciation occurred primarily by large-scale valley-ice lobe stagnation. Their interpretation was based on their recognition of extensive esker systems, massive crevasse-fill deposits, inversion ridges, and dead-ice sinks in the upper part of Kittatinny Valley. Crowl and Sevon (1980) and Cotter et al. (1986) demonstrated that glacial drift north of and including the Terminal Moraine is all of late Wisconsinan age and see no evidence to support a late Wisconsinan maximum position at or near the Ogdensburg-Culvers Gap moraine. A recent investigation by Larsen and Bierman (1995), whom used cosmogenic $^{26}Al$ dating of gneiss and quartzite erratics, also indicated the Terminal Moraine is of late Wisconsinan age.

Witte (1988) and Ridge (1983) accepted the late Wisconsinan age for the Terminal Moraine, and demonstrated that deglaciation was systematic in a northeast direction, and chiefly by stagnation-zone retreat. Ridge (1983) showed that Terminal Moraine was composed of several segments constructed at several ice-margin positions. Witte (1991, 1997a) further added that the "inversion ridges" of Connally et al. (1989) at Ogdensburg and Augusta are end moraines, and part of a much larger end-moraine complex that marks major ice-retreat positions of the Kittatinny and Minisink Valley ice lobes.

The deglacial history of the Laurentide ice sheet is well documented for northwestern New Jersey and parts of eastern Pennsylvania. Epstein (1969), Ridge (1983), Cotter et al. (1986), Stone et al. (1989), and Witte (1997a) showed that the margin of the Kittatinny Valley and Minisink Valley lobes retreated in a systematic manner with minimal stagnation. However, the age of the Terminal Moraine, timing of the late Wisconsinan maximum, and precise chronology of deglaciation are very uncertain. This is due to scant radiocarbon dates because of a lack of organic material that can be used to date deglaciation, inadequacies of dating bog-bottom organic material and concretions, and use of sedimentation rates to extrapolate bog-bottom radiocarbon dates. Also, there are few exposures of varves that can be used for chronology.

The few radiocarbon dates available bracket the age of the Terminal Moraine and retreat of ice from New Jersey. 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 near the Terminal Moraine by D. H. Cadwell (written communication, 1996) and basal organic material cored from Francis Lake by Cotter (1983) indicates a minimum age of deglaciation at 19,340 $\pm$ 695 yr B.P. (GX-4279), and 18,570 $\pm$ 250 yr B.P. (SI-5273) respectfully for the lower part of Kittatinny Valley. Because Francis Lake lies about 3 miles (5 km) southeast of the Franklin Grove moraine, this age is also 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 an estimated age of 17,210 yr B.P. for the Wallkill moraine by Connally and Sirkin (1973) suggest ice had retreated from New Jersey by 17,500 yr B.P.

Five ice margins (Figure 50), the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex have been identified that mark major recessional positions of the Kittatinny Valley lobe. The Sparta, Culvers Gap, Augusta, and Sussex margins are traceable onto the New Jersey Highlands, and all margins, except Sparta, are traceable westward into Minisink Valley. All appear to represent halts in ice retreat, and some mark minor readvances of the Kittatinny Valley lobe. This pattern of ice retreat is different from the more rapid style of retreat postulated for the lower part of Kittatinny Valley. These differences, as well as the close spacing of the Culvers Gap and Augusta margins, their large traceable extent, and correlation with extensive end moraines and ice-contact deltas in western New Jersey may indicate that other factors, besides local topographic control, have influenced the retreat history of the Kittatinny Valley lobe. Although the $O^{18}$ record shows that the climate remained cold and stable during the few thousand years that it took for ice to retreat from New Jersey, there existed minor fluctuations that may have influenced deglaciation.

## END MORAINES

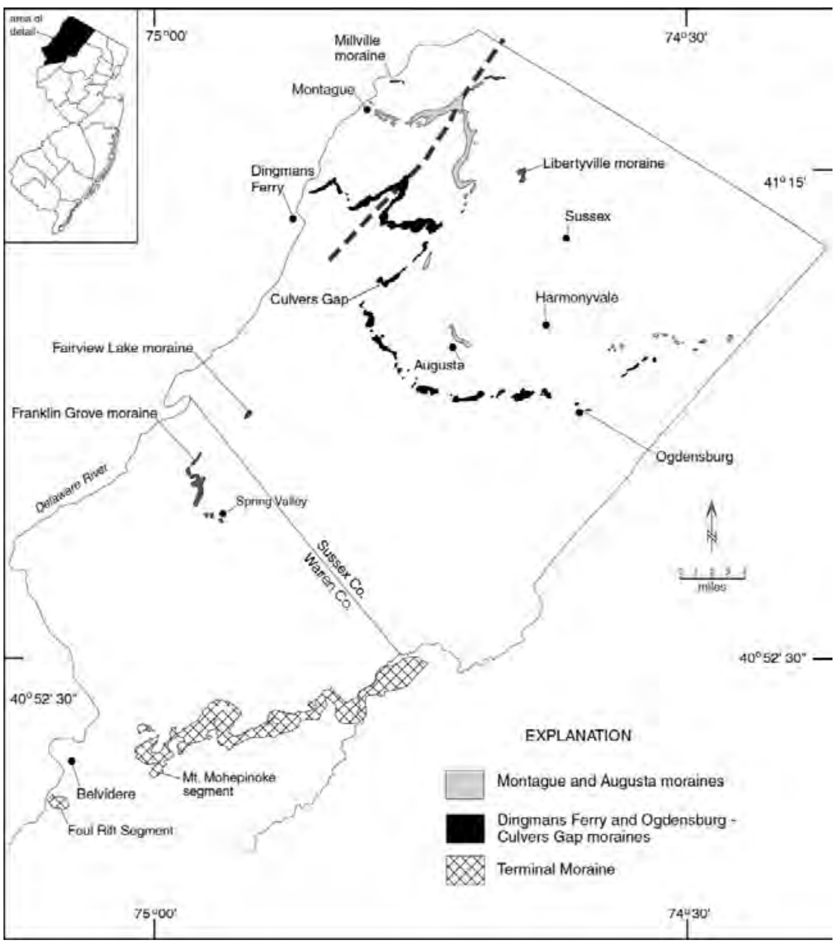



Figure 51. Surficial geologic map showing the late Wisconsinan Terminal Moraine, and recessional moraines in Sussex and Warren Counties, New Jersey. Dashed line marks the contact between the Dingmans Ferry and Ogdensburg-Culvers Gap moraines, and the Montague and Augusta moraines. Data from Ridge (1983), Stone et al. (1989), Witte (1988, 1991), Stanford et al. (1990), and Witte and Stanford (1995).

### Trace

In western New Jersey, end moraines are distinct, segmented belts of bouldery, hummocky glacial drift (Figure 51) that consist of poorly compact, sandy, bouldery till, minor lenses of glaciotectonized substrate (outwash, weathered bedrock, older till, and colluvium), and water-laid sand, gravel, and silt. The Terminal Moraine is as much as 130 feet (40 m) thick and one mile (0.6 km) in width. The recessional moraines are as much as 65 feet (20 m) thick and 2500 feet (760 m) wide. However, most are less than 1000 feet (300 m) wide. All the end moraines have asymmetrical topographic profiles with their distal slopes (outer edges) being the steepest. Their distal margins also have sharp boundaries, whereas the boundaries of their inner margins are typically indistinct. Morainal topography (sharpness, and overall relief) is also better formed along the moraine's outer margin than its inner margin. This was noted by Salisbury (1902, p. 251) who observed that "...the characteristic morainic topography made by the close association of hummocks, kettles, ridges, and troughs, is, in general, better marked in the outer half of the moraine belt than in the inner."

The Terminal Moraine follows a nearly continuous looping course through Warren County (Figure 51). In most places the morainal topography is distinct and easily recognized by its well-formed ridge-and-trough and knob-and-kettle topography. In a few places, where steep topography constrained its formation, morainal topography is only faintly noticeable. R.D. Salisbury and H.B. Kummel (Salisbury, 1902) described in detail the course of the moraine across New Jersey. Their excellent description of its trace through the southern part of Kittatinny Valley and across the Jenny Jump outlier stands today, except for a few areas in the Pequest and Delaware Valleys where outwash was mapped as part of the moraine. Clearly, the moraine's course was strongly influenced by topography, extending more southward in areas of lower elevation, and as it approaches the central axis of Kittatinny Valley. In many places a narrow belt of late Wisconsinan till extends as much as 3000 feet (900 m) out beyond the Terminal Moraine. This shows that the Terminal Moraine does not always represent the late Wisconsinan glacial border.

Recessional moraines in Kittatinny Valley include Franklin Grove, Fairview Lake, Ogdensburg-Culvers Gap, Augusta, and Libertyville (Figure 51). The smaller Fairview Lake and Libertyville moraines are correlated with heads-of-outwash situated farther east. The larger ones are more continuous in the valley, and both the Ogdensburg-Culvers Gap and Augusta moraines are traceable across Kittatinny Mountain where the former joins the Dingmans Ferry moraine and the latter joins the Montague moraine (Witte 1997a). The Franklin Grove moraine was first described by Salisbury (1902) and later named by Ridge (1983). The moraine trends northwestward from Spring Valley, through Franklin Grove, toward Sand Pond, and ends abruptly at the base of Kittatinny Mountain. This moraine does not continue across the mountain and it is absent east of Spring Valley. However, it is correlated with the Lake Pequest and Andover Ponds morphosequences situated farther east (Ridge 1983; Witte 1988, 1991) in Kittatinny Valley, and with the Sand Hill Church deposits in Pennsylvania (Witte 1997a; see STOP 9, Day 2). The Ogdensburg-Culvers Gap and Augusta moraines were first described by Salisbury (1902) and traced onto Kittatinny Mountain by Herpers (1961), and Minard (1961), and later remapped by Witte (1988, 1997a). The Ogdensburg-Culvers Gap moraine consists of several segments that trend westward from the New Jersey Highlands, through Ogdensburg to Culvers Gap in a distinct cross-valley loop. It continues along the southwest side of Kittatinny Mountain to where it crosses the main ridge crest, approximately 4 miles (6 km) northeast of Culvers Gap. From here its course traces a smaller loop through the Big Flat Brook valley and joins the Dingmans Ferry moraine. The Augusta moraine consists of several segments that trend westward from the New Jersey Highlands, through Harmonyvale to the base of Kittatinny Mountain, where it lies approximately 3 miles (5 km) northeast of Culvers Gap. From here it can be traced onto Kittatinny Mountain where it joins the Montague moraine, following a course similarly parallel to the Ogdensburg-Culvers Gap moraine. East of the Harmonyvale in Kittatinny Valley, morainal deposits are absent or they may lie buried beneath outwash. Ice retreat positions here are marked by ice-contact deltas in the Beaver Run and Wallkill River valleys, and they are correlated with the same ice margin as that marked by the moraine.

Morainal deposits in Minisink Valley include Dingmans Ferry, Montague, and Millville moraines (Figure 51). 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, this guidebook, p. 99). 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 et al. (1989), situated further west on the Pocono Plateau. Presumably the moraines in Minisink Valley have been eroded by glaciofluvial action. 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.

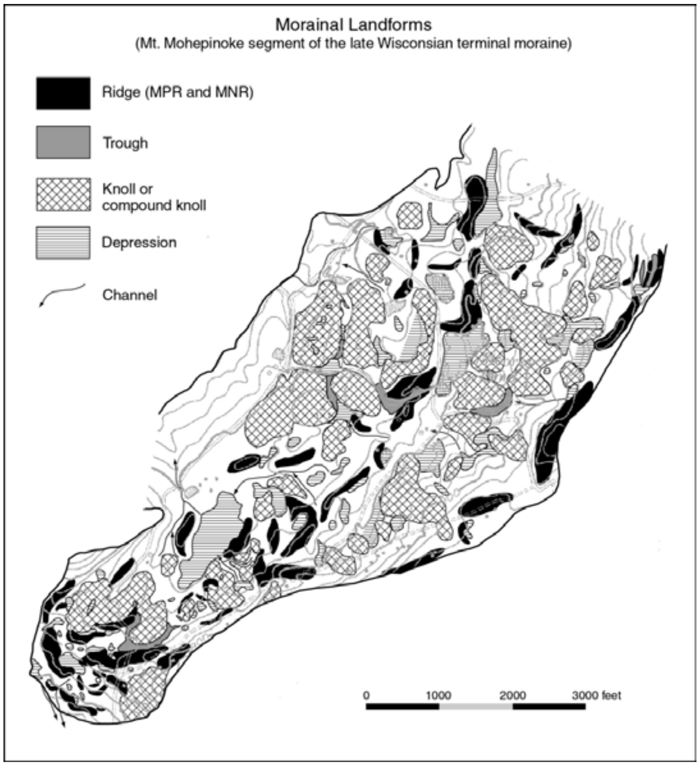



Figure 52. Morphology of the Mt. Mohepinoke segment of the Terminal Moraine, Warren County, New Jersey. Morainal landform elements collectively define areas of ridge-and-trough, and knob-and-kettle topography. Figure from Witte (2000).

**Morphology**

End moraines consist of a variety of topographic landforms that collectively form a belt as much as one mile wide of thick uneven bouldery drift. The complex assemblage of depressions, boulder fields, ridges, and mounds provides a diverse ecological setting for many kinds of plants and animals. To the casual observer the collection of ridges, mounds, and depressions appears random and chaotic in their trace across the countryside. Many end moraines have been simply described as a belt of hummocky drift. However, upon close inspection, they consist of several types of topographic elements that can be mapped, and characterized.

Morainal landforms may be grouped into positive and negative topographic elements (Figures 52 and 53). Positive elements include ridges, knolls, and plateaus. Ridges are further divided into moraine parallel ridges (MPR) and moraine nonparallel ridges (MNR). MPR’s generally lie along the outer margin of the moraine where they parallel its trace. They have narrow to broad crests, stand as much as 50 feet (15 m) high, and are as much as 2000 feet (600 m) long, although most are less than 500 feet (150 m) long. Many appear to have been formerly continuous, but may have been disconnected by collapse during melting of buried ice. Ridge crests follow straight to slightly arcuate traces that parallel the moraine’s outer border. In places they form nested sets that exhibit a remarkable degree of parallelism (Figure 53), suggesting they were built at several ice-margin positions. Their topographic profiles are typically asymmetric with their inner slopes the steepest. Inner slopes are also hummocky showing that this part of the ridge was laid down against ice. MPR’s typically occur along the outer part of the morainal belt. They are either push ridges, formed where the advancing ice had bulldozed ice-marginal sediment, or they are colluvial ramparts, laid down where the glacier margin remained stationary, shedding an apron of debris off its terminus. MNR’s are found throughout the morainal belt, and they are of similar dimensions as MPR’s, although they are not as numerous. Their ridge crests lie tangent to the moraine’s course, and they follow straight to sinuous traces. Side slopes are typically steep-sided and hummocky. In places the trace of their ridge crests define polygonal patterns. They may be crevasse fillings, formed where supraglacial debris had accumulated in deep fractures.

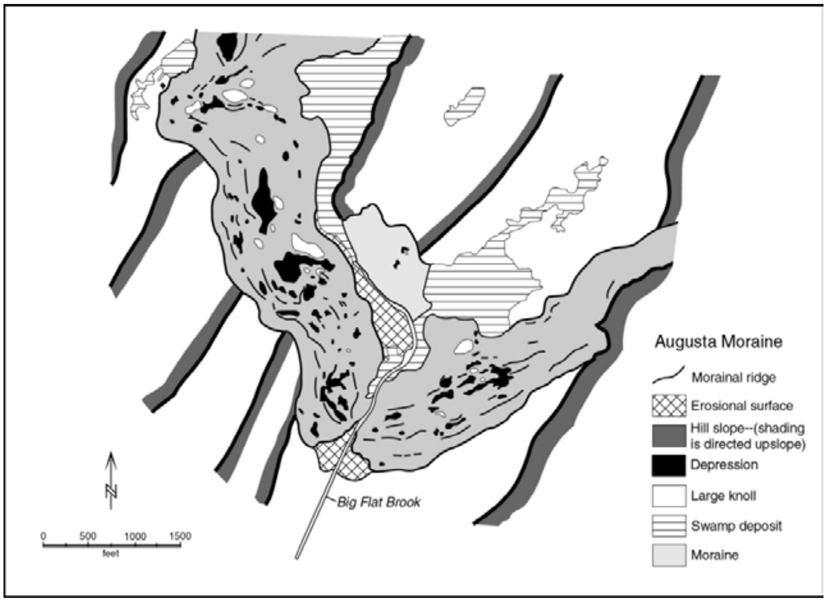



Figure 53. Morphology of the Augusta moraine where it crosses Big Flat Brook valley, Kittatinny Mountain, Sussex County, New Jersey. Morainal landform elements collectively define areas of ridge-and-trough, and knob-and-kettle topography. Figure from Witte (2000).

Knolls consist of low, rounded or elliptical hills that vary from larger isolated hills to compound forms that consist of several smaller hillocks. Relief is generally less than 25 feet (8 m), although in places it may be as much as 60 feet (18 m) and side slopes are variable. These features may be found throughout the morainal belt, but they typically are found along the moraine's inner margin. Collectively, they make up the largest areas in the moraine. These landforms probably represent places where supraglacial debris collected in hollows at the glacier's terminus. Over time, the icy substrate melted, letting down its sediment load on the land; the thicker areas of sediment now forming the higher parts of the moraine.

Plateaus form flat-topped, broad to slightly arched hills underlain by till. They are absent in the study area, but they have been found farther eastward (Stanford, 2000). Many parts of the moraine have a subdued morphology, marked by broad elevated areas of low hummocky relief (Figure 52). In places these higher areas have ice-contact scarps along their inner borders. Their origin is unclear. They may reflect places where readvancing ice has planed the moraine's surface, or the configuration of stagnant ice and supraglacial debris did not lend itself to forming distinct morainal topography.

Negative topographic elements include troughs, kettles, and meltwater channels. Troughs are elongated depressions that typically parallel MPR's. They are best formed in the outer part of the morainal belt, where in many places they separate nested sets of MPR's (Figure 53). They are as much as 40 feet (12 m) deep, 100 feet (30 m) wide, and 300 feet (90 m) long. These troughs represent places of little to moderate sediment accumulation between MPR's, or they may have originally been ice-cored ridges. Kettles are circular to irregularly shaped steep-sided depressions. In places they are only partially enclosed forming small amphitheater-shaped bowls. They are as much as 40 feet (12 m) deep, and as much as 500 feet (150 m) wide. Many depressions are wet and contain swamp or bog deposits. Other depressions are dry or contain seasonal water. Kettles have formed where detached blocks of residual ice have melted, leaving behind topographic depressions. In places, low-lying morainal areas are formed by several enclosed to partially enclosed depressions and bowls. They represent the opposite form of the compound morainal knoll and they formed where residual ice initially held up the higher areas along the glacier's margin.

In places the moraine is cut by small, narrow, straight to sinuous channels. These features may be as much as 40 feet (12 m) deep, and typically have bouldery floors. Today they are used by ephemeral streams where they carry off discharge from small springs. Some channels probably formed during the earlier phases of moraine formation when meltwater streams emanating from the active glacier margin flowed along the moraines outer border. Later, during stagnation, meltwater—chiefly from melting stagnant ice—drained from hollow to hollow and eventually formed a loosely organized drainage network.

Typically, the innermost parts of the morainal segments have fewer ridges, fewer elongated depressions, and are marked by knob-and-kettle rather than ridge-and-trough topography. The morphology expressed by the Augusta moraine (Figure 53) is typical for morainal segments that abut thick and widespread till. Overall these segments are larger, more continuous, and have more fully developed moraine-parallel ridges than those abutting thin patchy drift. This strongly suggests that unconsolidated material near the glacier's terminus may have supplied most of the sediment that makes up the moraine, rather than nearby glacially eroded bedrock.

**Composition**

End moraines consist of non-compact, bouldery, silty-sandy to sandy till with minor beds and lenses of water-laid sand, silt, and gravel (Figure 54). This material is distinctly different from the more compact, and less stony ground moraine or till that lies near the moraine. Additionally, stratified drift is <u>not</u> a major constituent, even in places where the moraine crosses river valleys or former glacial lake basins. The lithology of the moraine is decidedly local in origin. This was noted by Salisbury (1902, p. 254) who reported that "... the lithologic composition of the till varied from point to point, according to the nature of the formations over which the ice has passed." For example in the Delaware Valley where the Terminal Moraine rests on outwash, it contains many rounded, waterworn stones that mimic the provenance of the outwash. On Jenny Jump Mountain, crystalline materials make up the bulk of the moraine. Outcrops of morainal materials are rare due to the difficulty of digging the bouldery drift, but more importantly its lack of economic value. The best outcrops are places where the moraine has been removed to expose economic deposits of sand and gravel, such as Foul Rift. The few exposures observed by the author show that the moraine consists largely of till with minor interlayers and lenses of sorted sand, silt, and gravel. Upon close inspection most of the till is faintly layered with individual layers varying greatly in thickness from less than one foot to as much as 10 feet (3 m). Layering is typically subhorizontal, its base marked by a concentration of larger stones, and crude normal grading has been observed in some of these pseudo beds. The heterogeneity of morainal sediment, its indistinct layering and grading, and inclusion of water-laid sand, silt, and gravel beds and lenses suggest that most of this material has had a complex history of deposition and is chiefly a product of mass wasting.

Figure 54. Outcrop of till exposed in the Montague recessional moraine near Montague, New Jersey.

**Foul Rift Segment of the Terminal Moraine**

The Foul Rift pit lies in the Delaware Valley about two miles south of the village of Belvidere, New Jersey (Figure 50). The pit is named after a nearby section of class two rapids on the Delaware River formed over dolomite ledge and glacial boulders. The Foul Rift pit provides an exceptional opportunity to study—in three dimensions—glacial valley-fill laid down at the front of the Kittatinny Valley ice lobe, and to view a cross section of an end moraine. Figure 55 represents a measured section that summarizes the Foul Rift stratigraphy. Unfortunately, this section is now covered by a very large stockpile of cobbles and small boulders. Because this pit is active, most of the outcrops viewed earlier by the author are no longer available for inspection. However, the overall stratigraphy of the pit and character of materials exposed there have remained consistent.

The lower gravel and sand (Figures. 55 and 56,inset) is glacial outwash laid down in front of the advancing ice sheet. This unit makes up more than half the stratigraphic section at the Foul Rift pit. Based on the distribution of nearby bedrock outcrops and a reconstruction of the buried rock topography (Witte and Stanford, 1995) bedrock beneath the pit floor is about 200 feet (60 m) above sea level, rising upward to the east. Based on this estimate there may be an additional 50 to 75 feet (15 to 23 m) of material beneath the lowest part of the pit floor. These stratified materials consist chiefly of matrix-supported planar to cross-stratified cobble-pebble gravel, pebble gravel, pebbly sand, and minor lenses of sand. The provenance of the outwash has a decidedly Delaware Valley lithology. Clasts consist chiefly of dolostone, slate, graywacke, and quartzite, with secondary amounts of red sandstone. Gneiss and granite (Highlands source) account for less than 5 percent of the gravel fraction.

These materials were laid down by anastomosing sets of meltwater streams that formed a braided pattern of channels and bars across the valley floor. Most of the gravelly beds are bars, while most of the sandy lenses are channel-fill deposits. Individual beds may be as much as five feet thick, although most are less than two feet thick. Both normal and reverse grading may be observed and grain size changes rapidly in both vertical and horizontal directions. This shows that these materials were deposited under highly fluctuating water discharges, the result of daily and seasonal meltwater production. In a nearby pit, located about one mile (0.6 km) downstream in the Buckhorn Creek Valley, boulders, some as large as 3 feet (0.9 m) in diameter, form coarse beds. These features were probably deposited during a meltwater mega-flood related to an outburst of subglacially trapped meltwater. In places the stratified materials are cemented with calcium carbonate. The location of the cemented gravel probably represents the former water table where calcium carbonate was precipitated during cycles of wetting and drying. The cementing agent was probably derived from weathered clasts of carbonate rock.

Lying above the proglacial outwash is a compact, fissile, sandy-silty till that contains many striated and rounded clasts (as much as 15 percent by volume). Lithology of the clast fraction is similar to that of the underlying outwash. In places the till contains thin beds, small lenses, and clots of sand, pebbly sand, and pebbly gravel. These intra-unit materials exhibit horizontal to subhorizontal attitudes, typically have pinch and swell boundaries, and have the overall appearance of having been sheared. The lower till contact is typically abrupt, and there is very little mixing across boundaries, other than a few clasts that straddle the contact. Elongated clasts show a preferred long axis orientation downvalley (Figure 55). This material appears to be a basal till laid down at the base of the ice sheet when it advanced to its most southern position about 2500 feet downvalley from the pit. A large part of the till is reworked glacial outwash. The preservation of some primary bedding structures within these intra-unit beds suggest that some parts of the outwash were frozen before their incorporation within the glacier's sole.

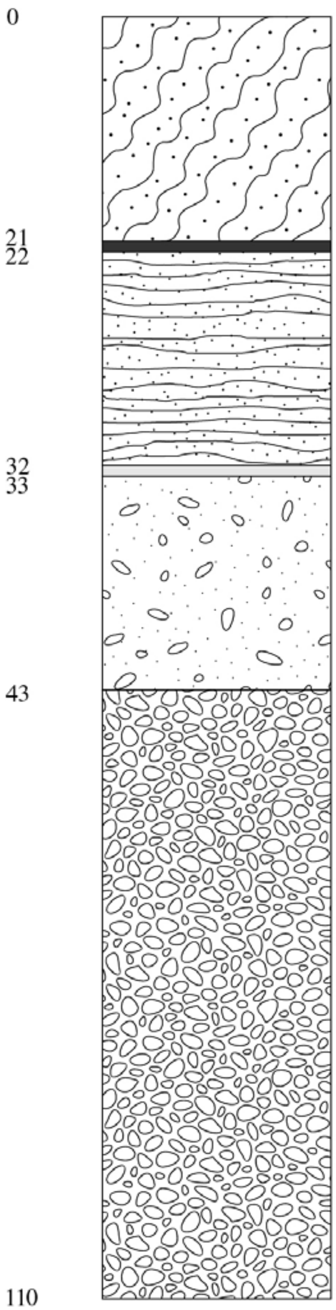
Based on the location of the lower till and the fact that it was deposited during glacial advance, it may be part of the same till sheet that covers the rock ridge south of the Foul Rift moraine. In places a thin layer of laminated silt and clay caps the till. This unit has been observed elsewhere in the pit. Previous exposures did show that this material extended several hundreds of feet southward. The silt and clay bed might be lacustrine, although its depositional setting is unclear. Its location atop the basal till suggests that the glacier had retreated north of the pit location at the time the fine-grained material was deposited. Possibly residual ice downvalley may have temporarily dammed a lake in front of the ice sheet. These materials may have also been deposited in shallow depressions formed on the surface of the till sheet by glacial scour.

## Section FR-4 - Foul Rift sand and gravel pit

Stratigraphic Column

Description of Materials

feet below
land surface




**Upper till** - Noncompact, stony, silty-sandy diamicton that contains by volume 15 to 25 percent, chiefly subrounded to rounded cobbles, pebbles and boulders. Contains intrabeds of sheared, stratified pebbly sand, gravel, and laminated silt and clay. Boulders in places form faint layers. Dolomite and sandstone clasts typically exhibit striations. These materials form the surface deposits of the Foul Rift moraine. Origin of materials is polygenetic ranging from an ablationary cover to ice thrusted substrate.

**Promorainal outwash** - Sheared beds of pebbly sand and sand. Deformation is much more pronounced in the upper part of this unit and was caused by overridding glacial ice.

**Promorainal outwash** - Mix of cobble-pebble, and pebble gravel, pebbly sand, and laminated silt and clay. Texture and primary sedimentary structures show that these materials were laid down by meltwater streams in both fluvial and lacustrine settings.

**Lacustrine sediment** - Laminated clayey silt with thin laminae of pink clay. Deposit largely derived from the settling of fines in a glacial. Clay - silt couplets may represent annual layers.

**Basal till** - Compact, sandy-silty diamicton that contains by volume 10 to 15 percent, chiefly subrounded to rounded pebbles, cobbles, and a few boulders. Contains intrabeds and lenses of sheared sand, pebbly sand, and pebble gravel. Dolostone and sandstone clasts exhibit striae and elongated clasts have a preferred orientation of northeast to southwest (see inset). Deposited at glacier's base. Largely derived from glacially eroded outwash.

**Proglacial outwash** - Planar to slightly curved, non-parallel tabular to lenticular beds of cobble-pebble, pebble gravel and pebbly sand, and minor lenticular beds of planar to cross-stratified sand. In places cemented by calcium carbonate. Gravel beds are typically less than 3 feet thick, generally rest on erosional contacts, have coarse clasts at their base, and exhibit normal grading. Imbrication of clasts is minor. Sand beds are generally less than 1.5 feet thick, and less than 10 feet wide. Their base conformably overlies coarser gravel and sand, where as the top of the beds have been truncated by erosion. Gravelly beds are channel bars and the sand beds are channel fills. These deposits were laid down by anastomosing sets of meltwater streams.

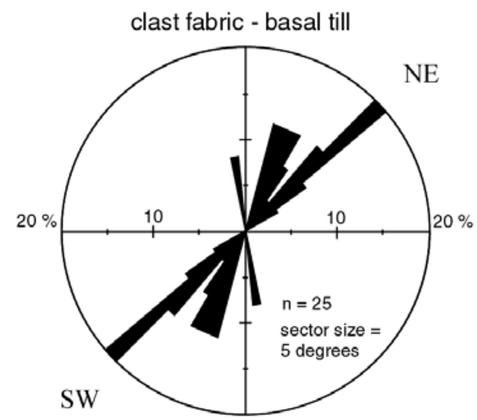



Figure 55. Stratigraphy of late Wisconsinan glacial drift measured in the Delaware Valley near Foul Rift, New Jersey. Inset Figure is a frequency rose diagram representing the azimuth of elongated pebbles (three to five inches in length) in the basal till. Figure from Witte (2000).

Lying above the till is a complex section of stratified gravel, sand, and silt. This unit has been the most difficult to decipher in terms of its history because its composition, bedding, and geometry vary throughout the pit. Previous exposures showed that materials ranged from cobble-pebble gravel to clayey silt. Bedding contacts typically exhibit pinch-and-swell traces across the outcrop face, and are sharply truncated in places. Texture and sedimentary structures show that these materials were laid down by meltwater streams in both fluvial and lacustrine settings. However, the complex geometric relationships between the various layers and lenses of material cannot be explained as a product of deposition, but as a product of postdepositional deformation. Given that these materials were laid down at the margin of an ice lobe and that active ice was present in the Delaware Valley, as shown by the lower till, deformation was probably caused by ice shove during a readvance of the Delaware Valley sublobe.

New exposures (Figure 56, located northeast of section FR-4) opened in 1999 showed large-scale recumbent folds and imbricate thrusts that further support the contention that this material has been ice shoved and probably overridden by ice. Near the upper center of the Figure and below a moraine parallel ridge, there are several stacked ramp-like structures that decrease in attitude going upwards. Beds of darker-colored material highlight these features. They consist of deformed clayey-silt (Figure 56, inset) similar to the fine-grained materials previously described. In other places the dark-colored beds are till. Most of the deformation is best preserved in the finer-grained materials rather than the coarser gravels. Apparently the gravelly material was largely deformed by intergranular rotation and sliding (similar to shoving a pile of marbles), whereas the finer material, because of its higher moisture content, and competence, deformed more ductilely.

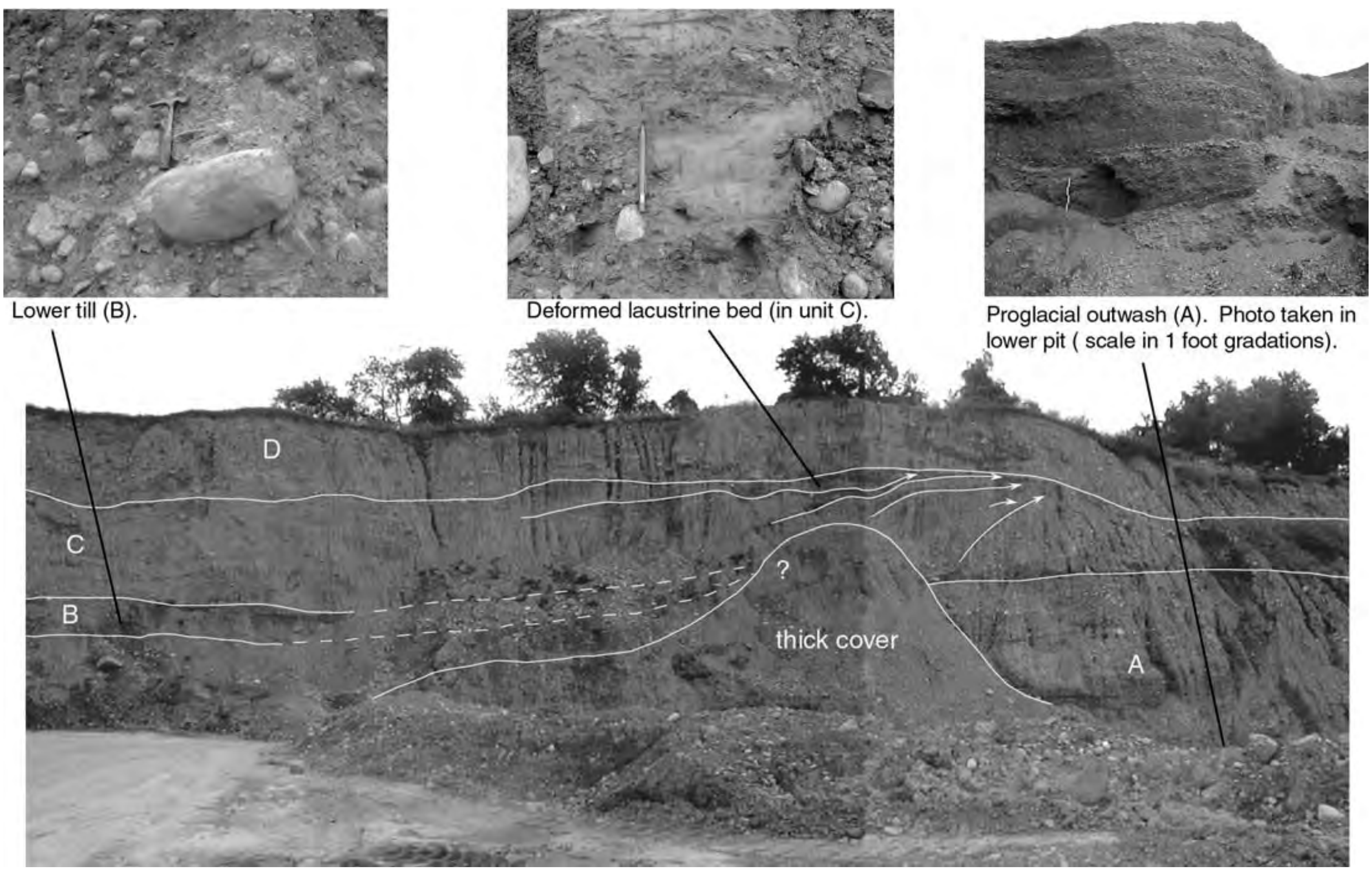



Figure 56. Composite section of the east wall, Foul Rift pit. Units: A - proglacial outwash, B - lower till (basal), C - deformed outwash, arrows denote thrusts, and D - upper till (Foul Rift moraine). Thrusts typically marked by deformed beds of silt and clay. Figure from Witte (2000).

The ramp-like structures may be a sequence of stacked thrusts that consist of ice-shoved outwash and glacial pond sediment.  The stacking may represent several advance and retreat cycles, or the thrusts may have developed during a solitary readvance.  Based on the amount of deformation observed in section FR-4 (Figure 56) it appears that deformation attenuates down valley with the highest degree of deformation occurring beneath a moraine-parallel ridge.

scale = 4 feet

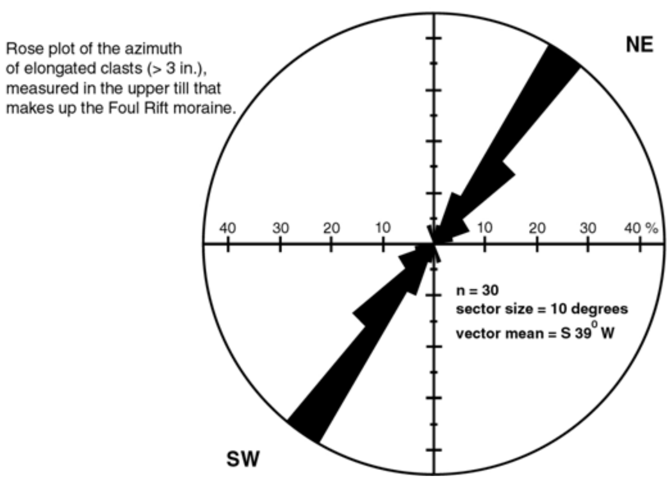



Figure 57.  Upper till exposed along the north rim of the Foul Rift pit. Compact, fissile, sandy-silty matrix containing by volume 7 to 10 percent subangular to subrounded stones.  Many stones are striated and elongated clasts have a pronounced downvalley fabric.  Although the till forms part of the Foul Rift moraine, it has characteristics of a basal till. It may represent a subglacial till facies associated with a push moraine.  Figure from Witte (2000).

The upper unit (Figures 55 and 56) consists of a poorly compact, stony, silty-sandy till containing lenses and layers of gravel and sand.   Some of these materials may be debris flows or flowtill, while others are glaciotectonized outwash.  Stoniness varies widely throughout the till, and dolostone and sandstone clasts are typically striated.  In places, boulders form weak layers.  This upper till unit makes up the Foul Rift moraine.  In most places this material looks to have been derived from an ablationary cover at the glacier's terminus.  New exposures cut into the northern pit wall suggest that the moraine may in part consist of basal till.  To effectively quarry the underlying sand and gravel, the upper till has been stripped by cutting a bench into the northern pit wall.  This process has unfortunately covered excellent exposures of basal till and deformed outwash, but it has revealed a character of the upper till never seen before.  The character of the morainal till exposed here (Figure 57) is much different than it is elsewhere in the pit.  It has a more massive structure, and lower stone content.  Its matrix is compact silty sand, and it contains a mix of rounded to subangular stones. Elongated clasts also have a moderately strong down valley fabric (Figure 57), and many of these exhibit imbrication down valley.  Based on its character this material appears to be a basal till, although here it forms the surface till that makes up the moraine.

Interpretations about the moraine's genesis have changed considerably over the years.  Most of the earlier workers (Lewis, 1884; Salisbury, 1902) suggested that the cross-valley ridge at Foul Rift was part of the Terminal Moraine.  Ward (1938) proposed that the moraine was a recessional kame complex formed behind the terminal moraine, which he placed farther downvalley.  Ridge (1983) suggested that the moraine was a frontal kame complex (coeval with the Terminal Moraine), largely consisting of stratified gravel and sand laid down among stagnant ice at the margin of the Delaware Valley sublobe. This interpretation was revised by Ridge (1985) upon seeing new exposures. He proposed that the Foul Rift moraine consisted of several push moraines, largely derived from the underlying outwash. The new interpretation emphasized the role of active ice at the glacier's margin.

The recent exposures along the east wall of the pit provide additional insights about the formation of an end moraine. The outcrop face shown in Figure 56 runs nearly perpendicular to the Foul Rift moraine, and it bisects a moraine parallel ridge. This view of the moraine and underlying materials provides strong evidence that active ice associated with one or several readvances formed large parts of the end moraine. Deformation seen beneath the moraine parallel ridge suggests that this feature is the result of ice shove. The till that makes up this part of the moraine is superincumbent on this ridge. It consists of ablation till, flow till, and blocks of overridden substrate. The basal till that lies along the north rim of the pit reflects a change in facies, going from a supraglacial setting to a subglacial one.

The submorainal ridge may have a deeper origin. The basal till measured in section FR-4 appears to have been a continuous sheet based on older exposures seen by the author. In Figure 55, the lower till crops out on the lower left side of photograph. From here it was seen to rise as much as 15 feet (4.5 m) and eventually terminate near a position well below the imbricate thrusts. It may lie beneath the slope cover, but if it does it has become much thinner. This geometry suggests that a gravel ridge existed before the deposition of the basal till. The origin of the ridge is unclear. Materials exposed there do not appear to be deformed, as they are higher in the pit. In part this may be due to the coarse texture of the material. This lower ridge may have formed out in front of the advancing ice lobe due to ice shove or it may have been part of a pressure ridge formed by the weight of the ice lobe. Is it just a coincidence that the moraine parallel ridge lies above this deeper ridge?

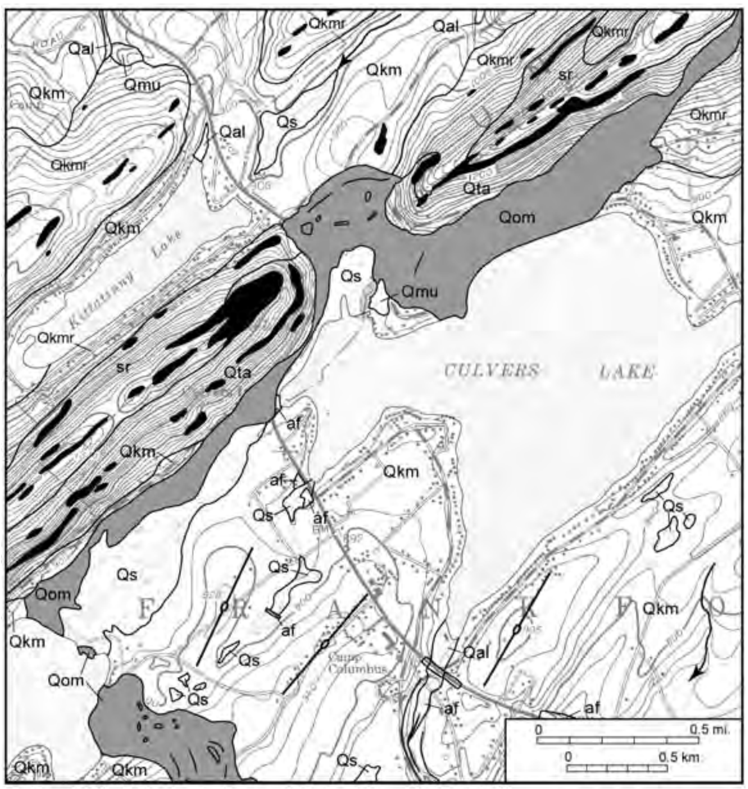



Figure 58. Surficial geology of a portion of the Culvers Gap quadrangle, New Jersey, in the vicinity of Culvers Gap showing local lobation of the Ogdensburg-Culvers Gap moraine. List of map units: af - artificial fill, Qs - swamp and bog deposits, Qal - alluvium, Qta - talus, Qkm - thick till, Qkmr - thin till, Qom (shaded gray area) - Ogdensburg-Culvers Gap moraine, Qmu - small undifferentiated meltwater deposits, Qft - meltwater-terrace deposits, sr - regolith, chiefly rock waste on steep hillslopes and ridge crests with minor talus, scattered erratics, and a few rock outcrops. Black-colored areas represent extensive rock outcrop. Symbology on Qom: arcuate lines are drawn along the crest of large morainal ridges, and small polygons represent kettles.

## GEOMETRY OF THE ICE MARGIN

Because many end moraines in northwestern New Jersey are nearly continuous belts, they define the edge of the ice sheet both geometrically and temporally. The moraines clearly show that the margin of the Laurentide ice sheet was distinctly lobate at both a regional scale and local scale (Figures 51 and 58). Based on the tracing of end moraines from the valley floor onto adjacent uplands, the surface gradient of the Kittatinny Valley lobe varied between 125 and

290 feet per mile within the first few miles from its margin. In northeastern Pennsylvania, Crowl and Sevon (1980) determined that the slope at the terminus of the Laurentide ice sheet varied from 80 to 405 feet per mile with a "best measure" estimated at 225 feet per mile. The trace of the moraines and estimates on the surface slope of the ice sheet show that extensive lobation as postulated by Ward (1938) did not occur.

End moraines are used to construct a chrono-morphostratigraphic framework for the late Wisconsinan deglaciation of northwestern New Jersey. Not only do they provide direct evidence of ice retreat (snapshot), but they also constrain the reconstruction of other ice-retreat positions based on the location of outwash heads, meltwater channels, and glacial lakes. The geometry of ice-retreat positions indicated by the moraines shows that the Laurentide ice sheet retreated in a systematic manner to the northeast, its margin consisting of two distinct sublobes with the largest in Kittatinny Valley and a smaller sublobe in Minisink Valley.

Because the Terminal Moraine and recessional moraines in northwestern New Jersey were formed at the margin of an active ice sheet, the ice margins they define may not accurately reflect the geometry of the ice lobe if a significant amount of stagnant ice existed beyond the active glacier. Stagnant ice may consist of a valley sublobe many miles long (Ward, 1938; Crowl, 1971), or small detached blocks left by a retreating glacier. A margin of dead ice may have also bordered on the glacier's active margin, more or less synchronously wasting back with the retreating active glacier margin. This style of retreat, called *stagnation-zone retreat*, was originally defined by Currier (1941) and modified by Koteff and Pessl (1981) to describe deglaciation in New England (largely defined by the outwash heads of ice contact deltas laid down in proglacial lakes) where end moraines were not found. This style of deglaciation was proposed by Ridge (1983) and Witte (1988) for Kittatinny Valley and later modified by Witte (1991, 1997a) to account for a more active glacier margin and formation of end moraines.

## FORMATION OF END MORAINES

The study of an end moraine's morphology, course, and composition show that they are complex landforms, their genesis not easily described by simplistic depositional models. The character of the moraine shows that both active ice and stagnation play a role in their formation. The following definition, modified from Flint (1971) adequately describes the character of these features:

> *An end moraine is a ridge-like accumulation of drift built along any part of the margin of an active glacier. Its topography is initially constructional, and its initial form results from (1) amount and vertical distribution of drift in the glacier, (2) rate of ice movement, and (3) rate of ablation.*

Flint stressed the role of active ice transporting drift to the glacier margin, and the amount of drift in the ice sheet. Presumably, the more active the glacier and the more drift it contained, the larger the end moraine it will make. In addition, syndepositional and postdepositional modification of the moraine through ice shove, collapse due to melting of buried ice, and resedimentation of supramorainal materials chiefly by mass wastage, all act to give the recessional moraines their overall form. Figure 59 shows through a time lapse series of panels how end moraines may have formed. Based on their course, morphology, and composition, end moraines formed under the following set of conditions:

1) Active ice must be present to transport glacial debris to the glacier's terminus. This requires that the glacier must be temperate (warm-based) and that part of its forward movement occurs by basal sliding.

2) There must be a substantial zone of basal debris (basal dirty ice) at the glacier's sole. Sediment, chiefly contained in debris bands, is carried upward at the glacier's terminus by compressive flow, shearing, and folding, where at its surface, sediment is released from ice by melting. Supraglacial

sediment is further transported by mass movement, running water, and in rarer instances ice shove.

3) Because the transport of glacial debris is slow, the formation of thick end moraine requires that the glacier's margin must remain static (neither advancing nor retreating) or that it only oscillates throughout a narrow marginal zone (x $10^2$ to x $10^3$ feet). Length of the stillstand is estimated at one thousand to fifteen hundred years for the Terminal Moraine and several hundred years for the larger recessional moraines. These estimates are based on a late Wisconsinan retreat history of about four thousand years for New Jersey (Witte, 1997a).

4) Accumulation of debris across the glacier's terminus is variable due to variations in basal debris content and rates of basal sliding at the glacier's sole. This coupled with differences in ice thickness, due to local topographic relief, results in differential melting and the creation of a more uneven supraglacial surface.

5) Eventually the debris-covered terminus becomes a margin of stagnant ice with the leading edge of active ice moving back up the glacier. As melting proceeds in the stagnant zone, supraglacial debris is slowly let down on the land, the former supraglacial topography becomes inverted with sediment filled basins forming the higher areas and high-standing ice blocks or ice-cored ridges forming the low areas.

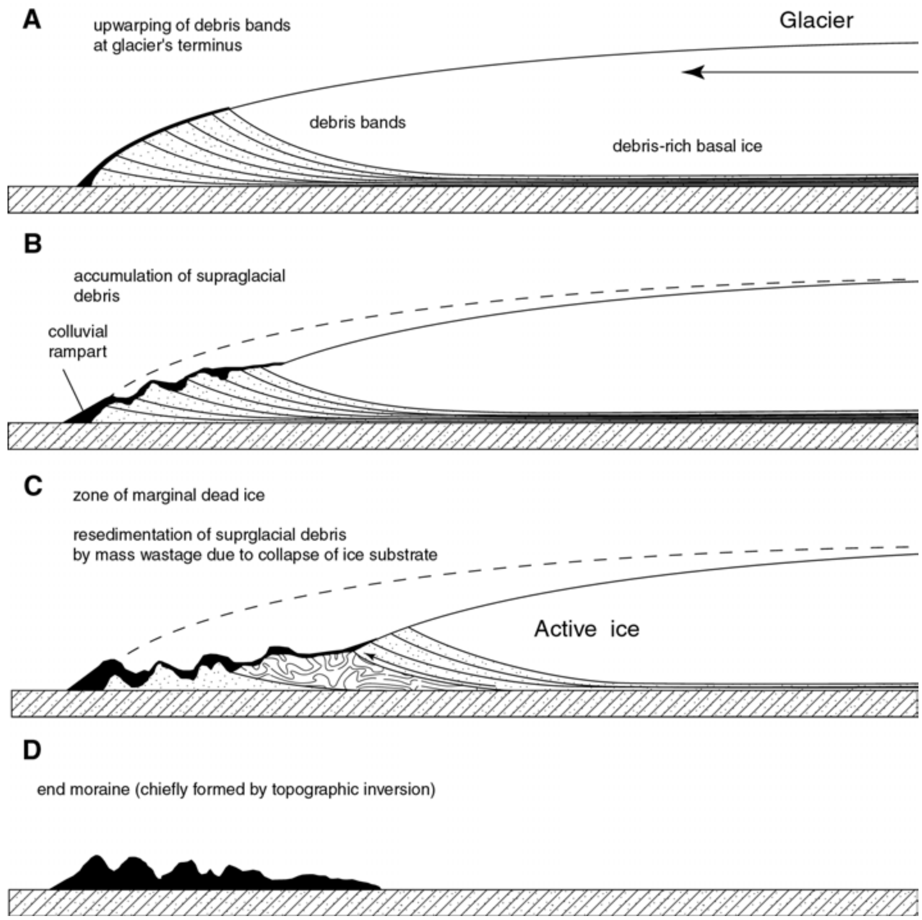



Figure 59. A sequence of panels showing how end moraines may have formed in New Jersey. Panel A - Active ice is present at the glacier's terminus. Sediment, chiefly basal debris transported along debris bands, is carried upward at the glacier's terminus by compressive flow, shearing, and folding of the debris-rich basal ice. Panel B - Sediment, released from ice by melting, accumulates along the glacier's margin where it collects in hollows and crevasses. Because the amount of debris is also variable, differential melting occurs, further increasing relief. Over time, and if the glacier's terminus remains in a relatively constant position, a debris blanket of varying thickness covers the marginal area of the glacier. Panel C - Continued accumulation of debris and differential melting causes the terminal area of the glacier to become very thin. This results in stagnation and the formation of a marginal zone of dead ice. Because the leading edge of active ice shifts backwards the transport of debris to the glacier's former active ice margin has ceased. Panel D - Gradually the supraglacial debris is let down on the land by the continued melting of dead ice and resedimentation by mass wastage. Except for colluvial ramparts and push ridges, morainal morphology is largely a product of topographic inversion where high areas of glacial ice covered by a veneer of debris now form the low areas (kettles and hollows) and low areas filled with a thick accumulation of debris now form the high areas (some morainal ridges, knolls and hillocks). Minor oscillations of the glacier's margin and several closely spaced stillstands may redistribute sediment, override stagnant ice, and leave additional stagnant ice forming a complex assemblage of morainal landforms. Figure modified from Flint (1971, Fig. 5-14).

6) Minor oscillations of the glacier margin can redistribute sediment, override stagnant ice, and leave additional stagnant blocks during the post-readvance phase of melting.

End moraines require active ice to form; yet, their overall morphology is the result of stagnation and redistribution of sediment by mass wastage. The lobate course of the moraines, their morphology, and evidence of glacial readvance suggests they were formed by 1) the transport of debris and debris-rich ice by the glacier at its margin, and 2) penecontemporaneous and postdepositional sorting and mixing of material by mass movement, chiefly resulting from slope failure caused by melting ice, and saturation and collapse of sediment. The source and mechanism of sediment transport is unclear. Most of the morainal material is of local origin. However, it is not known if the glacier was reworking drift at its margin or transporting sediment to its margin by direct glacial action. Inwash is not a viable mechanism because the larger deposits lie on mountain or ridge tops.

## CONCLUSIONS

End moraines in northwestern New Jersey were deposited at the margins of active ice lobes. They represent places where the glacier's terminus remained at a relatively constant position for one thousand to fifteen hundred years for the Terminal Moraine and several hundreds of years for the larger recessional moraines. They largely consist of till, formerly ice-entrained basal debris carried to the glacier's terminus, where it is released by melting. Over time the accumulation of debris and its redistribution across the glacier's periphery, chiefly by mass wasting, and—to a much lesser extent—ice thrusting, resulted in differential melting and stagnation of the glacier's marginal zone. The morphology of end moraines is largely the result of marginal stagnation and redistribution of sediment, chiefly by mass wastage. Moraine-parallel ridges may have formed by ice shove, or they are colluvial ramparts formed where debris was shed off the glacier's terminus. The Terminal Moraine and the Ogdensburg-Culvers Gap and Augusta moraines were laid down following a readvance of the Kittatinny Valley lobe. Marginal stagnation is caused by irregular topography and also by burial of the glacier's terminus, and not extensive melting and downwasting.

The reconstruction of glacial-lake histories, and delineation of ice-retreat positions marked by end moraines and outwash heads of ice-contact deltas show that the margin of the Kittatinny Valley lobe retreated in a systematic manner to the northeast. The interpretation of changes in ice flow during deglaciation and the presence of readvances marked by some end moraines show both show that live ice was present or not very far from the retreating stagnant margin throughout deglaciation. This evidence is in strong contrast to the concept of deglaciation by regional stagnation or valley ice-lobe stagnation as suggested by Connally et al. (1989).

Five ice margins, the Franklin Grove, Sparta, Culvers Gap, Augusta, and Sussex have been identified that mark major recessional positions of the Kittatinny Valley lobe. The Sparta, Culvers Gap, Augusta, and Sussex margins are traceable onto the New Jersey Highlands, and all margins, except Sparta, are traceable westward into Minisink Valley. All appear to represent halts in ice retreat, and some mark minor readvances of the Kittatinny Valley lobe. The close spacing of the Culvers Gap and Augusta margins, their large traceable extent, and correlation with extensive end moraines and ice-contact deltas in western New Jersey suggests that climate, in addition to local topography, have influenced the retreat history of the Kittatinny Valley lobe.

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

by
Ron W. Witte

## INTRODUCTION

Several investigations on glaciofluvial terraces in Minisink Valley (Figure 60) 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 (in press) near Stroudsburg, Pennsylvania, and Ridge (1983), Witte, (1988, 1991, 1997a), and Stone et al. (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 investigations have developed a

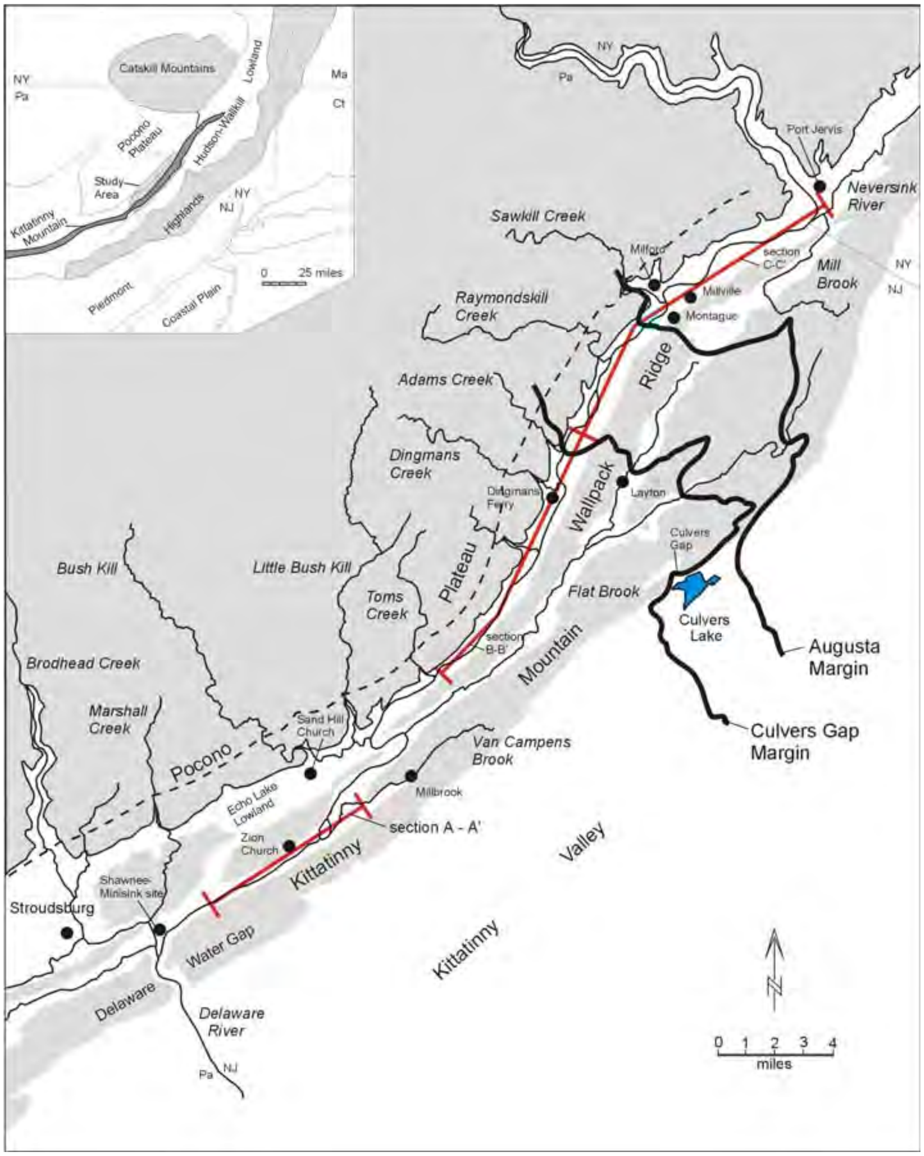



Figure 60. Physiography of Minisink Valley and surrounding area, the location of towns, streams, and lakes named in text, and position of longitudinal profiles shown on Figures 63a, b, and c.

Witte, R.W., 2001, Late Wisconsinan deglaciation and postglacial history of Minisink Valley: Delaware Water Gap to Port Jervis, New York, *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. 99 - 118.

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. This paper will examine the late Wisconsinan morphostratigraphic units in Minisink Valley to find out the nature of deglaciation there.

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 have been 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-sinuous meandering form. Archeological studies in the lower terrace have produced cultural artifacts that date from late Pleistocene through to historic time. Radiocarbon dating, microfaunal analyses, 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. This paper will examine the alluvial terraces in Minisink Valley, and examine the postglacial fluvial history of the Delaware River during the late Quaternary. 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) 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 that “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, as well as a lower and abandoned flood plain. Shortly afterwards, White (1882) reported on the glacial geology of Pike and Monroe Counties, Pennsylvania, and in it 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 (Figure 61) 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 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

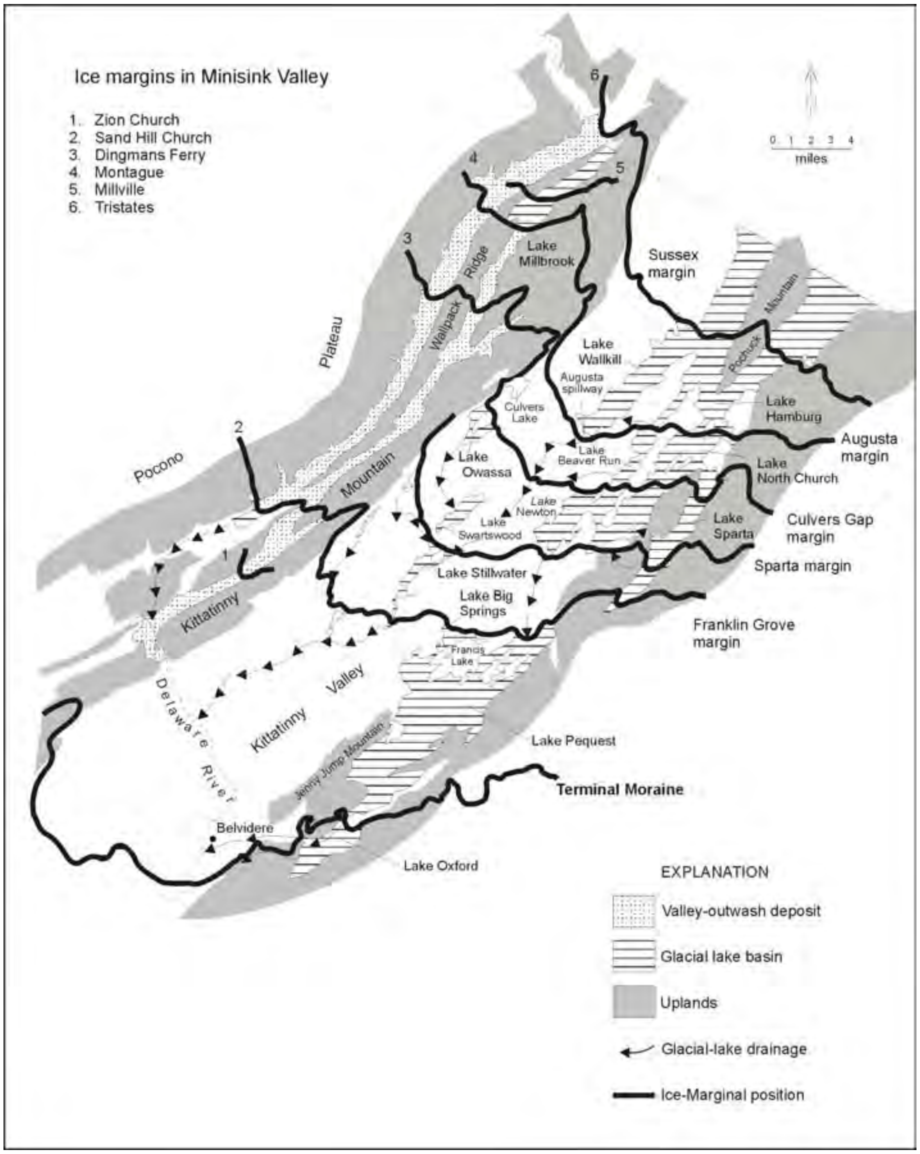



Figure 61. Late Wisconsinan ice-marginal positions in Minisink Valley and the upper part of Kittatinny Valley, location of large glacial lakes, and extensive valley-train deposits in northwestern New Jersey and northeastern Pennsylvania. Data modified from Epstein (1969), Crowl (1971), Ridge (1983), Witte (1991, 1997a), and Stone et al. (in press).

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 yr 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 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 (in review), and Stone et al. (in press) 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 (Figure 61). The Culvers Gap, Augusta, and Sussex margins are traceable across the New Jersey Highlands into the Newark Basin (Stone et al., in press), 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, 1985) and upper Shawnee Island site (Stewart, 1991). 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 (1991, p. 102) noted that “sedimentary sequences representing habitable landforms generally do not predate 8000/7000 B.C., 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 (1991) suggested that these paleosols reflect extended periods of landscape stability of about 500 to 1000 years.

## PHYSIOGRAPHY AND BEDROCK GEOLOGY

Minisink Valley (Figure 60), or the upper Delaware Valley, 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 Heilprin and Heilprin (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 translation of the word *Minisink* may be “the land from which water is gone” (Happ, 1938) or it may mean “stony country” (cited in Grumet, 1991, p. 176, as a personal communication from James Rementer in 1989). Perhaps the Minisink Lenape were the first geomorphologists to have worked in this area. The valley was also the proposed 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 consists of Silurian and Devonian strata that dip northwest and form a southwest-trending homocline (Drake et al., 1997; Sevon et al., 1989). The Delaware River (Figure 60) 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.

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 (450 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 (75 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**

Morainal deposits in Minisink Valley include the Dingmans Ferry, Montague, and Millville moraines (Figure 61). 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 lies only 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 et al. (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. (For a more comprehensive discussion on the recessional moraines in northwestern New Jersey, see Witte, this guidebook p. 81)

**Deposits of Glacial Meltwater Streams**

In Minisink Valley, sediment carried by glacial meltwater streams 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 62.

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 (Figure 61). 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, 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

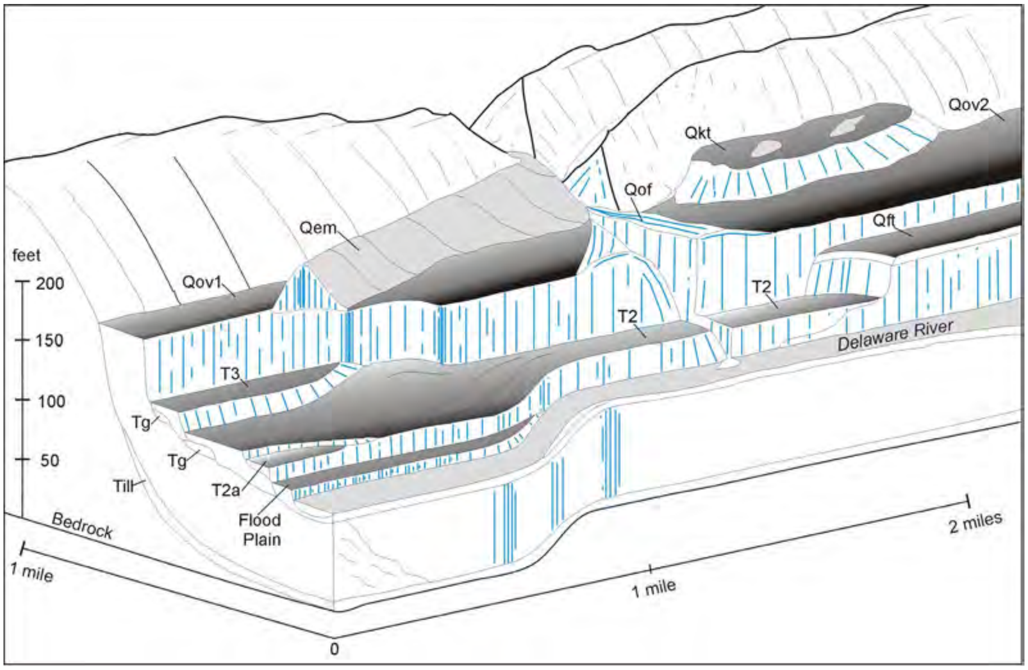



Figure 62. Assemblage and position of glacial and alluvial landforms in Minisink Valley. List of units: Qem = recessional moraine, Qov = valley train (lower number represents older deposit), Qkt = kame terrace, Qof = outwash fan, Qft = meltwater terrace, T3 = abandoned Pleistocene flood plain, T2 and T2a = abandoned Holocene flood plains, Tg = buried gravel cut terrace.

their upstream to downstream parts, the texture of the valley trains decreases from boulder-cobble-gravel to cobble-pebble gravel. 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 (Figure 63a, 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 1997b) 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 feed Minisink Valley. They may be as high as 560 feet (170 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. (The borough of Milford is situated upon the latter.) 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 $10^o$ 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.

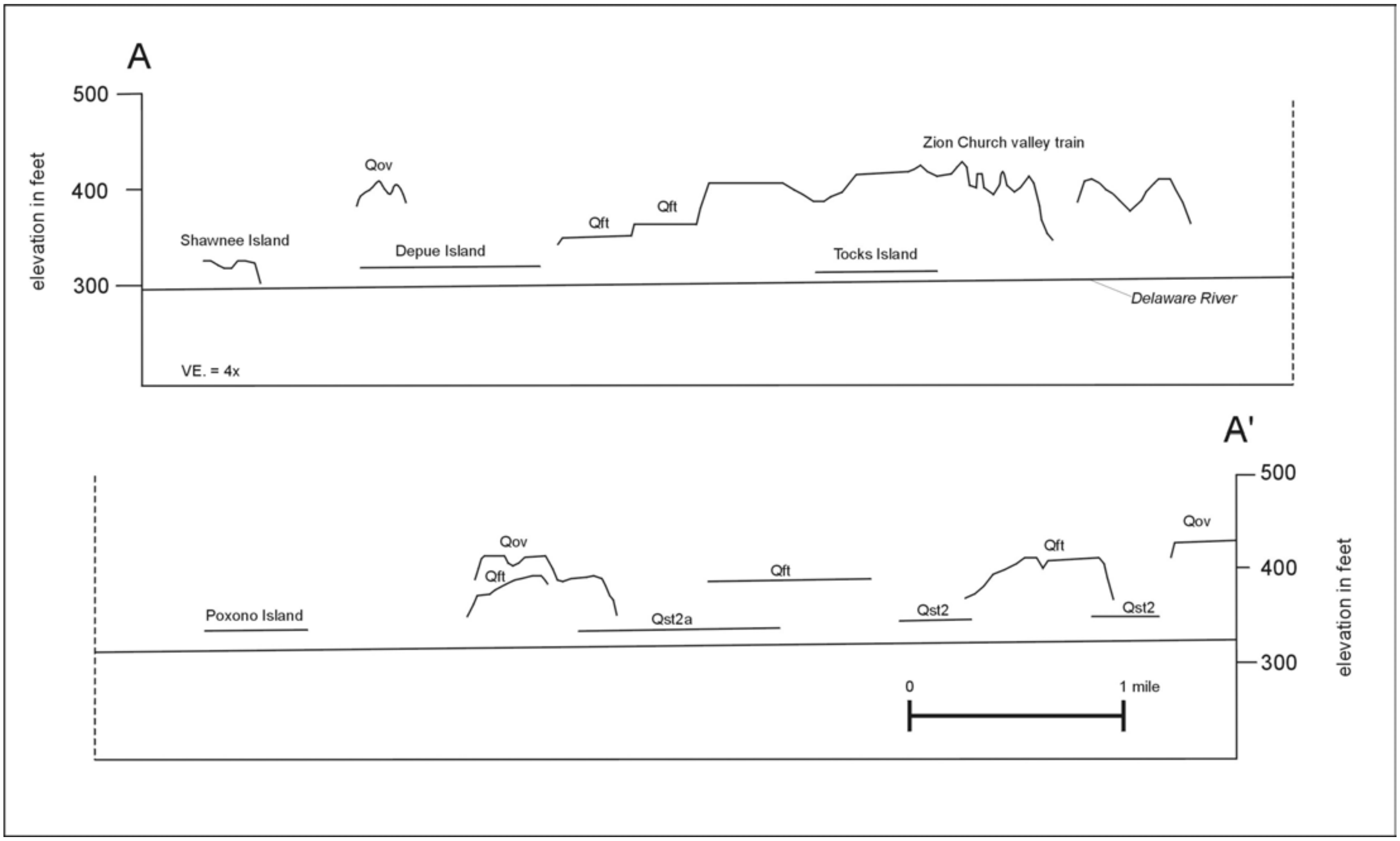



Figure 63a. Longitudinal profiles (section A - A ') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Bushkill and Flatbrookville, PA-NJ, 7-1/2 minute quadrangles. Location of section is shown in Figure 60. Profiles constructed by projecting elevation and contacts to a centerline 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.

Meltwater-terraces in Minisink Valley (Figure 62, and Figure 63a, 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.

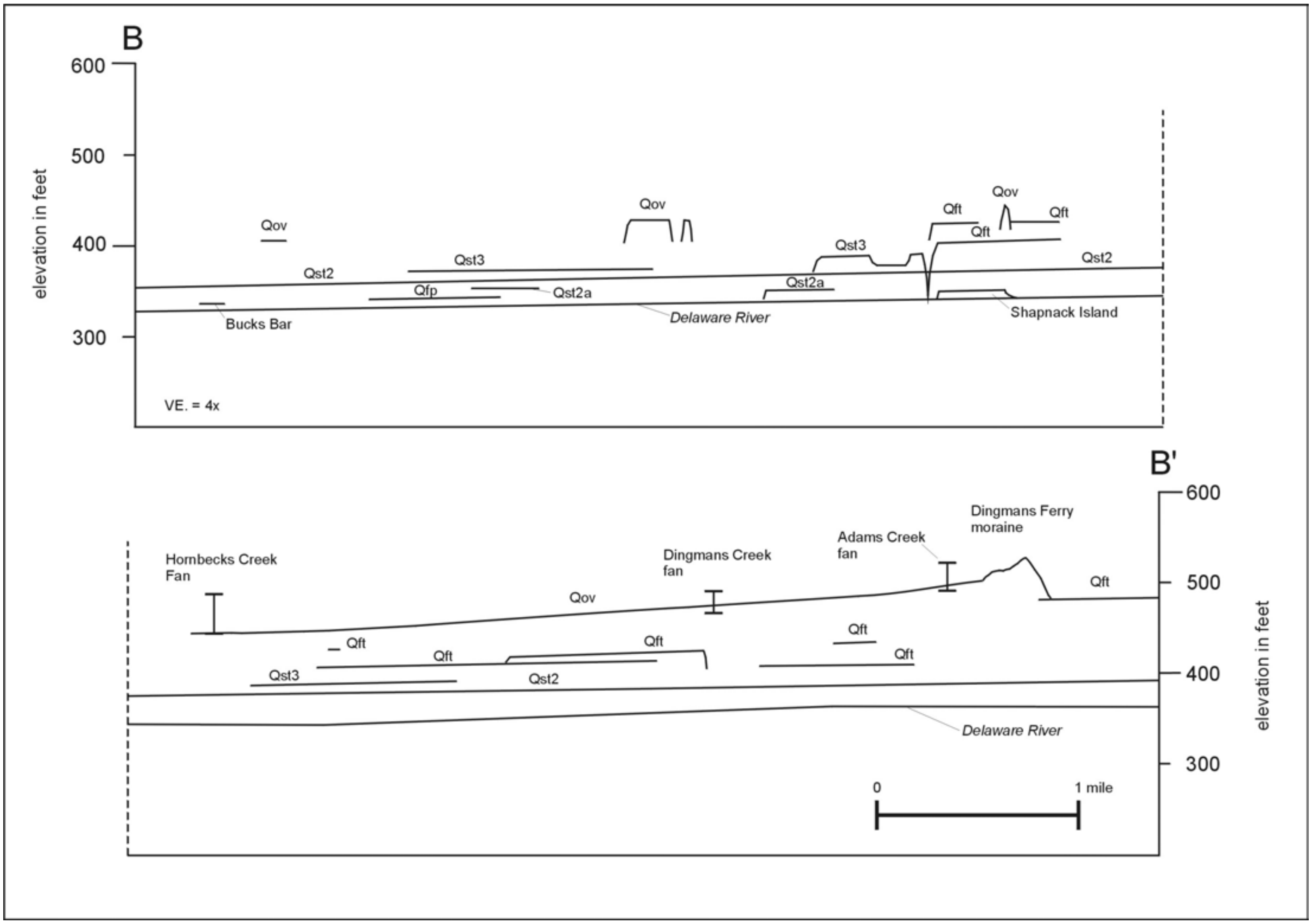



Figure 63b. Longitudinal profiles (section B - B ') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Lake Maskenozha and Culvers Gap, PA-NJ, 7-1/2 minute quadrangles. Location of section is shown in Figure 60. Profiles constructed by projecting elevation and contacts to a centerline 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, Qfp = modern flood plain. The range in elevation shown for the outwash fans represent the distal and proximal parts of their plains projected perpendicular to the section.

## 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 ± 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 ± 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 communication, 1996) indicates a minimum age of deglaciation at 19,340 ± 695 yr B.P. (GX-4279). Similarly, basal-organic material from Francis Lake in Kittatinny Valley, which lies approximately 8

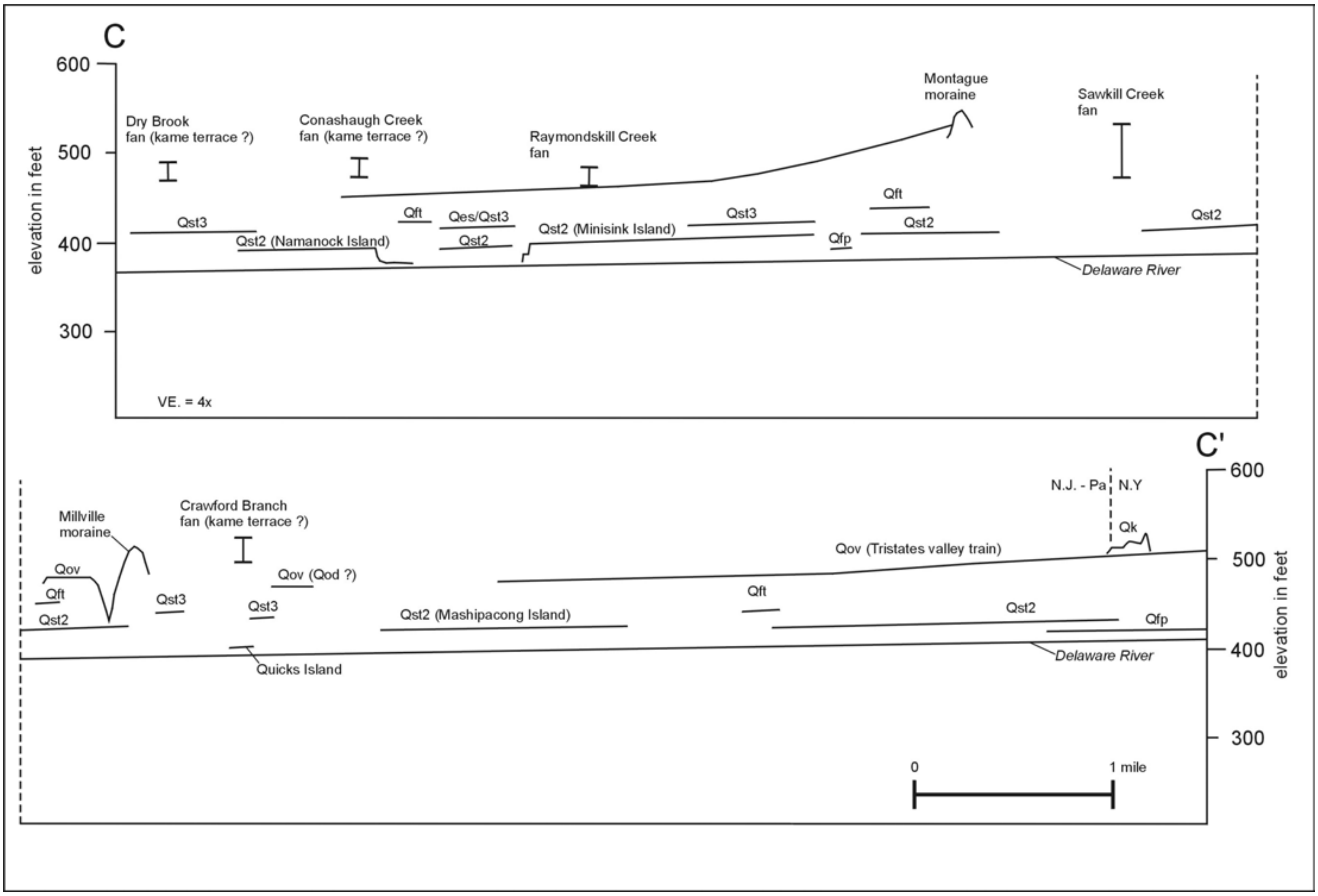
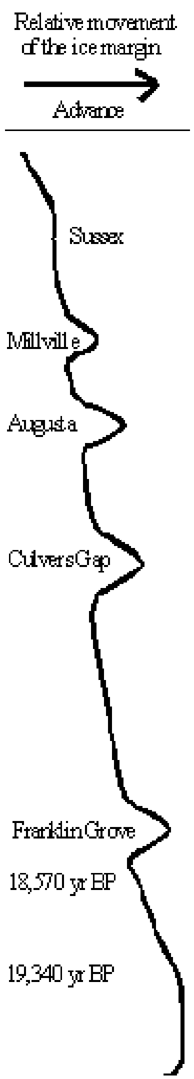



Figure 63c. Longitudinal profiles (section C - C ') of glacial outwash and postglacial alluvial terraces in Minisink Valley, Milford and Port Jervis South, PA-NJ, 7-1/2 minute Quadrangles. Location of section is shown in Figure 60. Profiles constructed by projecting elevation and contacts to a centerline 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 not labeled: Qov = valley train deposit, Qk = kame, Qft = meltwater-terrace deposit, Qst3 = abandoned Pleistocene flood plain, Qst2 = abandoned Holocene flood plain, Qfp = modern flood plain. The range in elevation shown for the outwash fans represent the distal and proximal parts of their plains projected perpendicular to the section.

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 (4.8 km) 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 (cited in 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 Wallkill moraine by Connally and Sirkin (1973) suggested 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 3.




| Minisink Valley | | Wallpack Valley | | Kittatinny Mountain | | | Kittatinny Valley | | Kittatinny Valley | |
|---|---|---|---|---|---|---|---|---|---|---|
| Local ice margin | Deposits | Local ice margin | Deposits | Local ice margin | Deposits | | Local ice margin | Deposits | Local ice margin | Deposits |
| | | | | | | | | | **Wallkill River** | |
| Tristates | valley Train | | | | | | **Papakating Creek** | | Sussex | ice-contact delta Lake Wallkill |
| Millville | moraine | Shimers Brook | Ice-contact delta | Steeny Kill Lake | moraine | ? | Armstrong | ice-contact delta (Lake Wallkill) | Hamburg | ice-contact delta Lake Hamburg |
| Montague | moraine and valley train | Montague | moraine and valley train | Montague and Augusta | moraine | | Augusta | moraine and ice-contact delta | North Church | ice-contact delta (Lake North Church) |
| Dingmans Ferry | moraine and valley train | Dingmans Ferry (Layton) | Moraine and valley train | Dingmans Ferry and Ogdensburg-Culvers Gap | moraine | | **Paulins Kill** | | Ogdensburg | moraine and ice-contact delta (Lake Sparta) |
| | | | | | | | Culvers Gap - Balesville - Lafayette - Germany Flats) | moraine and ice-contact deltas (Lake Newton) | | |
| | none observed | | none observed | | none observed | | Middleville | ice-contact delta | Sparta | ice-contact delta (Lake Sparta) |
| Sand Hill Church | ice-contact delta | **Van Campens Creek** | | | none observed | | Franklin Grove | moraine and valley train | **Pequest River** | |
| | | Mill Brook Village | moraine | | | | | | Turtle Pond - Andover | ice-contact delta |
| Zion Church | valley train | | | | none observed | | Hainesburg | ice-contact delta | Tranquility | ice-contact delta (Lake Pequest) |
| | | | | | | ? | Terminal Moraine | | | |

Table 3. Correlation of major ice margin positions and morphostratigraphic units in the the Kittatinny Valley and Minisink Valley areas. Cross-hachured boxes represent ice margin retreat from the drainage basin, and shaded boxes lie outside of the study area. Only the names of the larger glacial lakes are listed with the correlative morphostratigrapic unit. Ice movement and table boxes are not drawn to scale for distance of ice movement and time. List of works used to construct table. Kittatinny Valley - Ridge (1983); Stanford and Harper (1985), Witte, (1997a), Kittatinny Mountain, Wallpack Valley, and Minisink Valley - Minard (1961), Crowl (1971), Sevon et al. (1989), and Witte (1997a). Radiocarbon dates - 18,570 yr. BP (Cotter et al., 1986) and 19,340 yr. BP (Cadwell, written commun., 1996).

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 (1991), and Dent (1991). Cotter (1983), using radiocarbon dating and pollen stratigraphy, showed that the younger dates in the Minisink Valley area where 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 Valley, 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 62.

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 of glacial ice as envisioned by Happ (1938) and Crowl (1971). The longitudinal profiles of valley-train terraces and their downstream continuation from their heads (Figure 63a, 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 (Figure 61, 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 Valley lobe. The uncollapsed parts of its head-of-outwash lie at approximately 450 feet (135 m) above sea level (Figure 63a). Downstream, most of the valley train has been eroded by later meltwater and postglacial stream action.

The Sand Hill Church ice margin (Figure 61, 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, NJ) 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 1997c) confirm the existence of these deep depressions.

The next ice-recessional position is marked by the Dingmans Ferry moraine and valley train (Figure 61, Table 1, and Figure 63b). 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 (Figure 61, Table 1, and Figure 63c). 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 (Figure 61, Table 3). 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. (Figure 61, Table 3, and Figure 63c). 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 3) 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 (Figure 62). 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 either 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

Alluvial terraces in Minisink Valley consist of two abandoned flood plains that cover large parts of its floor. Additionally, buried cut terraces mark the former elevated position of the postglacial Delaware River. Taken together these deposits define a postglacial history of incision, and episodic flood-plain deposition punctuated by periods of land stability and soil formation. At the close of the Pleistocene, three factors seem to have had the greatest impact on the late glacial to early postglacial fluvial history of Minisink Valley. These are: 1) decrease in stream discharge due to the retreat of the Laurentide ice sheet out of the Delaware River drainage basin, 2) regional tilting of the land surface due to delayed isostatic rebound, and 3) the overall reduction in sediment supply due to floristic evolution in the drainage basin at the close of the Pleistocene.

## Geology of Alluvial Deposits

The modern flood plain, as it was defined by Leopold et al. (1964), lies as much as 12 feet (4 m) above the mean-annual elevation of the Delaware River in Minisink Valley. 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. In places they are covered by overbank sediment.

Stream-terrace deposits 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 Figure 62. Based on their continuity and uniform height above the Delaware they have been grouped in two distinct sets. The youngest (T2) terrace lies between 20 and 30 feet (6 and 9 m) above the mean-annual elevation of the 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 non-deposition and land stability. Stewart (1991) suggested that there are three buried-A horizons in Minisink Valley that form basin-wide chronostratigraphic units; these correlate with the Late Terminal/Archaic, late Middle Woodland, and Late Woodland cultural periods. Foss (1991) showed that most of the more mature paleosols he studied in Minisink Valley have cambic B horizons that indicate weathering of 500 to 1000 or more years. Argillic B horizons that require more than 4000 years to develop are absent.

The T2 terrace is an abandoned flood plain that covers large portions of the valley floor. Its higher parts lie next to the Delaware River and typically form a low levee. In a few places, the levee is well developed and it forms a prominent ridge that is as much as six 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 typically contains slack-water deposits and organic materials. In several areas 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. The T2 terrace may also include several levels as suggested by Wagner (1994). 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 T2 strata by Stewart (1991) and McNett et al. (1977) show that the base of the terrace in places is older than 11,000 yr B.P. Its upper foot (< 1 m) has been dated to historic times.

The oldest stream-terrace deposits form the T3 terrace; it lies 40 to 50 feet (12 to 16m) above the mean-annual elevation of the river, and as much as 20 feet above the higher parts of the T2 terrace. Longitudinal profiles of the postglacial terraces (Figure 63) parallel each other, although there is a slight suggestion of divergence going upstream. The T3 terrace 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 T2 terrace it is an abandoned flood plain. This material is similar to the levee and near levee facies described for the T2 terrace, but it has a slightly coarser texture. Paleosols are noticeably scarce, and in the few places where they were found they were represented by a truncated B horizon. In many places the T3 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 (3.6 m) 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 the T3 sediment had to be above the modern Delaware River and the late Pleistocene and Holocene rivers that laid down the T2 deposits

The T3 terraces are typically smaller than and flank the younger T2 terraces, and in some places, they lie surrounded by younger deposits. In several areas throughout Minisink Valley, gravel terraces lie between the T2 and T3 terraces. Apparently, the overlying materials were removed as the postglacial river cut down to its T2 level. No dates are available for the T3 terrace, 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.

## Record of Paleo-Environmental Change

The paleo-environmental record of Minisink Valley, from late glacial to modern time is summarized in Table 4. 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. Although, this report primarily involves events during the latter part of the Pleistocene, the entire postglacial record is summarized to give the reader a better understanding of the events that shaped Minisink Valley.

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

Table 4. 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 et al. (1977), (4) Stewart (1989). Delayed rebound from Koteff and Larsen (1989).

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 4). 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 paleo-environment interpreted from the pollen record shows that the study area initially was tundra with sparse vegetal cover. As the climate warmed, the open area of the tundra was replaced by 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 yr 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, 1991). Dent (1991) further suggested that a true boreal forest did not become established in Minisink Valley until 10,680 yr B.P. Mastodon remains, excavated from Leaps Bog (Hoff, 1969, and this guidebook, p. 146) near the lower end of the valley, were radiocarbon dated at 12,160 $\pm$ 180 yr B.P. (I-3929), and shows the presence of Pleistocene megafauna well into the later part of the spruce zone. Climatic warming is further evidenced by a large increase in the deposition of organic rich sediments (Cotter, 1983).

The pine pollen zone represents the period between 11, 250 and 9,700 yr B.P. and it marks the emergence of white pine as the dominant arboreal component. Dent (1991) 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, forming only 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 yr B.P. by Cotter (1983). However, Dent (1991) placed the break at 9250 yr 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, 1977), where charcoal associated with a hearth and Paleoindian artifacts yielded a date of 10,590 $\pm$ 300 yr B.P. (W-2994), and 10,750 $\pm$ 600 yr B.P. (W-3134), and the nearby upper Shawnee Island site (Stewart, 1991) which yielded a date of Early Archaic age (9380 $\pm$ 545 yr 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, 1991). The arrival of man in Minisink Valley shows that the climate, ecology, and the Preboreal to Boreal flood plain were suitable for occupation.

**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 very low sinuosity. 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 probably derived from local sources, eroded along the main reach of the valley and carried in by the many meteoric-sourced 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 62, the 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 T3 and T2 terraces, and radiocarbon dating of the T2 alluvial sequence indicate that the late glacial Delaware River was at least 30 feet (9 m) above the modern river.

In contrast to the late glacial river, the modern Delaware is a meandering stream of low sinuosity that is flanked by two abandoned flood plains (T3 and T2 terraces). The T2 alluvial sequences at the Shawnee-Minisink (McNett et al., 1977) and Upper Shawnee Island (Stewart, 1991) sites (Figure 63a) 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 meandering stream, represented significant hydraulic changes 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 yr B.P., based on the mapping and correlation of ice-marginal positions by Ozvath and Coates (1986) in the Western Catskill Mountains and by 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 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 Valley and to a lesser extent the Hudson Valley.

At some point in time 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 (1991) suggested the coarse gravel beneath the T2 terrace 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 (Table 4), the basal gravel is older than 11,000 yr 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 the T2 flood-plain deposits is not glacial outwash, but outwash reworked and incised by the postglacial Delaware River. The position and stratigraphy of the T3 terrace (Figure 62) show that it is older than the T2 terrace and it was laid down by a river that was higher than the T2 river. The T3 deposits represent 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 was fed by meltwater, and it had a braided channel form. Because the T3 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 T2 level appears to have been initiated by the onset of delayed isostatic rebound, and possibly a reduction in sediment supply due to the transition from a tundra to a closed boreal forest. The 14,000 yr B.P. maximum date for the start of rebound (Koteff and Larsen, 1989) and the 14,250 yr B.P. date marking the transition from herb to spruce pollen zones (Cotter, 1983) seem to be in accordance with the estimated age of the T3 terrace.

A large area of wind-blown sand on the T3 terrace just south of Minisink Island (Figure 64) provides additional evidence that the T3 terrace is of Pre-Boreal age. Here small sand dunes cover the T3 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 the T2 terrace next to and westward of the dune field. The position of the dune field suggests that it was deposited after the T3 flood plain was abandoned, but before the growth of an extensive cover of vegetation. The eolian sand may be reworked T3 material blown off a formerly larger and more extensive T3 flood plain.

**Conclusions**

Based on the estimated age of deglaciation for the Delaware River drainage basin, and a minimum age of 10,750 yr B.P. for the T2 alluvium, it is estimated that the T3 to T2 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. The T2 terrace represents episodic periods of alluviation throughout the Holocene. Leopold et al. (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 T2 flood plain outgrew its fluvial setting and eventually became abandoned, receiving sediment only during the greatest of floods.

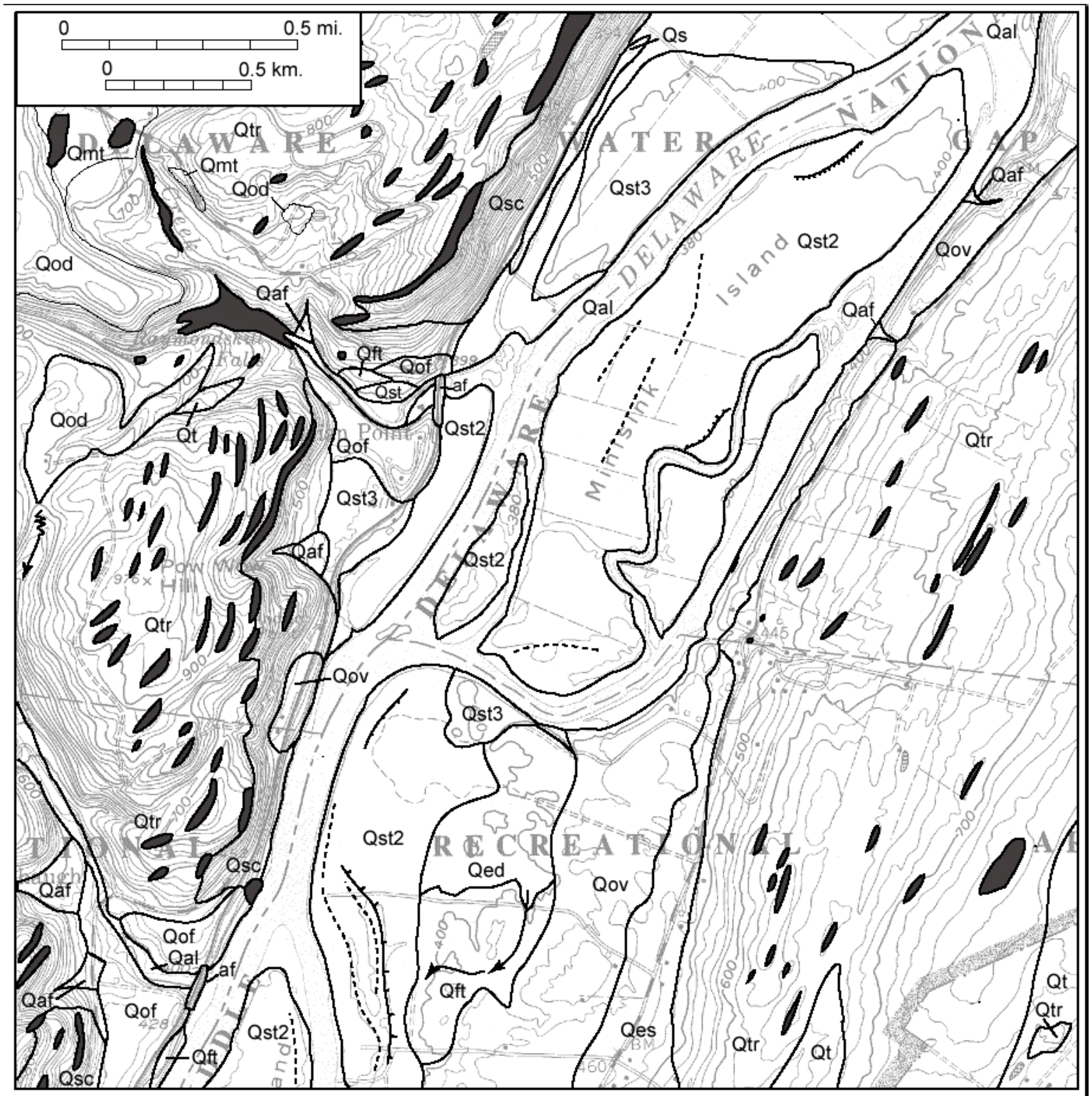



Figure 64. Surficial geologic map of part of the Milford, PA-NJ, 7-1/2 minute quadrangle near Raymondskill Creek, Pennsylvania. List of map units: af = artificial fill, Qal = alluvium, Qaf = alluvial fan, Qs = swamp and bog deposit, Qsc = shale-chip colluvium, Qed = sand dunes, Qes = thin sheet of wind-blown sand, Qst2 = abandoned flood plain (Holocene), Qst3 = abandoned flood plain (Pleistocene), Qft = meltwater-terrace deposit, Qov = valley-train deposit, Qod = ice-contact delta, Qf = outwash-fan deposit, Qt = thick till, and Qtr = thin till. Shaded areas represent extensive bedrock outcrop and the curved lines on the stream-terrace deposits represent channel scrolls. Surficial data from Witte (1997b).

# GEOLOGIC CONTROLS OF LANDSLIDES IN THE DELAWARE WATER GAP NATIONAL RECREATION AREA, NEW JERSEY-PENNSYLVANIA, AND LEHIGH GAP, PENNSYLVANIA

by
Jack B. Epstein

## *ABSTRACT*

Three types of landslides are recognized in the Delaware Water Gap National Recreation Area and adjoining Worthington State Park in northern New Jersey and northeastern Pennsylvania. These include soil slips on glaciated-polished bedrock surfaces, rockfalls that originated along fractures that parallel roads, and debris flows in glacial till. A soil slip occurred near Sambo Island on the Delaware River following heavy rain during October 20-21, 1995. Here, bedding planes of the Bloomsburg Red Beds are covered with a thin veneer of soil and glacial till in a moderately dipping northwest limb of an anticline in the Valley and Ridge physiographic province. The combination of heavy rain and lack of anchoring of the soil by tree roots that did not penetrate the polished bedrock surface resulted in the landslide. Similar geologic conditions (moderately steep bedding, glaciated-polished bedrock surfaces and shallow soil) can be used to determine areas of potential future landsliding. A debris flow in rain-saturated glacial till developed along a road cut in 1996 in a steep bank along a narrow tributary valley in the Pocono Plateau. Glacial till is common throughout the area and landsliding may be anticipated in areas where the bases of steep slopes are excavated. Two rockfalls occurred in New Jersey where the Old Mine Road parallels longitudinal joints near the crest of an anticline just north of Delaware Water Gap and on the northwest limb of an anticline opposite Tocks Island. These fractures are common in bedrock throughout the park. The runout of the till debris flow and one rockfall were mitigated with gabions. Stress measurements by the US Army Corps of Engineers on northwest-dipping bedding-plane faults in New Jersey near Tocks Island suggest that there is the potential for massive failure of rock above these structures should they be exposed by construction. Potential movement along cross joints in sandstone and conglomerate at Lehigh Gap, 29 miles southwest of Delaware Water Gap, have created a rockfall hazard that required mitigation.

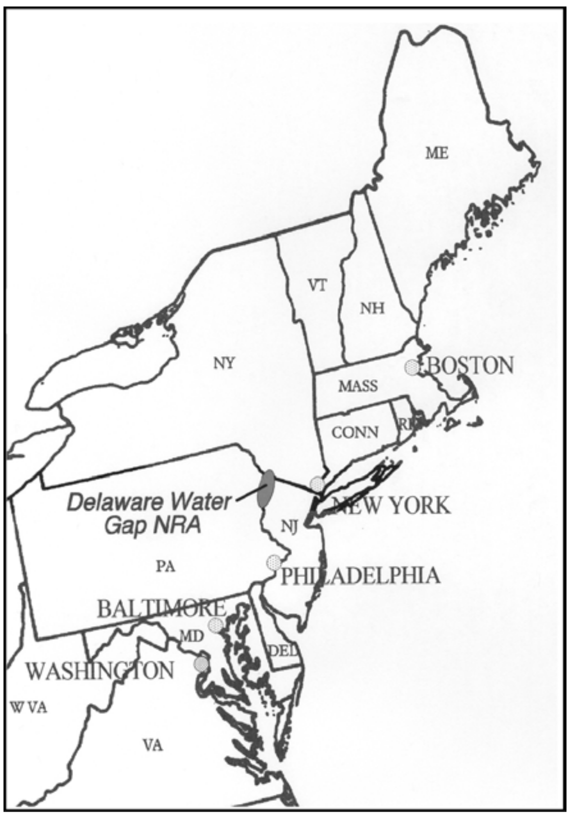



Figure 65. Index map showing the location of the Delaware Water Gap National Recreation Area within the heart of the northeast US population center.

## INTRODUCTION

The Delaware Water Gap National Recreation Area (DEWA) in Pennsylvania and New Jersey was established by an act of Congress in 1965, and lies within the heart of the Boston-Washington urban corridor. It is the largest National Park facility in the northeastern United States (Figure 65) and is the sixth most heavily visited NPS facility in the country with about 4 million visitors yearly. DEWA is about 40 miles long and includes a scenic and mostly undeveloped stretch of the free-flowing Delaware River between Port Jervis, New York, and the Delaware Water

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.

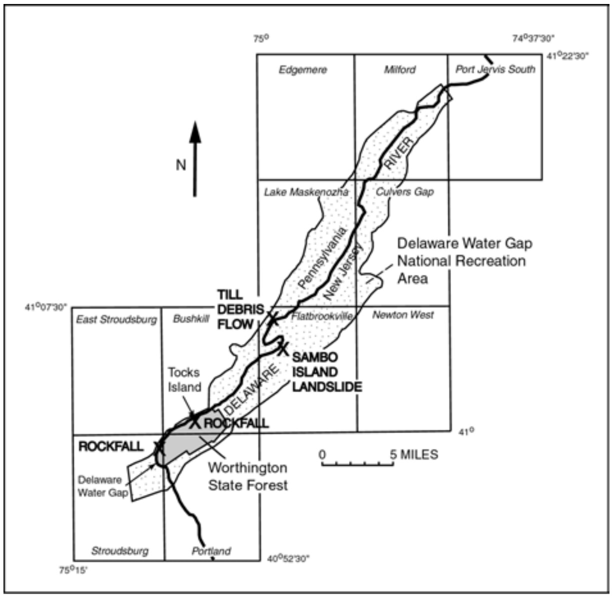



Figure 66. Index map showing landslides and topographic coverage in Delaware Water Gap National Recreation Area and Worthington State Forest, Pennsylvania-New Jersey.

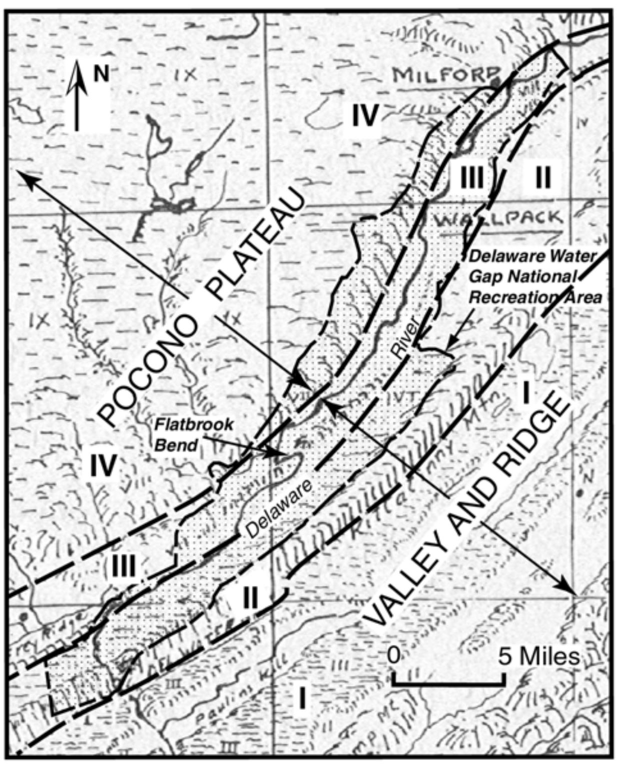



Figure 67. The Delaware Water Gap National Recreation Area spans two physiographic provinces, the Pocono (Allegheny) Plateau and the Valley and Ridge. The boundary is marked by gently dipping rocks of the Marcellus Shale and Mahantango Formation (Unit IV) of Middle Devonian age along cliffs northwest of the Delaware River, separated from Ordovician through Middle Devonian rocks (units III – I) to the southeast that are more complexly folded.

Gap in New Jersey and Pennsylvania. It occupies parts of eleven 7.5-minute quadrangles (Figure 66). The area offers a variety of recreational opportunities and opportunities to study the biologic diversity, cultural history, and geologic development of this part of the Appalachians.

Bedrock and surficial geologic mapping by the US Geological Survey, New Jersey Geological Survey and Pennsylvania Geological Survey has established a database useful for understanding slope instability within the recreation area and surrounding terrain. The instability has resulted in several types of landslide, posing a risk to people and property. Mitigation costs have been an estimated $150,000. Three distinct types of landslides were identified and studied: (1) a soil slip-debris flow on moderately dipping, glacially polished bedrock surfaces in the Bloomsburg Red Beds, (2) a debris flow in till on a steep slope, and (3) rock falls or slips generated along joints along the Old Mine Road (Figure 66). Many bedding-plane faults have been documented in the Bloomsburg Red Beds. During contemplation of construction of the Tocks Island Dam during the 1960's, the U.S. Army Corps of Engineers noted that the residual stress in these faults exceeded the hydrostatic stress. This suggests that if a bedding plane fault were exposed during construction, the entire body of bedrock above the fault had the potential for mass movement down the fault plane. This possibility should be considered in any future contemplation of highway construction, especially where bedding dips at an angle steeper than friction angle along bedding and daylights towards the construction site. A potential rock fall near the Appalachian Trail under National Park Service jurisdiction was alleviated at Lehigh Gap, 29 miles southwest of Delaware Water Gap.

## GEOLOGY OF THE RECREATION AREA

The Delaware Water Gap National Recreation Area spans two major physiographic provinces, the Valley and Ridge and Appalachian Plateau, the latter locally known as the Pocono Plateau (Figure 67). The rocks within the park area aggregate more than 8,000 feet in thickness and range in age from the Middle Ordovician to the Upper Devonian, approximately 440 to 380 million years ago. The structure in the rocks and the resulting landscape features trend northeastward. These rocks are varied in lithology and structure and, for convenience of discussion, they are subdivided into four units. Unit IV is within the Pocono Plateau, the others are in the Valley and Ridge province. Table 5 summarizes the lithologic and structural characteristics of the component formations within these units.

| Unit | Age | Stratigraphic Unit | Lithology and structure | Approximate Thickness (feet) |
|---|---|---|---|---|
| IV | Middle-Upper Devonian | Catskill Formation<br>Trimmers Rock Formation<br>Mahantango Formation<br>Marcellus Shale | Gray, red, and green sandstone, siltstone, and shale, coarsening upwards. Gently northwest dipping with minor low-amplitude folds. Base marked by poorly exposed shear zone. | 5,000+ |
| III | Upper Silurian-Middle Devonian | Buttermilk Falls Limestone<br>Schoharie Formation<br>Esopus Formation<br>Oriskany Group<br>Helderburg Group<br>Rondout Formation<br>Decker Formation<br>Bossardville Limestone<br>Poxono Island Formation | Heterogeneous units of limestone, shale, siltstone, sandstone, and dolomite in asymmetric folds, some overturned, with wavelengths about 1,200 feet and amplitudes of 250 feet in the southwest decreasing to gently dipping monoclines in the northeast. Base believed to be disharmonic and faulted. | 700-1,500 |
| II | Middle and Upper Silurian | Bloomsburg Red Beds<br>Shawangunk Formation | Gray and red sandstone, siltstone, shale, and conglomerate in folds, some overturned, with wavelengths of about 1 mile and amplitudes averaging 3,000 feet, becoming gentler to the northeast. Base marked by a major unconformity and faulting. | 3,000 |
| I | Upper Ordovician | Martinsburg Formation | Slate and greywacke in folds averaging 2,000 feet in wave length and 1,200 feet in amplitude. | 1,000+ |

Table 5. Description of rock units and their structural characteristics within the Delaware Water Gap National Recreation Area.

Unit I comprises slate and sandstone of the Martinsburg Formation of Middle Ordovician age, found only in the southwestern end of the Park. The Martinsburg erodes into rolling hills that slope down to the southeast towards the Paulins Kill which is underlain by carbonate rocks of Cambrian and Ordovician age. No significant landslides are present in the Martinsburg in the Park.

Unit II comprises resistant sandstone and conglomerate of the Shawangunk Formation which holds up Kittatinny Mountain, rising to altitudes above 1600 feet, with less resistant shale, siltstone, and sandstone of the Bloomsburg Red Beds forming the northwest slopes. Slope failures include soil slips and debris flows in the Bloomsburg and rockfalls in both formations. Bedding-plane faults in the Bloomsburg have the potential for causing mass movement of overlying rock if construction daylights this zone of weakness.

The course of the Delaware River in the northeast section of the Recreation Area is on cherty limestone of the Buttermilk Falls (Onondaga) of Unit III and shales of the Marcellus of Unit IV. After cutting through unit III at the S-shaped Flatbrook Bend, the river flows on the weak shale and carbonate rock of the Poxono Island Formation (Unit III) and upper part of the Bloomsburg Red Beds (Unit II). Other than large talus blocks, probably of Pleistocene age, there are no major landslides in unit III. At Hibachi Rock (Appendix C, Stop 8) is a large dolomite boulder from unit III that has toppled into the

Delaware and is a favorite diving platform for swimmers.

Unit IV lies northwest of the Delaware River and forms the base, steep slopes, and tableland of the Pocono Plateau. The basal shales of the unit, which form the escarpment and steep slopes along US 209, are over steepened because of Wisconsinan glacial erosion and are constantly spalling off, forming shale-chip rubble at the base of the cliffs. (See Field Conference STOP 10, Day 2, for a discussion of these deposits.) One landslide is reported here in glacial till overlying this unit.

Figure 68. Sambo Island landslide, developed on glacially polished bedrock bedding surfaces, extends for more than 600 feet on a slope that averages about 36°. It originated as a soil slip, entrained additional soil, glacial till, and trees downslope, and spread out into a debris fan at the bottom after encountering a rock rib formed by a stratigraphically higher bedding surface. Note the leveed channel on the lower right.

**SAMBO ISLAND LANDSLIDE**

During October 20-21, 1995, heavy rain fell in the Delaware Water Gap area generating a landslide on the anticlinal ridge south of Sambo Island along the Delaware River (Figure 68). The ridge is composed of red siltstone, shale and sandstone of the Bloomsburg Red Beds (unit II, Figure 67). The northwest slope of the ridge is a dip slope with bedding dipping gently at the crest at about 850 feet altitude, increasing to nearly 40° farther down the slope (Figure 69). The bedding surface exposed at the landslide site has been polished and striated by glacial erosion (Figure 70). Scattered outcrops of bedrock dot the ridge, but in most places bedrock is covered by three types of surficial materials: (1) soil composed of shale chips and organic matter; (2) large blocks of sandstone, and (3) till. The glacial erosion and till are the result of action by the last glacier that departed from this section of the Delaware River less than 20,000 years ago. The soil and sandstone debris formed subsequent to glacial retreat.

The soil cover in the slide area averages about two feet thick (Figure 71), ranging up to 8 feet thick. It consists mostly of rock chips, weathered from Bloomsburg shale and siltstone, averaging about 1 inch in length, and mixed with fine organic matter. Where the underlying bedrock is sandstone, angular blocks as much as 20 feet long are produced by spalling off from the bedrock surface and by transportation down slope. Slow mass wasting by creep is evidenced by many trees that are bowed at their bases. Production of the weathered fragments is facilitated by cleavage, bedding parting and joints in the shale-siltstone, and by bedding parting and joints in the sandstone. Most trees in the immediate slide area have an accumulation of rock chips plastered on their up-hill side to a height of two feet, indicating fairly recent downward movement of these materials.

Glacial till (Figure 72) is present in patches along the entire slope of the mountain and is more than 20 feet thick in places where it blankets large areas of bedrock. Hummocky topography along much of the lower slope of the mountain is suggestive of old landslide deposits. At the slide area the till is thickest at the base of the slope where it is nearly 5 feet thick (Figure 73). The material is a moderate brown (5YR4/4) clay-silt till with subangular to rounded cobbles and boulders as much as four feet long in the slide area. The clasts include a variety of rock types, including quartzite and sandstone from the Shawangunk Formation, Bloomsburg Red Beds, and Catskill Formation; and limestone and chert from various stratigraphic units. Some igneous boulders are syenite derived from an intrusion at Beemerville, NJ, 17 miles to the northeast. A thin veneer of soil containing red shale chips overlies the till in many places. For several hundreds of feet downstream, riffles in the Delaware River are due to large erratic till boulders that may have been emplaced by older landslides. Some of these boulders are as much as six feet long.

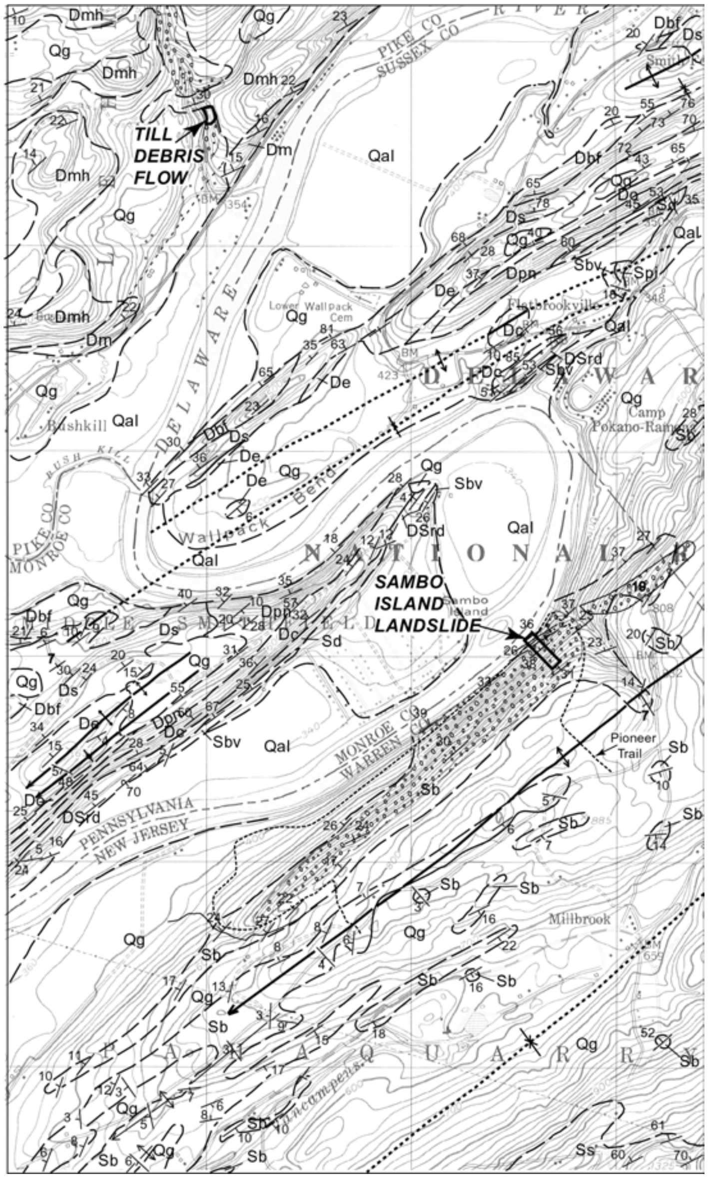
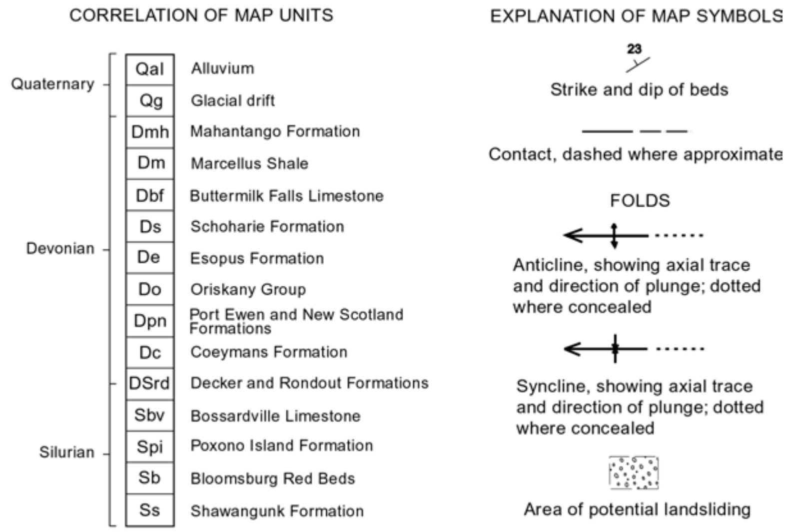
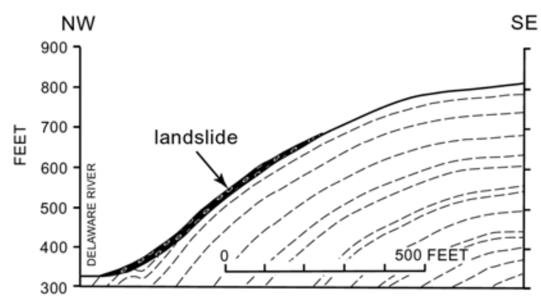



Figure 69. Geologic map of the Wallpack Bend area, Flatbrookville quadrangle, NJ-PA, showing location of Sambo Island landslide (with a cross section) and till debris flow. The area of potential landsliding south of the Delaware River is delimited.




Cross section through the Sambo Island landslide. Dashed lines are bedding.

Between 11 PM, October 20, 1995, and 3 PM, October 21, a total of 3.3 inches of rain fell near the site of the landslide (Figure 74), during two periods separated by about 6 hours. Immediately following the rain the river's discharge increased nearly ten-fold, peaking at about 27,000 cfs (Figure 75). During the 16 hours, there were two separate periods of heavy rain which saturated the thin soil making it unstable and causing it to slide on the polished bedrock surface. The contrast between the loose porous soil and the hard smooth bedrock surface was favorable for such failure because of a buildup of pore-water pressure over the less permeable bedrock surface. During recent periods of rainy weather, water flows along the smooth bedrock from under the soil cover.

The slide initiated in about 1.5 feet-thick, moderate-reddish-brown (10R4/4) shale- and siltstone-chip soil and organic matter at an altitude of 680 feet, encompassing an oblong area 37 feet wide and 90 feet long (Figure 76). The bedrock here forms a dipslope (34$^{o}$), is composed of red siltstone, and is partly burrowed and mudcracked. The surface is highly polished and striated (Figure 70), glacial


Figure 70. Glacially polished Bloomsburg bedding plane in the initial failure area of the Sambo island landslide. Bedding slopes 32$^o$ NW. Glacial striae trend S 40$^o$ W, parallel to the trend of the ridge.

Figure 71. Pressure ridge (arrow), about 3 feet high, showing downslope slip of soil adjacent to the initial failure area of the Sambo Island landslide. Note the lack of penetration of tree roots into the polished bedrock.

Figure 72. About 4.5 feet of glacial till overlies Bloomsburg bedrock just below the initial failure area. A thin veneer of soil containing red shale chips overlies the till.

Figure 73. Four-foot deep channel developed in debris fan in glacial till at the base of the Sambo island landslide.

movement towards S40$^o$W is indicated by the striae. At the bottom of the initiating area, the slide narrows to 28 feet at a three-foot-high pile of a debris-flow levee on the bedding surface, below which there is about 4.5 feet of moderate brown (5YR4/4) clay-silt glacial till overlying the bedrock (Figure 72). The soil thickens to about three feet in a pressure ridge (Figure 71). Initial movement of the slide compressed the soil to form the pressure ridge, then movement was retarded at this debris dam, and finally failure progressed by entrainment of material in a narrow slide area averaging about 40 feet wide, but reaching 50 feet in width, and for a total length of 600 feet to the bottom of the ridge. It extended out into the Delaware River for an additional 60 feet. Fortunately, no canoeists were present in the river at the time.

For most of the slide area, the soil has slipped off a single bedding plane, which remains fairly constant in dip (38$^o$ near river level, although complicated by a small fold there) down to an altitude of 440 feet. At about 130 feet above the river, higher bedding planes are exposed and the slope is offset upwards by about 10 feet. Glacial till (Figure 73) was encountered and the material spread out as a gullied debris fan (Figures 68 and 73), 40 feet wide at the top, and 125 feet wide at the bottom where it removed the Pioneer Trail just above the Delaware River. The debris included bedrock fragments, soil,

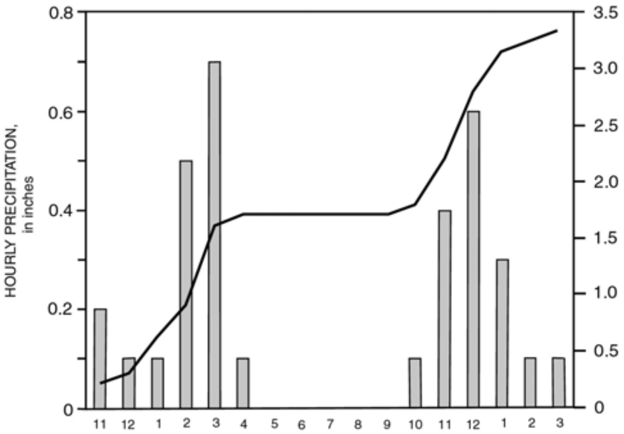



Figure 74. Bar graph showing hourly precipitation and curve showing cumulative precipitation between 11 PM, October 20, 1995, and 3 PM, October 21, 1995, at Dingmans Ferry, 11 miles northeast of the Sambo Island landslide.

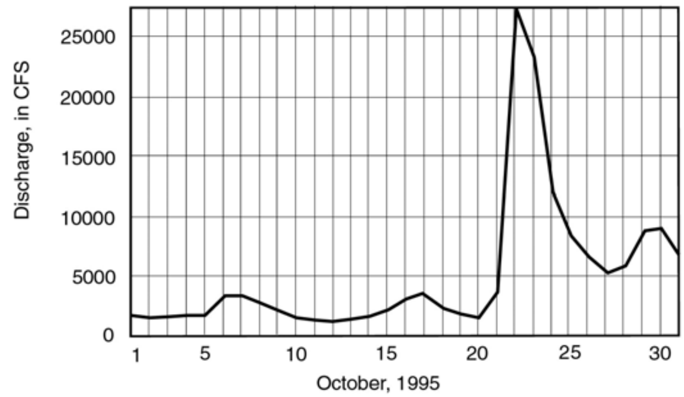



Figure 75. Streamflow for the Delaware River at Tocks Island during October, 1995.

Figure 76. Initial failure area at the top of the Sambo Island landslide. This area is 37 feet wide and extends down 90 feet to a pile of debris. The bedding dips $32^{o}$, paralleling the slope. The slide area extends 600 feet down to the Delaware River.

Figure 77. Irregular tension cracks (below dashed line) several inches high and a few tens of feet northeast of the initial failure area. Arrows show slight downward movement of the soil.

till, and trees, which slid out into the river. The total volume of soil displaced is calculated to be about 48,000 cubic feet (1,800 cubic yards).

The instability that produced the landslide probably began as sheet wash of soils, as evidenced by many small (less than 20 feet long) debris fans made of forest litter and piling up of soil up to two feet high on the uphill side of trees. Initial failure may have been sited at fractures, similar to ones present near the slide (Figure 77). Some of these tension cracks were developed along animal trails.

In summary, the landslide near Sambo Island is a soil slip of a thin veneer of shale-siltstone-chip gravel and glacial till that merged into a debris flow at the bottom. It is sited along the steep cutoff northwest slope of Kittatinny Mountain at the outer bend of a meander in the Delaware River. The presence of old landslide deposits indicate the potential for a landslide hazard here and elsewhere where similar conditions exist. The factors that make some of the area prone to instability and landsliding are (1) fairly steep slopes, more than 30 degrees in many places, (2) a thin soil cover, including glacial till, that rests on glacially polished bedrock surfaces, (3) tree roots, which otherwise could anchor the soil to bedrock, that do not penetrate the bedrock, (4) water seeps along the soil-bedrock boundary, making for a low-stress condition at this interface, (5) removal of the toe of the potential failure area by stream erosion (or the works of man), and (6) tension cracks similar to those seen near the landslide. Figure 69 outlines the slope adjacent to the Sambo Island landslide that meets these criteria for instability.

## ROCKFALLS ALONG THE OLD MINE ROAD: ROCKSLIP NEAR TOCKS ISLAND

The Old Mine Road in New Jersey cuts through siltstone and very fine-grained sandstone of the Bloomsburg Red Beds for a distance of 2,400 feet adjacent to Tocks Island (Figure 66). Smooth longitudinal joints whose strike parallels the road (averaging about N70$^o$ E) and which dip steeply towards the road (50-70$^o$ northwest) produce slabs averaging about 1 foot thick (Figure 78). These joints are smooth and regular and suggest sheet jointing or exfoliation resulting from expansion due to release of confining pressure due to rapid erosion along this stretch of the Delaware River.

The rock is further cut by irregular steeply dipping cross joints (striking 2$^o$-53$^o$ northeast; Figure 78*C*,*D*, and *E*) and fracture cleavage (averaging N60$^o$E, 65$^o$SE; Figure 78*D*). The road cut is as much as 25 feet high along this stretch. The slope above the road cut comprises ribs of bedrock outcrop and colluvial debris on slopes of about 38$^o$. The bedrock is massive and bedding is generally indistinct; it is recognized mainly by green reduced layers. The bedding dips more gently than the longitudinal joints, ranging between 16$^o$ and 44$^o$ northwest (Figure 78*D*). The bedding attitudes shown on the geologic map of the Bushkill quadrangle (Alvord and Drake, 1971) at this locality are in error; they record the longitudinal joints instead. Bedding has not influenced slope instability along this road cut. Figure 79 illustrates the geologic features at the site and their control on rock slips at this locality.

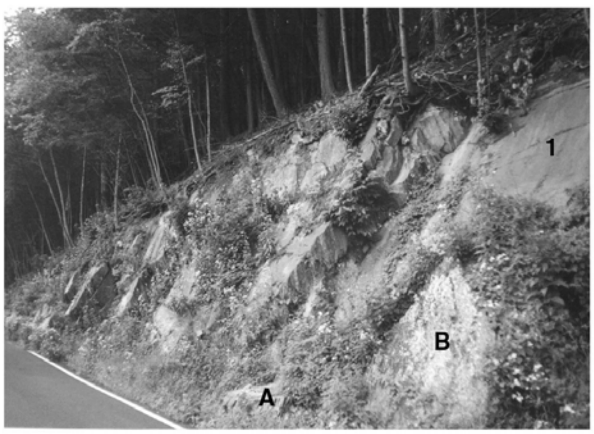
Because the toe of the rock mass has been removed by road construction (Figure 79), rock masses as much as 10 feet long have slid down the smooth joint surfaces. Figure 78*C* shows two masses that fell off the outcrop, partially blocking the road in 1999. The soil above the bedrock is thin, about 1 foot thick, and tree roots do not penetrate into the bedrock (Figure 78*A* and *B*). Thus, the soil is not stabilized by the trees and, just as at the Sambo Island landslide, there is potential for soil slip along this stretch of steep slopes. These small rock falls are a continuing problem.

## ROCKFALL IN DELAWARE WATER GAP

Worthington State Forest includes part of the gap of the Delaware River in New Jersey (Figure 66) and is surrounded by the Delaware Water Gap National Recreation Area. During the early 1980's a landslide, presumably a rockfall, removed a part of the west side of the Old Mine Road in the park, one thousand feet northeast of where I-80 crosses the Delaware River (Figure 80). The failure occurred in interbedded siltstone and shale and minor sandstone of the Bloomsburg Red Beds on the crenulated southeast limb of the Cherry Valley anticline which, 800 feet to the northeast, brings gray quartzite and conglomerate of the Shawangunk Formation to the surface. A 50-foot cliff parallels the road here and the slope above is steep, averaging about 43$^o$. Figure 81 is a diagrammatic cross section at the site. The landslide along the road probably occurred along sheeting fractures, but failure may have also been by rock toppling along steep cross joints.

*A*. Longitudinal sheeting joints dipping 48º NW and producing slabs between 10 inches and 3 feet thick. Shallow tree roots do not penetrate bedrock.

*B*. Bare Bloomsburg bedrock due to rockfall along sheeting joints. Note shallow soil and tree roots that do not penetrate the bedrock.




*C*. Sheeting joints (1) dipping steeply to the northwest and two rock slabs (A, 8 feet long; B, 10 feet long) that slid off during 1999. Other slabs that covered the road have been removed.


*D*. Longitudinal (sheeting) joints (1) dipping 57º NW and cross joints (2) separate the rock into slabs between 2 and 8 feet long. Rock cleavage (3) dips steeply southeast and bedding (dashed line) dips gently NE.

Figure 78. Exposures of the Bloomsburg Red Beds along the Old Mine Road in New Jersey, opposite Tocks Island, Delaware Water Gap National Recreation Area, showing how joints, sheeting, and cleavage affect development of rock slabs prone to sliding.

*E*. Irregular cross joints cutting steep longitudinal joint surface and separating rock into masses 2-8 feet wide.

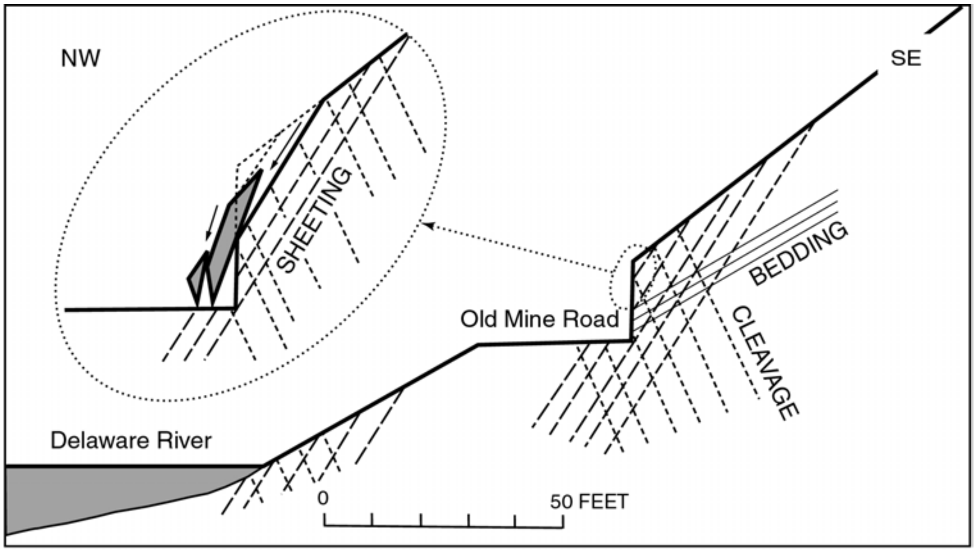


Figure 79. Cross section showing orientation of longitudinal joint set (sheeting) and cleavage in the Bloomsburg red Beds along the Old Mine Road opposite Tocks Island on the Delaware River, New Jersey. Because the road cut along the Old Mine Road is steeper than the dip of the joints, the toe of the rock mass has been removed making the area susceptible to rock falls along the joint surfaces. Cross joints, parallel to the plane of the section, aid in breaking the rocks into slabs as much as 10 feet long. Along the Delaware River, the surface slope is less than the dip of the joints, so there is little likelihood that rock falls will form.

The collapsed section of the road is now supported by a concrete gabion, about 100 feet long and 15 feet high (Figures 81 and 82), located along the abandoned railroad grade below the road, now used as a foot trail. A one-foot diameter weep pipe, about 30 feet northeast of the gabion, drains a wet area along the road. At the time of one visit, water coming out of the pipe was much less than the water falling from the rocks above and into a drainage along the road. At this locality the rock is highly fractured and veined in a zone 10 feet high that parallels a bedding-plane fault. Bedding trend is N68$^{o}$E, 41$^{o}$SE, and slickensides plunge 27$^{o}$, S34$^{o}$W, with steps indicating movement downdip to the southwest. Much water soaks into the substrate below, a possible concern for future failure. Just above the fractured zone is a set of sheeting fractures, similar to those at the Tocks Island rockfall, along which the failure probably took place (Figures 81 and 83). The strike of sheeting parallels the road at the failure site (Figure 84) and dips 45$^{o}$NW. Because it parallels the slope of the ridge and because its orientation is different than most other joints produced by tectonic forces, exfoliation is believed to be the origin of these joints.

Failure may have also occurred by separation and toppling along cross joints (Figure 85) that trend slightly west of north (Figure 84). These joints, prominent south of the gabion, are aligned about 25$^{o}$ from the trend of the road. Rock fragments and angular boulders have fallen into the cracks and over time have wedged the rock apart. The potential for future toppling is obvious, especially for those rocks on the steep hill high above the roadway.

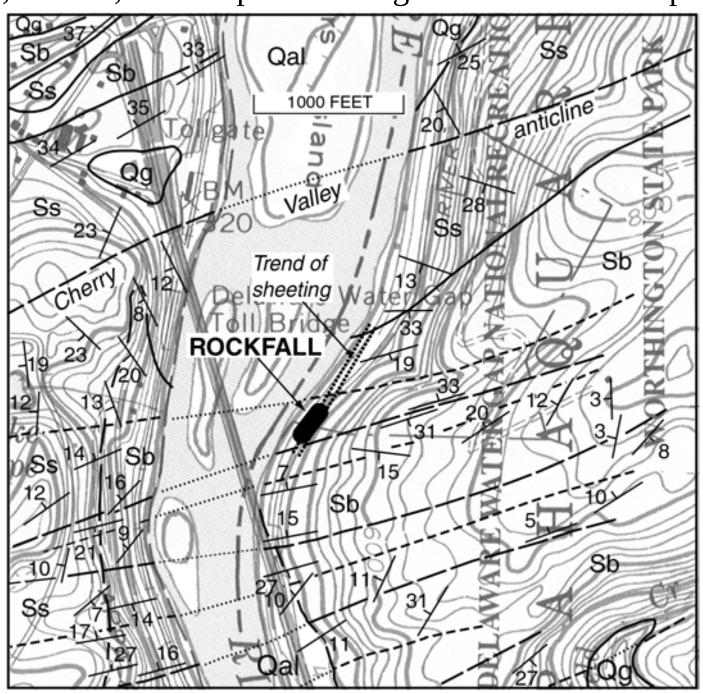


Figure 80. Geologic map of part of the Delaware Water Gap showing location of rockfall along Old Mine Road in relation to geologic structure. Heavy dotted line is trend of sheeting fractures. Qal, Holocene alluvium and stream terrace deposit; Qg, Wisconsinan glacial drift; Sb, Bloomsburg Red Beds; Ss, Shawangunk Formation. Solid lines, contacts. Long, dashed lines, anticlines. Short, dashed line, synclines. Modified from Epstein, 1973.

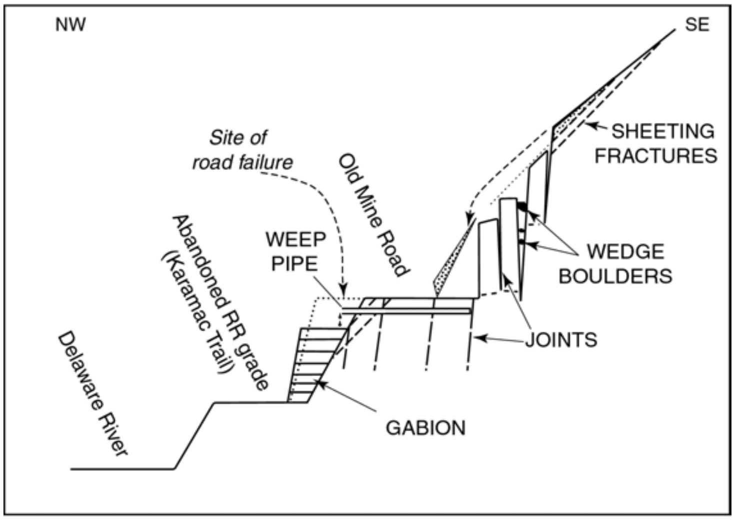



Figure 81. Generalized profile showing the relations of joints, sheeting, wedge boulders, and topography that resulted in rock sliding and toppling along the Old Mine Road in Worthington State Park, New Jersey. Dashed lines beneath road bed indicate joints and sheeting fractures that probably were related to the road collapse about 40 years ago. Additionally, the dotted block contributed to the failure by sliding off the sheeting joint. Dotted line indicates configuration of pre-rockfall topography.

Figure 82. Concrete gabion along abandoned railroad grade below the Old Mine Road in Worthington State Park.


Figure 83. Sheeting joints along which the landslide along the Old Mine Road near Delaware Water Gap probably occurred. The shear zone is often wet and water seeps into the road bed below, more than is emitted from the weep pipe shown in Figure 81.

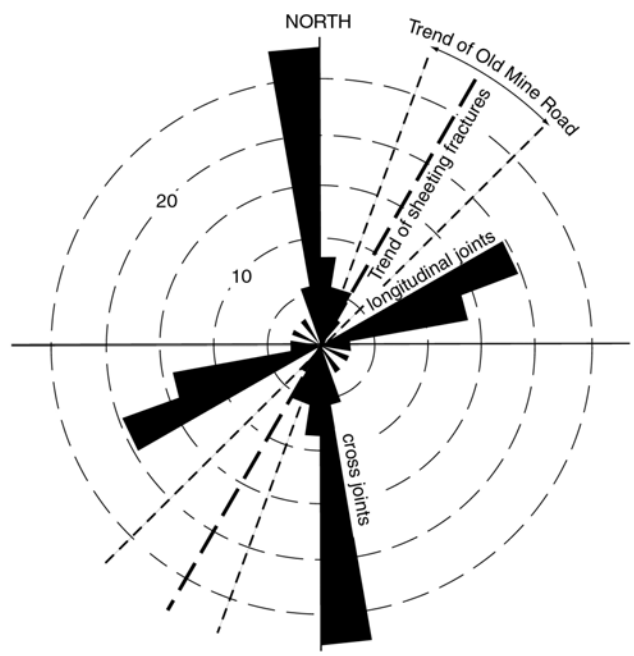



Figure 84. Rose diagram showing trend of longitudinal joints, cross joints, and sheeting fractures along Old Mine Road in Worthington State Park, New Jersey.

Figure 85. Irregular cross joints, trending slightly west of north and at an angle to the trend of the Old Mine Road (see Figure 84). Boulders in the fractures are wedging the block apart, creating the potential for toppling.

Thus, several geologic factors appear to be involved in two distinct types of failure and potential failure. These include joint sheeting dipping towards the roadway, shearing along a fault, and rock masses being forced apart by wedging along cross joints. Additionally, rock cleavage, which averages about N0$^o$E, 65$^o$SE, is nearly parallel to the Old Mine Road, creating another plane of weakness along which the rock may separate.

## BRODHEAD ROAD DEBRIS FLOW IN GLACIAL TILL

The Wisconsinan glacier retreated from the Delaware Water Gap area less than 20,000 years ago. It left behind a variety of geologic deposits, including sorted sand and gravel, mainly in the valleys, and glacial till, consisting of a heterogeneous mixture of clay, silt, sand, and boulders. If till becomes water saturated on moderately steep slopes, there is the potential for downhill movement of this material. A debris flow can form involving the till and trees and other vegetation on the slope. Movement may be rapid, becoming a hazard to property and life, or slow. A debris flow developed about 1996 on the east side of Brodhead Road, 1.3 miles northeast of Bushkill, PA, and 1500 feet north of US 209 in the Flatbrookville quadrangle (Figures 66, 69, and 86).

Figure 86. Debris flow in clay-silt till along Brodhead Road. The largest boulder in the till is two feet long.

The bottom of the slope was excavated by construction of the road, cutting out the toe, and expediting the sliding. The landslide is 60 feet high with the steeper head scarp 20 feet high. The angle of the original slope was 37$^o$, now steepened by the sliding in the upper part. The landslide occurred in a moderate- to dark-yellowish-brown (10 YR 5/4-4/2), poorly sorted, compact, clay-silt till with scattered boulders as much as 2 feet long, but averaging about 3 inches long. Boulders more than 5 feet long have been noted nearby. Stones were derived from underlying siltstone bedrock and gray and red sandstones from northerly formations including the Trimmers Rock and Catskill Formations. The most recent slide is 140 feet wide, but there are older slide scars extending for an additional 300 feet along the road. The slope is steep because it was eroded along the outside meander of the creek that now lies west of the road. Another smaller slide area lies

Figure 87. Gabion, about 450 feet long, along the glacial debris flow along Brodhead Road.

about 200 feet south along a cut bank on the opposite side of the creek. Two wire-mesh gabions, each 200 feet long and separated by 50 feet, were constructed along the road along the total slide area to prevent future movement onto the road (Figure 87).

The potential for future landsliding may be determined by creating a map (Figure 69) showing the occurrence of till and slopes that are greater than about 30 degrees, especially if construction cuts out the toe of the slope.

## BEDDING-PLANE FAULTS IN THE BLOOMSBURG RED BEDS AND THEIR POTENTIAL FOR FAILURE

Faults along bedding in the Bloomsburg Red Beds were recognized by Epstein and Epstein (1967, 1969) throughout eastern Pennsylvania. They consist of polished and slickensided bedding surfaces, occasionally in a zone as much as one foot thick. Steps on the slickensides invariably indicate that overriding beds moved to the west, regardless of the rock's position on a given fold. Wedges (small ramp thrusts) and dragged cleavage also corroborate this general sense of movement (see Figure 82 in Epstein and Epstein, 1967). These faults are commonly zones of weathering and ground water flow.

During engineering evaluation for the proposed Tocks Island dam after the Delaware Water Gap National Recreation Area was authorized by Congress in 1965, the U.S. Army Corps of Engineers (ACE) encountered 360 feet of laminated and massive, partly desiccation-cracked, red siltstone and shale with some greenish-gray mottling and minor gray quartzitic sandstone in a 600-foot long exploratory adit in the Bloomsburg at the base of Kittatinny Mountain in New Jersey near Tocks Island (Figures 66 and 88). The maximum stress level parallel to one of these faults was determined as 1,000 psi during excavation for the proposed spillway (**x** in Figure 88), and was attributed to the weight of the entire rock mass between the fault and the surface (Dan Parillo, ACE, written communication, 1970). The maximum stress level below the fault was 525 psi, indicating that there is little or no strength across the fault and if the toe were daylighted the entire uphill mass would move downhill, according to Parillo. Many of the bedding faults shown in Figure 88 are zones of abundant water flow.

Even though the Tocks Island dam was never constructed, there should be concern about future road building or other construction along the Old Mine Road in New Jersey. Should any of the bedding faults be intersected during road construction, there is the potential for massive failure along these zones of weakness. There are many of these zones in the Bloomsburg and individual potential construction sites need to be analyzed for their presence. One such fault zone, which may have contributed to failure, is the rockfall in Worthington State Park, discussed above.

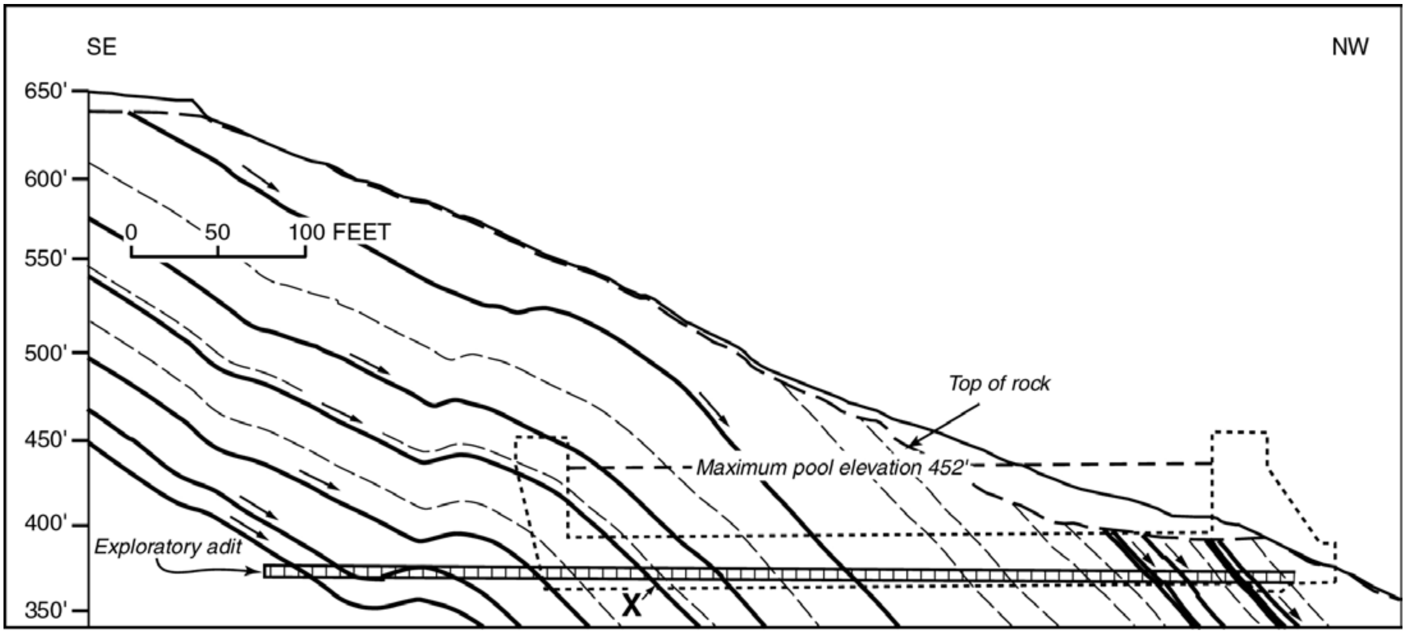



Figure 88. Cross section through the Bloomsburg Red Beds in the lower part of the northwest slope of Kittatinny Mountain in new Jersey near Tocks island (modified from Depman and Parrillo, 1969). Structure determined from exposures in 600-foot-long exploratory adit and several drill holes. Solid heavy lines are bedding-plane faults, arrows indicate direction of movement of overriding beds; short-dashed lines are normal bedding surfaces. "**X**" is bedding-slip fault along which stress measurements are discussed in text. Dotted lines show proposed spillway for the now-deauthorized Tocks Island dam.

## ROCKFALL HAZARD AT LEHIGH GAP

Joints are a natural consequence of folding of rocks. Joints control the directions that rocks break and they may facilitate landsliding. These fractures generally form in distinct sets, especially in hard competent rocks such as sandstones and conglomerates in the Shawangunk Formation which holds up Kittatinny and Blue Mountains in eastern Pennsylvania and New Jersey.

Figure 89 shows the general orientation of longitudinal and cross joints, that are the most prominent to develop because of folding. In eastern Pennsylvania and New Jersey the longitudinal joints strike (trend) northeast and the cross joints are approximately perpendicular to them, cutting through the mountain at right angles. The cross joints are planes of weakness which are sought out by streams to carve their valleys. Water gaps form in localities of abundant fractures. Folds in rocks are produced by compression due to the force of moving plates of the earth's crust. The longitudinal joints which form at right angles to the direction of compressive stress are generally smooth and planar. Cross joints, formed by pulling apart of the rock under tension, may be irregular in shape.

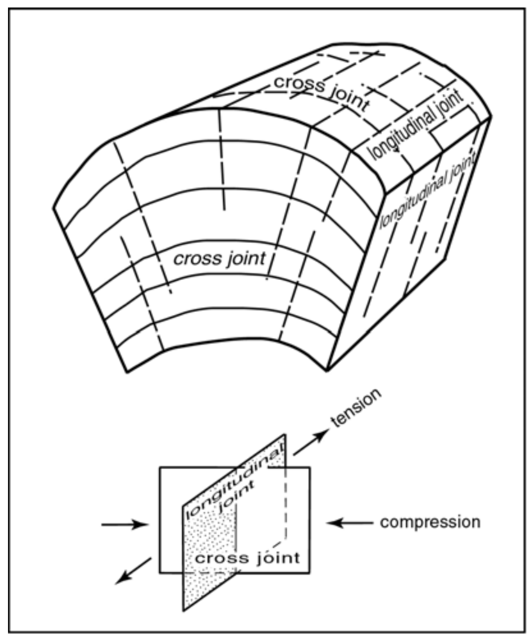



Figure 89. Diagram showing orientation of longitudinal and cross joints in folded rocks typical of the Appalachian Valley and Ridge province. Longitudinal joints form at right angles to the direction of maximum compressive stress and are generally smooth, whereas cross joints are pulled apart by tension and are more irregular.

The confining pressure against cross joints may be lessened by rapid erosion of rock along streams or by the excavation of rock by man, such as during highway construction. This may cause rock masses to move outward and become a rockfall hazard, such as that described at Delaware Water Gap. Cross joints are ubiquitous throughout the Appalachian Mountains. As excellent example of a rockfall hazard is along PA 248 in Lehigh Gap, 29 miles southwest of Delaware Water Gap (Figure 90).

The Lehigh River cuts through Blue Mountain just south of Palmerton, PA. The geology at Lehigh Gap has been described by Epstein and Epstein (1967) and Epstein et al. (1974). Here, northwest moderately dipping shales and graywacke of the Martinsburg Formation, partly with well-developed slaty cleavage, is overlain by resistant sandstone and conglomerate of the Shawangunk Formation (Figure 90*A*). PA 248 is cantilevered between two railroads as it enters the gap (Figure 90*A*), one to the west near the Lehigh River, and the other on the slope 90 feet above, abandoned a few years after the highway was completed in 1960. Falling rocks onto and erosion along the upper railroad grade are a recurring problem, although not as serious as the potential of rockfalls initiated along cross joints.

The rocks in Lehigh Gap have been denuded of vegetation by previous smelting operations at New Jersey Zinc Co. plants in Palmerton. The abundant vegetation in the gap during pre-smelting days can be seen in Plate 11B of Miller, et al. (1941). The Martinsburg formation contains slaty cleavage except within a couple of hundred feet of the Shawangunk contact (Epstein and Epstein, 1967, 1969). The cleavage, where present, breaks the rock into small fragments leading to spalling from the steep face above the railroad grade. To prevent this material from falling on the road below, a 350-foot-long wire-mesh gabion, beginning 240 feet south of the Martinsburg-Shawangunk contact and terminating at the south end of the Martinsburg outcrop where bedrock meets coarse colluvium, was constructed along the edge of the railroad grade (Figure 90*A*).

The cross joints are irregular to roughly planar (Figure 90*B*, *C*, and *D*). A diagram showing the trend of joints in sandstone and conglomerate of the lower Shawangunk Formation is shown in Figure 91. Longitudinal joints parallel the trend of the mountain, averaging about N68$^{o}$E, and the cross joints trend about N20$^{o}$W. The abandoned railroad and the highway below parallel the cross joints. The joints break the Shawangunk rocks into blocks as much as 10 feet long, each weighing many tons. Outward movement from these fractures were noted in 1989, and, because of the potential for these rocks falling on the highway below and because the site is adjacent to the Appalachian National Scenic Trail, the National Park Service requested the Federal Highway Administration to analyze mitigation procedures. A geotechnical evaluation was also prepared by G.F. Wieczorek and R.W. Jibson of the U.S. Geological Survey at that time (G.F. Wieczorek, written communication, 1989). It was determined that there was potential for rapid failure of large blocks of sandstone and conglomerate resulting in their cascading down to the road below. The retaining wall along the road below could be destroyed resulting in severe damage to the roadbed and risk to any vehicles present at the time. The potential slope failure in the Lehigh Gap area was mitigated following the 1990 evaluation by removal of large blocks of rocks from the fractured Shawangunk rocks and by construction of a gabion along the outcrop of the Martinsburg Formation to prevent erosion and rock spalling.

The cross fractures recorded at Lehigh Gap are exactly similar to those in the larger surrounding area (Epstein et al., 1974, p. 271). Figure 92 is a radar image of the region in which many lineaments define the cross-fractures. Many streams, gullies on mountain fronts, and sections of the Lehigh River, including that at Lehigh Gap, are controlled by these cross fractures. An appreciation of these structures and their orientation in relation to roads and other constructions is important to avoid potential slope instability problems.

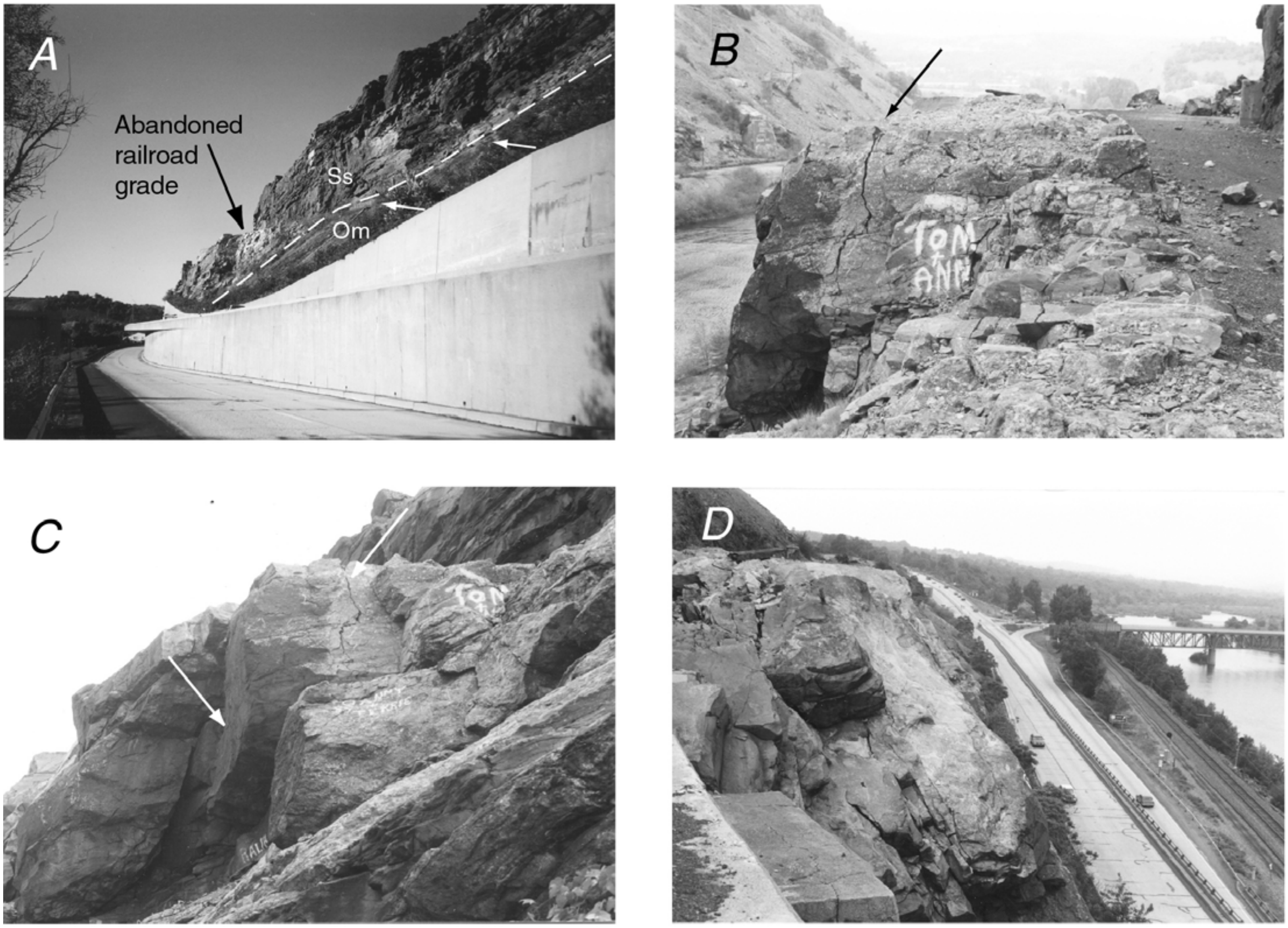



Figure 90. Potential rockfall due to cross fractures along PA 248 and abandoned railroad grade in Lehigh Gap.
*A.* Northbound lane of the highway, as it appeared in 2000, is cantilevered above the southbound lane beneath the contact between the Shawangunk Formation (Ss) and Martinsburg Formation (Om). The location of the highway was constrained by a railroad along the Lehigh River to the left and a railroad, now abandoned, above. A wire-meshed gabion (arrows) lines the edge of the railroad grade to protect against falling rocks and erosion.
*B.* Cross joints in the Shawangunk Formation (arrow) opening parallel to the abandoned railroad grade; Lehigh River below. Compare with the cross joints in Figure 78E.
*C.* View of cross joints from highway below. Some of the rock has been removed subsequent to taking this picture in 1990.
*D.* Cross joints in the Shawangunk Formation parallel the highway below.

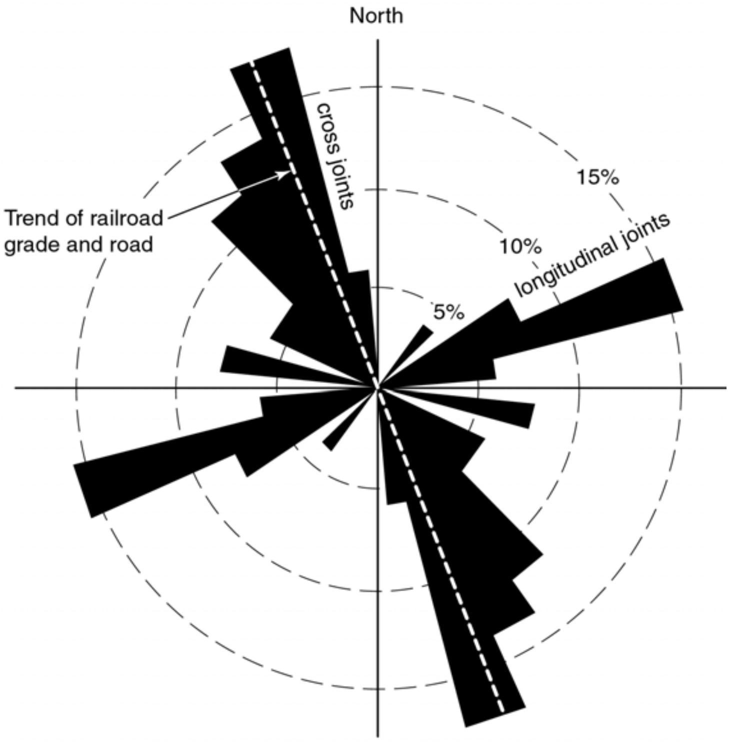



Figure 91. Rose diagram showing strike of 52 joints at Lehigh Gap, 29 miles southwest of Delaware Water Gap. Dashed line shows trend of the abandoned railroad grade above PA 248.

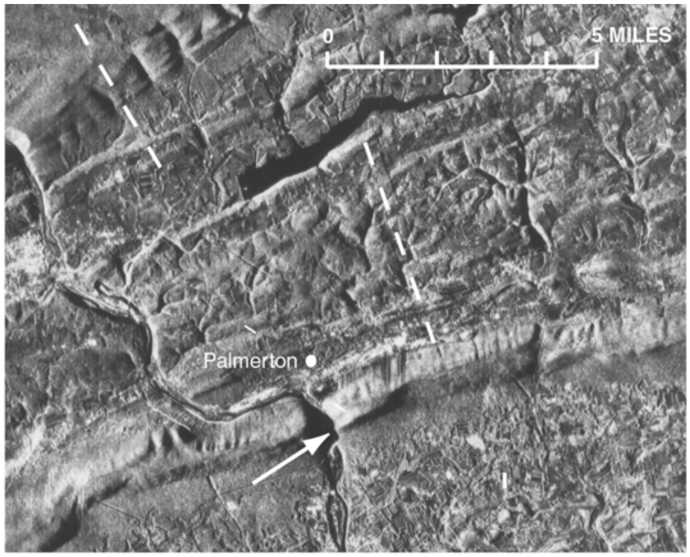



Figure 92. Radar image of the Lehigh Gap area showing location of the potential rockfall along PA 248 south of Palmerton, PA (arrow) and fracture control of many of the NNW-trending lineaments (dashed line). From Newark, NJ; PA; NY radar mosaic, U.S. Geological Survey, 1984.

## CONCLUSION

Bedrock instability is favored by a variety of fractures in bedrock, such as joints, bedding, and cleavage. The relation of slope to the orientation of these fractures, as well as the type of geologic material, is important in determining the potential for failure. Ground instability within the Delaware Water Gap National Recreation Area, and at Lehigh Gap near the Appalachian Trail, is influenced by slope declivity, shallow soils with tree roots that do not penetrate into underlying bedrock, polished glaciated bedrock surfaces beneath thin regolith, till on steep slopes, and relation of joint and cleavage orientation to roads. Appreciation of these factors and an adequate geologic database would be helpful in avoiding or mitigating slope-failure hazards.

# THE FALLS ON DINGMANS CREEK, PIKE COUNTY, NORTHEASTERN PENNSYLVANIA

by
William D. Sevon and Jon D. Inners

***ABSTRACT***

Four falls, Dingmans (lower), 130 ft, Deer Leap, 34 ft, Fulmer, 78 ft, and Factory (upper), 34 ft, represent major interruptions, knickpoints, in the longitudinal profile of Dingmans Creek, a Delaware River tributary. The lips and faces of all the falls have multiple notches that are many feet down and back (upstream) from former valley-wide, valley-head, vertical faces. Narrow valleys occur between the Delaware River valley and Dingmans Falls and between Dingmans and Deer Leap Falls. These valleys are incised into bedrock, have bedrock floors thinly veneered with gravel, have several low (<6 ft high) falls along their length, and are 1.5 and 1.2 mi. long, respectively.

The Deer Leap-Fulmer-Factory Falls complex is in a short (1,000 ft long), narrow, steep-sided valley with a bedrock floor. Dingmans Falls and the valley below are on siltstone (Upper Devonian, Mahantango Formation) with thin-spaced bedding and wide-spaced, sub-vertical fractures. The other falls are on sandstone and siltstone (Upper Devonian, Millrift Member, Trimmers Rock Formation) that have close- to moderate-spaced bedding partings, local fracture cleavage, and moderate-spaced, near-vertical fractures.

The following interpretations are made. Knickpoint erosion is by quarrying or plucking. The Mahantango Formation is more difficult to erode than the Millrift Member. Erosion of falls and valleys by glacial ice was minimal because Dingmans Creek is normal to ice-flow direction in this glaciated area. The partly eroded, former valley-head, falls faces represent pre-Pleistocene knickpoint positions. Notching and recession of falls lips and faces were multi-phase erosion events caused by meltwater flow during 3 or 4 deglaciations. Holocene erosion of the falls is minimal. Similar sequential events occur at Pinchot, Raymondskill, and Bushkill Falls, all falls in the same bedrock strike belt.

## INTRODUCTION

The falls on Dingmans Creek, Pike County, northeastern Pennsylvania (Figure 93), represent a diversity of falls types at two different locations. The lower two falls, Dingmans and Silverthread, occur at the end of an access road entered from US 209 just south of PA 739. The upper three falls, Deer Leap, Fulmer, and Factory, occur at Childs Park, about 1.2 miles upstream from Dingmans Falls. Childs Park is accessed at 1.7 mi. along a road that turns off PA 739, 1.2 mi. west of US 209.

These falls, except for Silverthread, are knickpoints on Dingmans Creek and a correct interpretation of their history is important in evaluating the history of landform development in Pike County. Silverthread Falls, on a stream tributary to Dingmans Creek is a somewhat different falls, but may have a similar history.

The rocks will be described first and then the falls and their history. Discussion of the falls starts with those at Childs Park because it is there that the erosional model was developed and is best displayed. Dingmans and Silverthread Falls are discussed next. Geology of the area has been mapped and described by Fletcher and Woodrow (1970) and Sevon et al. (1989). Some elements of this paper were presented previously (Sevon and Inners, 2001).

Sevon, W. D. and J. D. Inners, 2001, The falls on Dingman's Creek, Pike County, northeastern 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. 136 - 145.

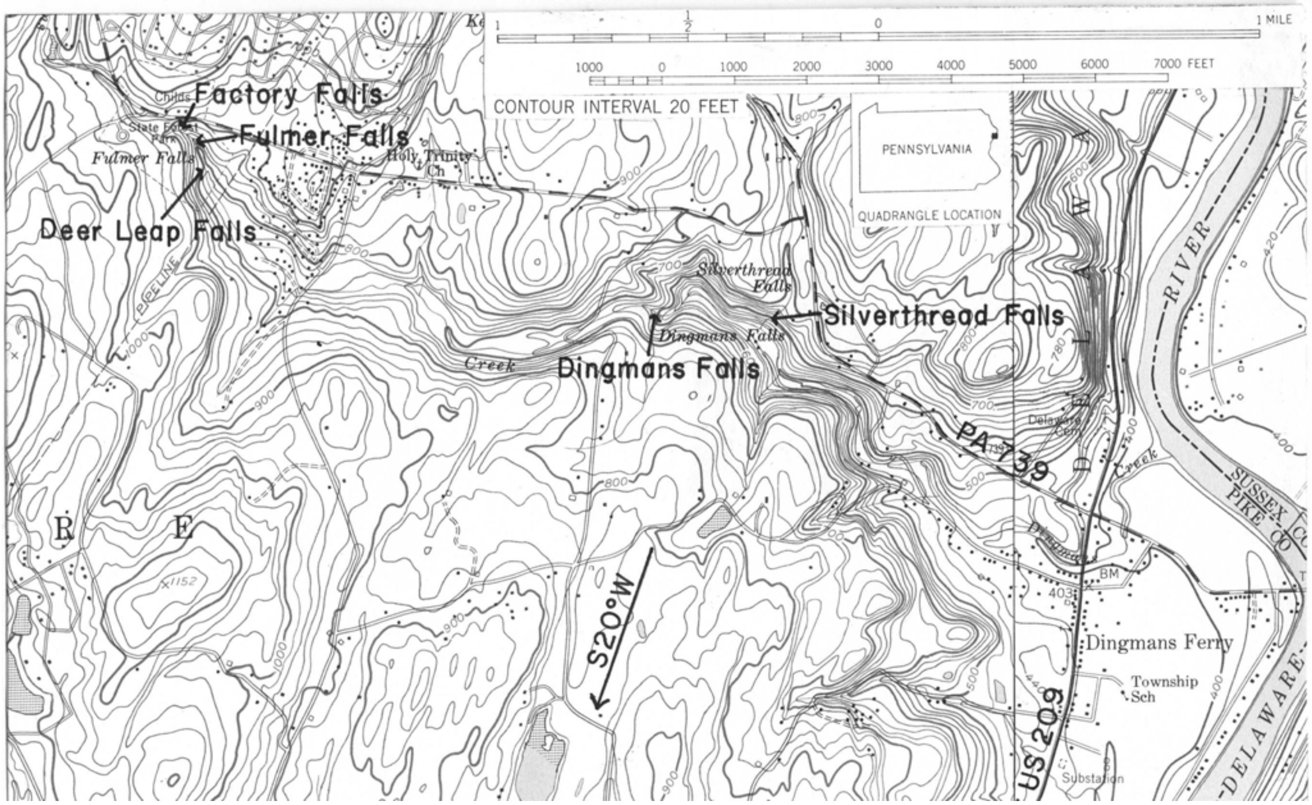



Figure 93. Location map for the falls on Dingmans Creek, Pike County, Pennsylvania. Arrow oriented S20°E is general direction of Ice flow in this area

## THE ROCKS

The rocks exposed at Childs Park are all part of the Millrift Member of the Trimmers Rock Formation (Fletcher and Woodrow, 1970; Sevon et al., 1989). Some covered intervals occur, but between the base of Deer Leap Falls and the top of Factory Falls 150-200 ft of rock are exposed. This rock is predominantly very fine-grained sandstone or siltstone. In general, the Millrift consists of about 60 percent sandstone and 40 percent siltstone and shale (Fletcher and Woodrow, 1970). The rocks are divisible into two general categories represented in outcrops readily visible at the top of Deer Leap Falls and adjacent to Fulmer Falls. These outcrops are beautifully etched by weathering and clearly display the bedding characteristics of the rock.

Beds at Deer Leap Falls are oriented N10°E/9°W. Fractures have the following orientations: N30°W/90° and N62°E/88°S. Spaced cleavage orientation is N65°E/82°S with 1-6 in. spacing, but spacing is mostly 2-3 in. (Figure 94A). Bedding at Fulmer Falls is variable but generally around N30°E/8°W. Fracture cleavage there is N82°E/80°E. Other similar bedding and fracture orientations occur depending on the bed selected for measurement.

At Deer Leap Falls the rock is mostly thinly laminated with individual lamina generally 1-5 mm thick (Figure 94B). The laminae occur in packages that are a few to several inches thick. Each package is separated by a distinct, bedding-plane-parallel break, but this break does not seem related to a change in lithology. At a quick glance the laminae within each package appear quite uniform and parallel. Close examination shows that there is cut and fill in places, but that the fill laminae are laterally continuous for 10's of inches or more. Some beds appear more massive and thick, but frequently these beds are equally finely laminated, just not etched to the same extent by weathering. The reason for the dissimilarity in degree of etch-weathering has not been determined, but may be related to grain size.

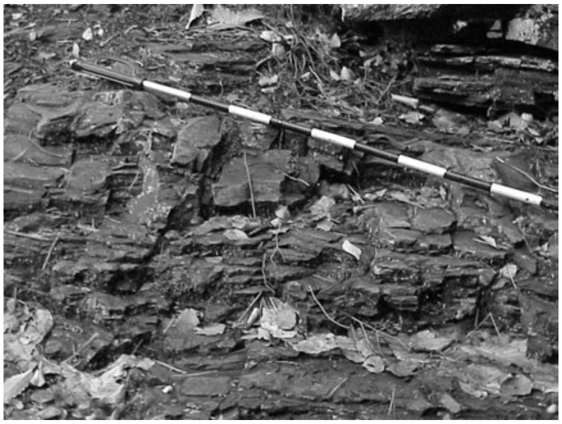


Figure 94A. Bedding and spaced cleavage in Millrift Member, Trimmers Rock Formation at lip of Deer Leap Falls. Scale intervals = 4 in.

Figure 94B. Bedding in Millrift Member, Trimmers Rock Formation at lip of Deer Leap Falls.

Figure 94C. Slump structures in Millrift Member, Trimmers Rock Formation at Factory Falls. Scale intervals = 4 in.

Figure 94D. Bedding and ripple structure cross sections in Millrift Member, Trimmers Rock Formation near lip of Fulmer Falls. Scale intervals = 4 in

Figure 94E. Interference ripples in Millrift Member, Trimmers Rock Formation at base of Factory Falls Scale intervals at bottom = 1 in.

Figure 94F. Bedding and hummocky cross-stratification in the Millrift Member, Trimmers Rock Formation near the base of Factory Falls. Scale intervals = 4 in.

Good etched surfaces for examination occur adjacent to the trail at the top of Deer Leap Falls.

Along the left-bank trail upstream from the bridge above Deer Leap Falls is outcrop that shows elongate, rounded masses that are either load casts or slump structures. We favor the latter. Similar but larger masses occur beneath an overhang in left-bank outcrop at Factory Falls (Figure 94C).

A beautifully etch-weathered outcrop occurs on the left bank of Dingmans Creek at the overlook for Fulmer Falls. Here are packages of sediment that are quite dissimilar to the packages at Deer Leap Falls. Each package here has a sharp base, well-defined cross-bedded or planar-bedded lamina of fine-grained sandstone, and a thin, finer-grained bed at the top (Figure 94D). These packages are essentially couplets that are generally 2-4 in. thick. The finer-grained bed weathers and erodes to create a small recessed area beneath the overlying package. The cross beds are presumably related to interference wave ripples similar to those seen on bedding surfaces above Fulmer Falls (Figure 94E).

Several moderate-sized bedding surfaces occur along the left bank between Fulmer Falls and Factory Falls upstream from the small pavilion. Two of these surfaces display hummocky cross stratification (Figure 94F) and one displays interference wave ripples (Figure 94E). Additional aspects of bedding and sedimentary structures occur in the rocks along the left bank at Factory Falls.

The texture and sedimentary structures in the Millrift Member of the Trimmers Rock Formation at Childs Park, particularly their stratigraphically upward changes, indicate a facies that represents deposition in shallowing water and an approach to the shoreline. This is what would be expected here where the rocks at the top of Factory Falls are less than 100 ft below the base of the Towamensing Member of the Catskill Formation, a non-marine rock sequence.

Between Deer Leap and Dingmans Falls is a valley about 1.2 mi long. The upper half of the valley is cut into rocks of the Sloat Brook Member, Trimmers Rock Formation. The Sloat Brook Member consists of about 90 percent siltstone and shale with the remainder being sandstone (Fletcher and Woodrow, 1970). The rocks are softer and more easily eroded than those of the Millrift Member. Thus, the valley shows the effects of glacial erosion and is broader and less deep than that cut by fluvial erosion into the Millrift Member above or the Mahantango Member below. The lower half of the valley is cut into siltstone of the Mahantango Formation.

Both Dingmans and Silverthread Falls are formed on siltstone of the Mahantango Formation. The siltstone is uniformly thin bedded with bedding planes generally less than an inch apart. However, bedding partings are not always well defined and some of the Mahantango appears somewhat massive. In addition, bedding partings are often irregular in shape and not sharp, continuous planes. Bedding at Dingmans Falls is oriented N31$^{o}$/8$^{o}$N and has widely-spaced joints oriented N69$^{o}$E/80$^{o}$N and N4$^{o}$E/86$^{o}$E. Bedding at Silverthread Falls is oriented N18$^{o}$E/10$^{o}$N and has joints oriented N7$^{o}$E/90$^{o}$, N60$^{o}$E/90$^{o}$, and N44$^{o}$W/79$^{o}$E. It is interesting that the bedding at these falls displays no cleavage or close-spaced joints. This is a contrast to outcrop in shale-chip rubble quarries along US 209 that show well-developed cleavage generally oriented about N67$^{o}$E/66$^{o}$S. The development of cleavage in the Mahantango either disappears rapidly to the west away from the Delaware River because of tectonics or its development along the Delaware River is related to pressure-release phenomena (Ferguson, 1967, 1974; Wyrick and Borchers, 1981).

A narrow valley occurs between the Delaware River valley and Dingmans Falls. This valley is incised into Mahantango bedrock, has a bedrock floor thinly veneered with gravel, has at least one low (<6 ft high) fall along its length, and is 1.5 mi. long.

## THE FALLS

Three falls at Childs Park, Deer Leap (lower), 34 ft high (Figure 95), Fulmer, 78 ft high (Figure 96), and Factory (upper), 34 ft high (Figure 97) represent major interruptions, knickpoints, in the longitudinal profile of Dingmans Creek, a Delaware River tributary. The falls complex is in a short (1,000 ft long), narrow, steep-sided valley with a bedrock floor. It has several small, knickpoint falls (Figure 98) between the larger falls. The lips and faces of all the large falls are notched many feet down into former higher and wider lips and back (upstream) from former valley-wide, valley-head, vertical faces. For example, Deer Leap Falls is first notched back 14 ft from the original valley-head, valley-wide, falls-face, then 7 ft back in a narrower notch, and finally 21.6 ft back in a 10 ft deep and 6.5 ft wide final notch (Figure 95). The following interpretations are made. Knickpoint erosion is by quarrying or plucking of pieces of rock whose size is controlled by bedding-plane partings, fracture cleavage, and joints. Erosion of falls and valleys by glacial ice was minimal because Dingmans Creek is normal to ice-flow direction in this glaciated area. The partly eroded, former valley-head, valley-wide falls faces represent pre-Pleistocene knickpoint positions. These valley-head falls faces were the result of long-continued and uninterrupted erosion. Notching and recession of falls lips and faces were multi-phase erosion events caused by meltwater flow during 3 or 4 deglaciations. Holocene erosion of the falls is minimal.

Figure 95. Deer Leap Falls, Childs Park, Pike County, northeastern Pennsylvania. The three stages of notching and down-cutting are shown well by the ledges on the left side of the falls. The large vertical face on the right, front side of the falls correlates with the front face on the left side.

Figure 96. Fulmer Falls, Childs Park, Pike County, northeastern Pennsylvania. Notching and down-cutting at Fulmer Falls is a little less obvious than at Deer Leap Falls, but follows the same pattern. The eroded front face on the right is the original valley-wide face. The third erosional episode cut back to the present falls position and also widened the falls face so that it is wider than the second episode notch through which the water flows. This can be seen somewhat in this photograph by observing the path of the flowing water.

Figure 97. Factory Falls, Childs Park, Pike County, northeastern Pennsylvania. Factory Falls looks different than the two lower falls in Childs Park, but it is interpreted to have the same erosional history. Here, however, the erosion is shown less by the notching and more by the flat expanses of bedrock that occur at different levels. The front rock face on the right side of the lower falls is the original valley-wide face.

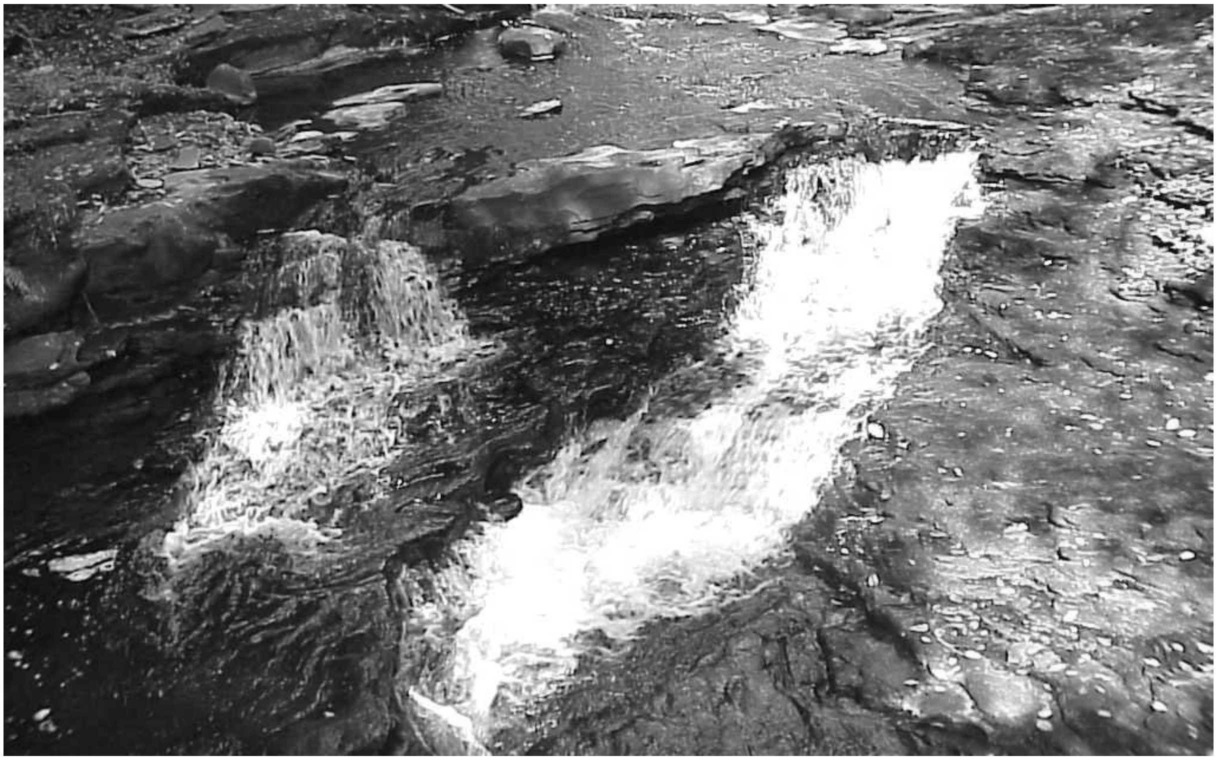
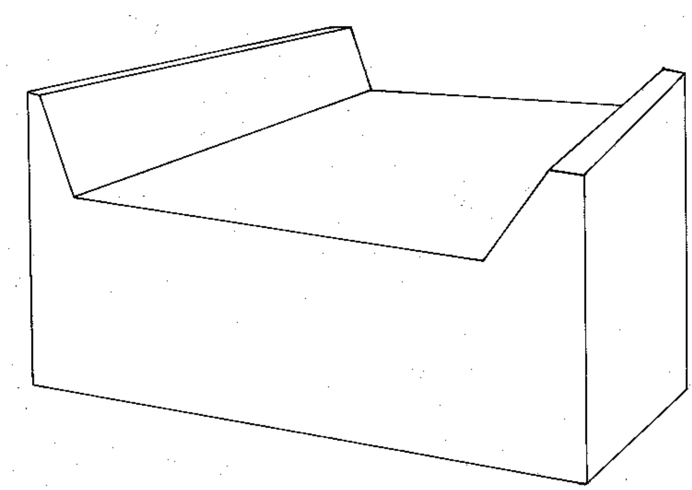


Figure 98. Low, unnamed falls in Childs Park, Pike County, northeastern Pennsylvania. This falls is typical of the several small knickpoints that occur along Dingmans Creek. It occurs just upstream from a bridge that crosses Dingmans Creek about half way between Deer Leap and Fulmer Falls.

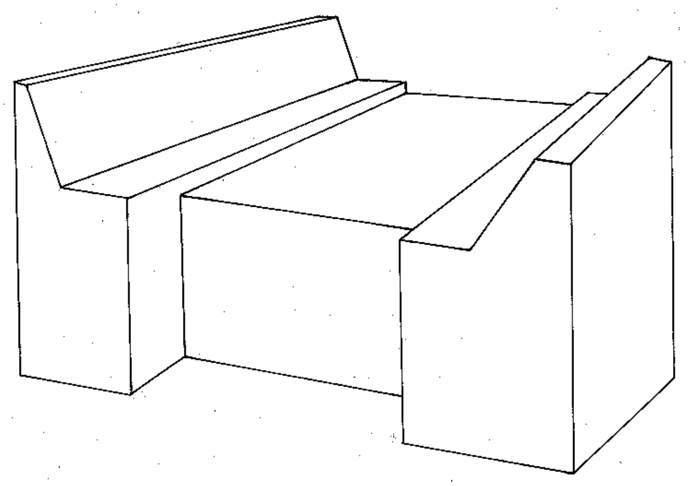
The model for erosion of the falls is discussed in four stages with each stage illustrated by a drawing.

Stage 1



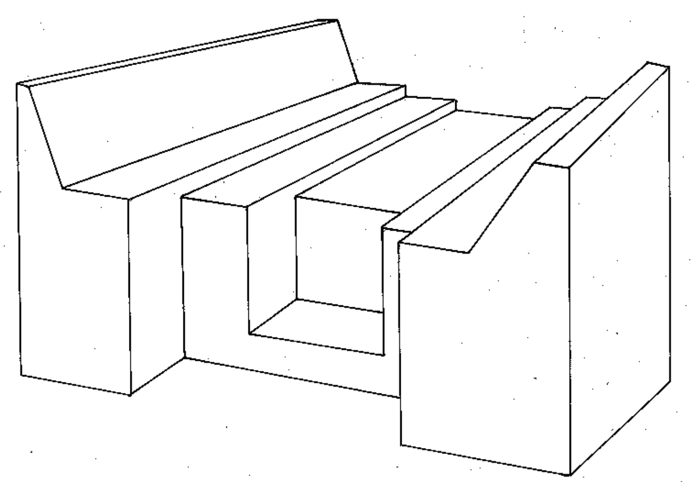
At the start of the scenario discussed here, Stage 1, Deer Leap, Fulmer, and Factory, had a valley-wide, vertical, falls face. We suggest that this was the situation at the start of the Pleistocene and that prior erosion and migration of the knickpoints represented by the falls occurred during the late Tertiary.

Stage 2



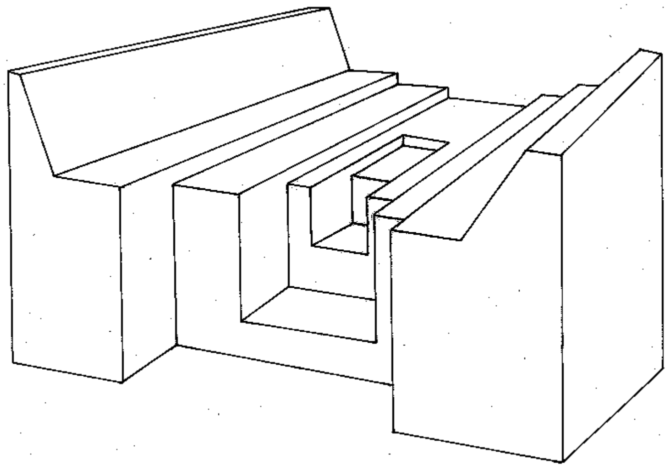
During deglaciation following pre-Illinoian ice advance, meltwater caused accelerated erosion of bedrock at the falls. This erosion was by quarrying and plucking of bedrock. The result of the erosion was notching of the falls face and lowering of the streambed. Present form at each of the falls indicates that this was the largest erosion event, a fact possibly correlative with the size of the pre-Illinoian glacier, the largest known glacier in northeastern Pennsylvania.

Stage 3



During deglaciation following Illinoian ice advance, meltwater caused accelerated erosion of bedrock at the falls. This erosion was accomplished in the same manner as that during Stage 2. However, the amount of erosion was less than that during Stage 2, possibly because the Illinoian glacier was smaller than the pre-Illinoian glacier. Even though the whole area of the falls was covered by ice as in Stage 2, the thickness and maximum extent of the ice would have been less and the volume and the length of time of meltwater flow would presumably have been less than in Stage 2.

Stage 4



During deglaciation following Late Wisconsinan ice advance, meltwater caused accelerated erosion of bedrock at the falls. This erosion was accomplished in the same manner as that during Stages 2 and 3. However, the amount of erosion was the least of the three stages, presumably because the Late Wisconsinan glacier was the smallest of the three glaciations. A fourth glaciation in northeastern Pennsylvania similar to that in northwestern Pennsylvania is a possibility, but has not been established and any possible effects on these falls has not been noted.

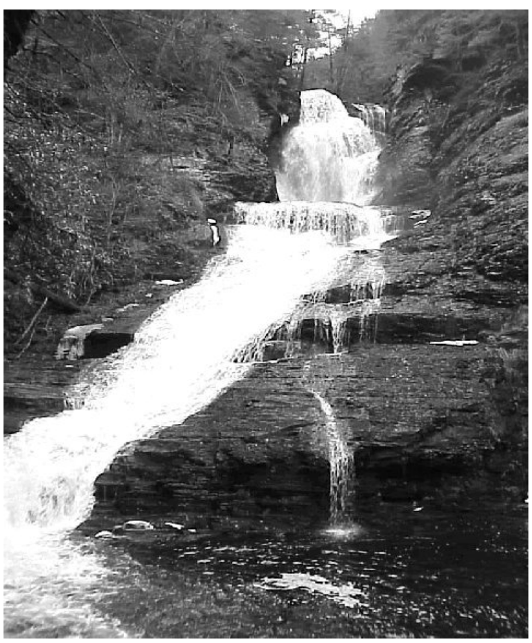


Figure 99. Dingmans Falls, Pike County, northeastern Pennsylvania. Dingmans Falls is eroded on siltstones of the Mahantango Formation and has a buttress shape not present in the falls at Childs Park. Notching does occur in the upper part of the falls and it is assumed that the stages of development for this falls are the same as for those at Childs Park.

Dingmans Falls (Figure 99) appears to have a similar, but more complicated history. First of all, at present Dingmans Falls and Dingmans Creek above the falls are oriented normal to the valley immediately below the falls. Interpretation of the topography (Figure 93) in the falls area indicates that, when Dingmans Creek was flowing at a level more than 100 ft above the present level, it turned northeast about 600-700 ft upstream from the present falls, traveled about 400 ft to the northeast, and then flowed straight into the lower valley. The position of Dingmans Creek may have been relocated during the earliest glaciation or it may have been relocated by tributary piracy. We suspect the former. The present valley extends southwest beyond the base of Dingmans Falls and may be a remnant of a tributary valley. Glacial erosion of a valley parallel to ice flow (Figure 93) would have been considerable as is evidenced by the great depth of erosion that has occurred in the Delaware River valley. This erosion would have created a new channel into which Dingmans Creek could easily have been diverted. The upper part of Dingmans Falls exhibits the same sort of notching history that is present at the Childs Park falls and that suggests that the diversion was related to the earliest glaciation.

Dingmans Falls looks different than the falls at Childs Park because the lower part is a buttress falls. Notching occurs at two levels below the upper lip of the falls. There is a much higher notched level above the main lip of the falls. There is also a low falls within a short distance upstream from the main falls. We have not studied Dingmans Falls adequately to give a detailed description of its history. It appears that the model developed for the falls at Childs Park explains the history of Dingmans Falls following its position relocation, but more work remains for verification.

Silverthread Falls (Figure 100) is possibly the most scenic of the falls in the area because it presents the appearance of a narrow spread of water falling precipitously down vertical rock faces. The unnamed tributary to Dingmans Creek has eroded the gently north dipping Mahantango siltstone headward between very well-developed N7$^o$E/90$^o$ joints. The close spacing of these joints at the falls may indicate the presence of a fracture trace. There are a series of steps in the falls and it is probable that they relate to the notching episodes modeled for the falls on Dingmans Creek. This falls has not been investigated in any detail.

Figure 100. Silverthread Falls, Pike County, northeastern Pennsylvania. Silverthread Falls differs from the other falls discussed here in that it is on a stream tributary to Dingmans Creek and that stream has eroded a channel whose shape and location is determined by joints that are oriented N7$^o$E/90$^o$. The falls have several set-back steps that may correlate with the notching events at the falls on Dingmans Creek.

As was discussed earlier, the valley below Dingmans and Silverthread Falls is floored with bedrock thinly veneered with gravel. This bedrock floor disappears beneath outwash and floodplain sediment in its lower part where Dingmans Creek valley joins the Delaware River valley (Figure 93). The Delaware River valley is known to have up to 130 ft of fill in the Dingmans Ferry area, but the maximum depth to bedrock in the center of the valley is not known. It is our interpretation that the pattern and consistency of the erosional model presented here indicates a normal base-level juncture of Dingmans Creek and Delaware River in pre-Pleistocene time. The excessive deepening and subsequent fill within the Delaware River valley is concluded to be all the result of multiple Pleistocene events of glacier erosion followed by deglaciation meltwater deposition in a valley that controlled the direction of ice flow.

## SUMMARY

The falls on Dingmans Creek are eroded into rocks of the Mahantango and Trimmers Rock Formations. The falls have been eroded by plucking and quarrying. Interpretation of a consistent and similar notching pattern for all of the falls leads to an interpretation of erosion during several stages of deglaciation from an initial position developed during the Late Tertiary. Holocene erosion appears to be minimal. A similar erosional history is interpreted at Bushkill, Raymondskill, and Pinchot Falls, all falls eroded into the same bedrock.

# THE MARSHALLS CREEK MASTODON

by

Donald M. Hoff

The Leap peat bog near Marshalls Creek, Monroe County, Pennsylvania, held an almost complete American mastodon skeleton in total secrecy until it was discovered and excavated during the summer of 1968 (Hoff, 1969). The discovery site is located 750 ft (240 m) south of US 209 at a point approximately 1 mi (1.6 km) northeast of the intersection of that road with PA 402 in the village of Marshalls Creek ($41^o$ 02' 51"N /$75^o$ 06' 36"W, Bushkill quadrangle; Figure 101). This specimen is now called the Marshalls Creek mastodon after the area of entombment.

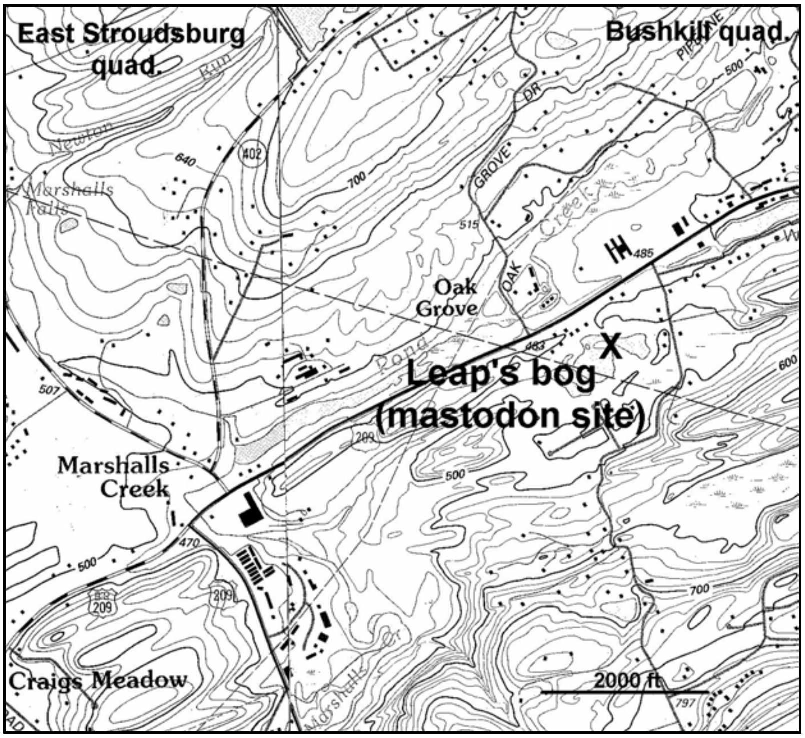



Figure 101. Location map of Leap's bog in Middle Smithfield and Smithfield Townships, Monroe County. The **X** marks the approximate site of the discovery of the Marshalls Creek mastodon in 1968.

American mastodons are well-known Pleistocene mammals. They roamed south of the North American glaciers during the Ice Age, moving north as the glaciers melted and retreated. Numerous fragmentary to complete mastodon skeletons have been found in bogs, swamps, and sinkholes in the eastern United States. The skeleton of the "Warren mastodon" found near Newburgh, New York, in 1845 is considered complete, and was described in great detail by J.C. Warren (1852) in his voluminous monograph, "Mastodon giganteus" of North America.

The American mastodon (Figure 102) weighed an earth-shaking seven tons or more, and stood approximately 8.5 to 10 ft (2.7 to 3 m) high at the shoulders. This huge beast was very fond of open forest where they browsed on sylvan vegetations. Judging from rib cage plant remains, they also ate coarse grasses, swamp plants, and mosses. Cuvier's Mastodon is often used as a generic name and also as an informal descriptive word. Blumenbach's Mammut has priority in vertebrate nomenclature. The American mastodon became extinct about 9,000 years ago and was contemporary with mammoths; Cervalces, a mooselike mammal with huge antlers; Castoroides, a giant beaver

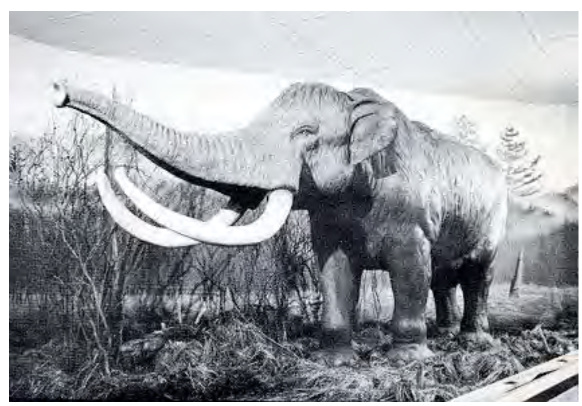


Figure 102. Marshalls Creek mastodon reconstruction in the State Museum. Photo by Robert M. Sullivan.

Hoff, D. M., 2001, The Marshalls Creek mastodon, *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. 146 - 149.

about the size of a black bear; peccaries, piglike mammals; and other interesting Pleistocene to present-day mammals (Kurten and Anderson, 1980). Sinkhole deposits often yield a great diversification of species.

Mastodons had a well-developed trunk like those of the mammoths and modern elephants. Mastodons had a flattened brow and were heavyset. Mammoths had a high-crested brow and were rangy in comparison. Mastodon means "nipple tooth" which describes the large, low cusps on the surface of their molars. Mammoth and elephant molars were/are like a millstone, with low ridges and intervening valleys on the grinding surface.

The Marshalls Creek mastodon skeleton was discovered on July 5, 1968, during a dragline peat mining operation by John Leap, owner of the Lakeside Peat Humis Company. Mr. Leap threw out his drag bucket for more peat and, during the retrieve, hooked onto what he thought was an old stump. He applied more pressure with his controls and the "old stump" snapped. Mr. Leap dumped this bucket of peat, along with fragments of the "old stump," into a waiting truck, which hauled the material to a stockpile.

Paul Strausser, an employee of the peat company, identified the "old stump" fragments as bone rather than wood. Further dragline operations were halted in the discovery area when it became obvious that there was something unusual in the bog. The bone fragments were subsequently identified as belonging to the left rear portion of an American mastodon skull.

Almost one month later, news about the Leap bog discovery reached the natural science staff of the William Penn Memorial Museum, now The State Museum of Pennsylvania. We made a preliminary investigation of the site and found it covered with about 2 ft (0.6 m) of water. In spite of the obvious great problems confronting us, we decided to excavate and established proper institutional-property owner relationships with Mr. Leap.

Figure 103. Leap’s bog and the adjacent glacial-kettle pond as it appeared in March, 2001. The actual discovery site of the mastodon skeleton was probably just off the photo to the left.

The Leap bog had been a shallow lake after retreat of the glacier. The lake subsequently became a bog through deposition of calcium carbonate-rich muck containing plant remains followed by deposition of plant material suitable for commercial purposes. Mr. Leap sold a large percentage of his peat production to golf courses in Pennsylvania and New Jersey. Prior to peat mining operations, the bog had been drained down to grow asparagus on the bog's perimeter. This provided "dry land" to move the dragline about on large, thick wood mats which afforded stability. Mining the peat reverted the area back to a shallow lake and marsh with lily pads (Figure 103).

On August 8, 1968, the natural science crew from the William Penn Memorial Museum arrived at the Leap bog to start operations. An unmined area a short distance to the northeast of the mastodon site provided a space for dragline operation and processing bone for shipment to the William Penn Memorial Museum.

The position of the skeleton was outlined by setting a large-diameter corrugated steel pipe in the muck followed by bailing and probing inside the pipe. Additional locating was accomplished from a rowboat.

A dam extending from the nearby dry area was constructed around the work site by using water resistant composition board and two by fours. Mr. Leap used his dragline to dig a sump for pumping water from the work site, and to remove many tons of muck from between the dam and skeletal area. This was a very timesaving operation which had the effect of placing the huge fossil on a pedestal.

Excavation of muck from the area indicated that an average of five to six ft (1.5 to 1.8 m) of bog deposits had covered the mastodon skeleton. The bones were embedded in soft calcium carbonate-rich marl containing plant remains under the top 2.5-ft (0.8 m) peat layer that had been removed during discovery. It was the effect of the calcium carbonate environment that had left the mastodon bones in a remarkable state of preservation.

The salvage site had now been prepared for bone removal. All bones were recovered by digging into the exposed pedestal with hand tools and removing them one by one from their matrix. The skull and lower jaw were removed first, and the remainder of the bones were excavated by troweling through the pedestal from the pelvic end. The tusks were never found, but all molar-teeth were in place and exhibited great preservation including a crust of tartar.

As excavation progressed, it became apparent that the mastodon skeleton was in a rather disarticulated condition, a condition not caused by the dragline during discovery of the remains. Some of the ribs and bones of the feet were found about 6 ft (1.8 m) from their proper position in relation to the remainder of the skeleton. This would not be the case if the animal had walked into the water and became mired in the muck to a point where escape was impossible. Imprisonment in the muck would have held the bones in place, especially the feet.

How did the life of such a great beast come to an end? The mastodon probably walked into the ancient shallow lake as a very sick animal and died. Although several of the recovered bones show effects of a bone disease, the conditions that caused the mastodon's death will never be known.

A dead animal in a pond will soon start to putrefy and bloat with gases resulting from decay. The bloated mastodon, assuming that it was not imprisoned in the muck, probably drifted around in the water with its flesh turning into a slime of putrefaction. There might also have been some action by scavengers. Some of the bones possibly dislodged from their proper positions before the remains of the mastodon slowly settled to the lake's bottom.

Every bone, including the tremendous 4-ft (1.2 m) skull, was immediately washed with clean water after removal from the bog. The skull, pelvis, and other large bones were then liberally soaked with gum acacia solution after which soft tissue paper was placed upon them with further application of gum acacia and tissue. Strips of burlap cut from old feedbags were saturated with plaster and placed on top of the gum acacia-soaked tissue. The plaster and burlap, known as "field bandages," reinforced and protected the large bones for shipment back to the William Penn Memorial Museum. The small bones were wrapped individually in paper and placed in boxes. (Gum acacia is no longer used as a preservative as it crystallizes in time. Polyethylene glycol, PEG, is presently the desirable bone preservative.)

We placed the processed bones on the ground floor of John Leap's storage barn until we could drive the museum truck to Marshall's Creek and transport them to Harrisburg. In the interim, a lightning strike set the barn on fire. It was the cool thinking and quick action by John Leap and Paul Strausser in moving the bones out of the burning barn that saved the day. The mastodon's soul had been highly angered!

Completion of the on-site salvage project on August 22, 1968, after two weeks of digging, was only the beginning of a lengthy program. Museum preparators Arlton Murray and John Schreffler completed the preliminary work on the mastodon bones in the William Penn Memorial Museum preparation area.

The field bandages were carefully removed, and every bone was painted liberally with polyvinylacetate lacquer. Fragments of the damaged bones were placed in their respective positions and joined together with steel rods and glue, while any voids were filled with beeswax. This work preceded the final mounting of the skeleton, and was completed in December of 1968.

A final survey of the salvaged mastodon skeleton indicates that it is approximately 90 percent complete. Among more than 18 mastodon finds in Pennsylvania to date, the Marshalls Creek mastodon remains the only find suitable for a mounted display. Two specimens of wood that were carefully collected from near contact with the skeleton gave radiocarbon dates of 12,160 + 180 yr B.P. and 12,020+ 180 yr B.P. (Buckley and Willis, 1970)

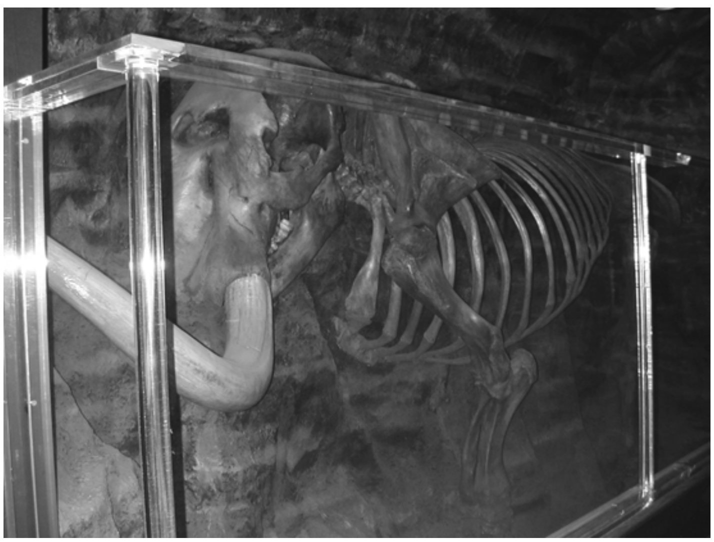


Figure 104. Mount of the Marshalls Creek mastodon skeleton at the State Museum.

Design of the present Hall of Geology (opened to the general public in August, 1976) in the William Penn Memorial Museum did not allow space for a full body mount of the Marshalls Creek mastodon. Only the left side is mounted, but includes the complete skull and mandible (Figure 104). The tusks are replicas from the Peary mastodon found near Wheaton, Illinois. A diorama opposite the mounted skeleton depicts the mastodon (Figure 102) being stalked by a pair of spear-wielding paleo-Indians. Though the extinction of mastodons and other large Ice-Age mammals in North America has long been linked to the hunting pressures of early Native Americans (see Ward, 1997), no evidence linking the demise of the Marshalls Creek mastodon to this cause has ever come to light.

# PAHAQUARRY COPPER MINE

by
Donald H. Monteverde

Pahaquarry Copper Mine lies along the western slopes of Kittatinny Mountain in the Delaware Water Gap National Recreation Area (DEWA) (Figure 105). 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 in New Jersey and possibly the United States, dating back to Dutch explorers as early as the 1650's. Researchers traced this declaration back to an 1828 publication of several letters in Hazard's Register. 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.

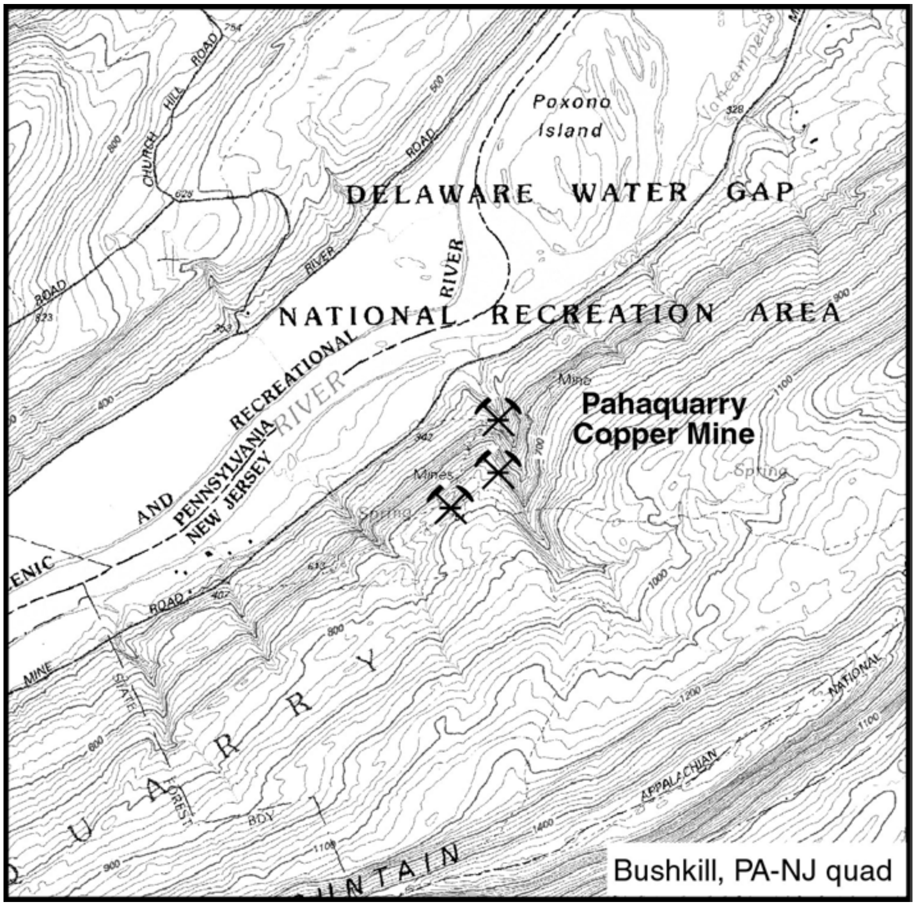



Figure 105. Pahaquarry Copper Mine location within the Delaware Water Gap National Recreation Area, on the New Jersey side of the Delaware River.

Even though Pahaquarry is not the first copper mine in New Jersey, it left a long and varied history. It underwent five separate periods of exploration and attempted exploitation, including in 1750's, 1829-30, 1847-48, 1861-62 and 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). The reader is referred to their work for all original sources.

The Bloomsburg Red Beds is the Pahaquarry mine's host rock. These Upper Silurian clastic units overlie 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 western slope of the ridge. Copper in the Bloomsburg is an uncommon occurrence. When hiking across the Bloomsburg and 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). The ore occurrence, genesis and regional geology that led to the Pahaquarry mine will be described. The description of the ore mineralization and genesis will be based entirely on Woodward (1944).

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.

## HISTORY OF MINING AT PAHAQUARRY

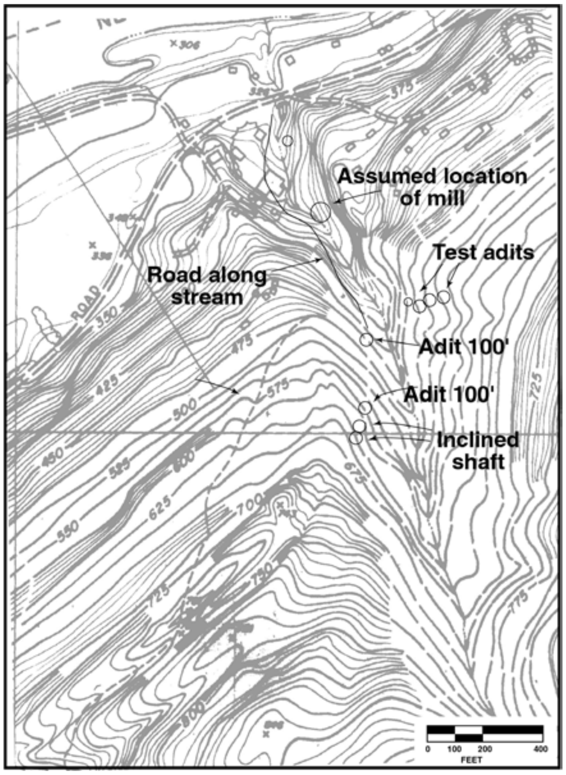



Figure 106. Map depicting Pahaquarry mine workings from the 1753-1760 historic period. Several prospects were dug on the northeast side of Mine Brook before work concentrated on the southwest side. The workings included strike parallel adits and dip parallel shafts. Modified from Burns-Chavez and Clemensen, 1995.

### *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 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 106). Southwest of the brook seemed more promising and underwent the greatest development at this time. There they excavated two adits of 50 to 100 feet each, as well as two 45$^o$ inclined shafts that attempted to follow the copper mineralization. In addition to the mineral workings several other structures including a stamp mill, several dams and water races, as well as buildings for lodging and equipment storage were built on the Delaware River terrace.

Mineralized rock excavation was labor intensive. A single jack was used to drill the rock before black powder blasting broke up the rock. A single jack involved holding a drill with one hand while hitting it with a four-pound sledgehammer in the other hand. After each hit, the drill was turned one half revolution before the next strike. Three feet depth proved to be about the maximum attainable before blasting proceeded. Black powder wrapped in paper was placed in the drill holes using 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 high-grade ore was then transported to a smelter for 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 copper ore.

### *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. Furthermore, 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.

### *1847-48*

Interest in 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 old adits, digging

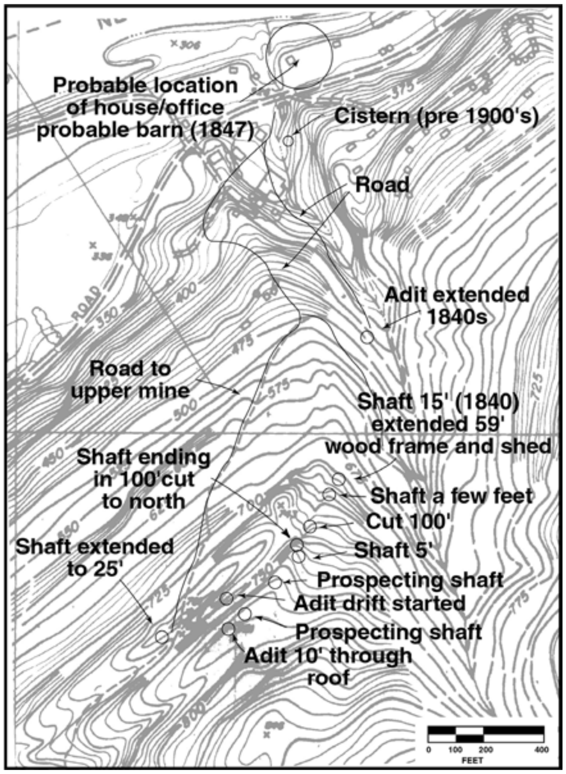



Figure 107. Map depicting Pahaquarry mine workings from the 1847-1862 historic period. Workings extended the colonial adit as well as prospected the southwest ridgeline. Modified from Burns Chavez and Clemensen, 1995.

of new adits and other prospect openings. Ten total excavations occurred including lengthening an original adit and digging four additional shallow shafts as prospects up on the higher topography southwest of Mine Brook (Figure 107). Additional diggings include two adits, one of unknown length, and a second running 10 feet into the rock before digging a 75-foot crosscut. On the higher topography, they excavated 100-foot-long by 15-foot-deep cut that led to another new opening. The last two shafts consisted of one 15 feet deep and another 20 feet deep, each with a diameter of 7 x 15 feet.

Mine development progressed by methods similar to those employed in the 1750s. Drills were hardened steel now, which increased their usage, but they were still employed as single jacks. 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 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.

### *1861-61*

The Civil War induced the resurrection of the Alleghany Mining Company in 1861 under a new directorship. An economic geologist worked on site and planed its further development. A new wood frame covered an 1847 shaft, which was extended from 20 feet to 59 feet. A second shaft was deepened from 15 down to 25 feet. Some amounts of ore were collected, but again, the low-grade ore defeated this latest attempt to develop the copper mineralization.

### *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 ore. 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 one-year lease agreement with the Alleghany Mining Company. They excavated a new adit into the hillside and removed 100 tons

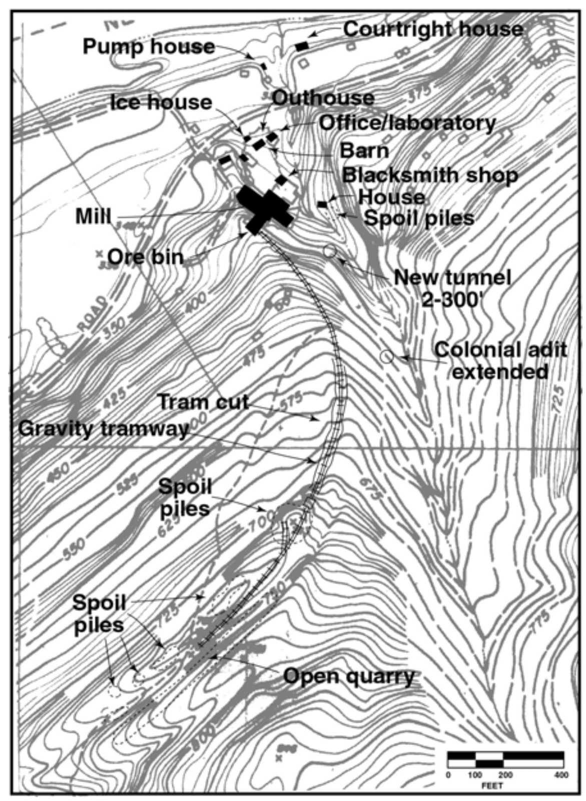



Figure 108. Map depicting Pahaquarry mine workings from the 1901-1912 historic period. Building construction, including the large mill was a major part of Pahaquarry development. Open pit mining (see Figures 110 and 111) also commenced. Modified from Burns-Chavez and Clemensen, 1995.

Figure 109. Colonial adit entrance extended during the 1901-1912 mining period. The adit entrance is gated and usually locked. Bloomsburg outcrop composed of gray quartz sandstone. Siltstone and shale occur within the more recessive layers on the face. Bloomsburg beds dip northwestward while cleavage dips moderately to the southeast. View faces southwest.

of ore (Figure 108). 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 site. So in 1902, Montgomery bought the Alleghany Mining Company lands at sheriff sale. They constructed numerous structures including powder house, blacksmith shed, barn, oil, and icehouses. By 1904, the adit reached 300 feet in length (Figure 109).

This time mining technological advances allowed easier ore extraction. Gas driven generators supplied power for mechanical drills. Advances in blasting methods also moved forward. 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.

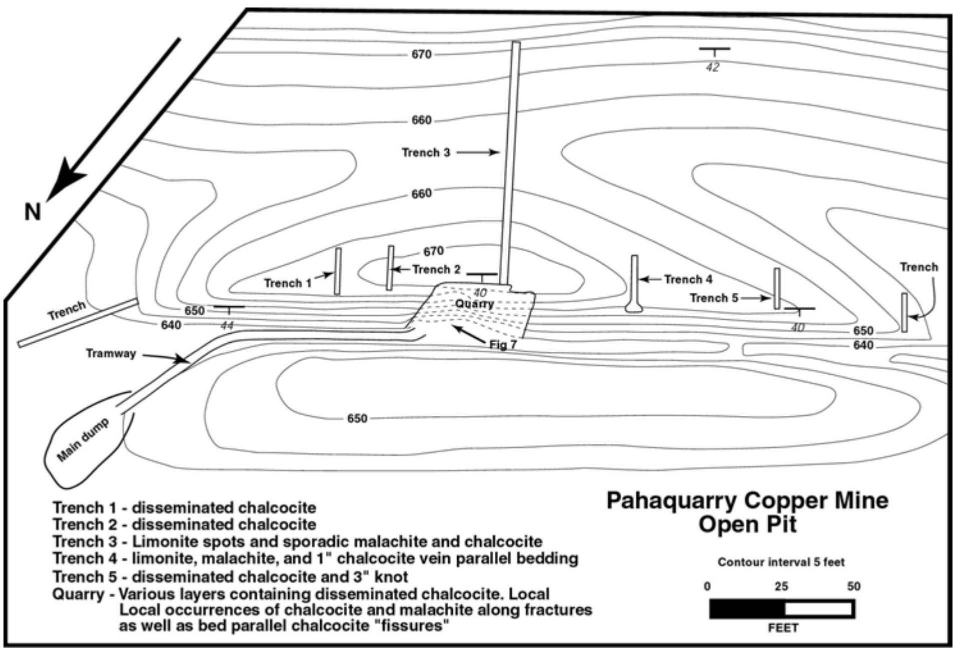



Figure 110. Map showing the open pit quarry and trench locations, southwest of Mine Brook as outlined by the USGS in 1943 for a strategic mineral assessment. The open pit work dates to 1901-1912 period while the trenches probably are part of the 1847-62 period. Note location of Figure 111 photograph. Modified from Cornwell, 1945.

Meanwhile exploration continued along the upper ridges above 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 quarry-ore removal, a tramway was constructed that brought the

Figure 111. Pahaquarry open pit photograph along ridge southwest of Mine Brook. Mostly gray quartz sandstone can be seen dipping northwestward. Occurrences of chalcocite both as disseminated grains and as several inch thick zones paralleling both fractures and bedding are visible. Moderately dipping spaced cleavage is evident in the sandstone. The layer within the cut yielding a speckled appearance is covered by current ripples that indicate a northwest-directed flow. View is east-northeast.

ore downhill 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 (Figures 110 and 111). 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 three copper ingots for a total value of $15.00, not nearly enough to maintain a mining concession. A new concentrating process was attempted which also failed. This ended copper mining at Pahaquarry in 1912.

## GEOLOGY

### Regional

Kittatinny Mountain geology is dominated by the Silurian-age Shawangunk Formation and Bloomsburg Red Beds. 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 member (Lizard Creek Member) (Epstein and Epstein, 1972). Bedding generally dips northwest, though there are many folds and faults cutting the Shawangunk. Regionally the Shawangunk averages 1,400 feet thick. These coarse clastic sediments are resistant to weathering and support the ridgeline of the Blue-Kittatinny-Shawangunk Mountains. 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 Late Silurian-age Bloomsburg conformably overlies the Shawangunk throughout the length of the Blue-Kittatinny-Shawangunk mountain chain. 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 and upward exhibit crossbedding and laminations. Siltstones overlie the sands and gradually grade upward into shale that may be mudcracked. Bloomsburg deposition occurred in a shallow to marginal marine paleoenvironment. The 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 locality exists at mile 15.4 on the alternate road log, south of the Pahaquarry trail parking lot. Dorsal and ventral plates of the ostracoderms *Vernonaspis* and *Americaspis* have reportedly been found 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 uniform rock strengths and combined thickness they reacted to the northwest-directed progressive strain of the Alleghanian orogeny in a uniform way. The weaker siltstone and shales below (Late Ordovician in age) and limestone units above (Late Silurian to Devonian in age) more readily folded and faulted than the stronger, more resistant Shawangunk-Bloomsburg rock package. This allowed Epstein et al. (1967) 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 is steep to moderately southeast dipping.

**Ore geology**

The host Bloomsburg beds dip northwest, forming a dip slope (average bedding orientation is N53$^o$E/42$^o$NW) 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 in the cuts. Sandstone beds dominate the area. Excellent glacially polished ledges of sandstone can be seen in the quarry on the upper ridge southwest of Mine Brook. There, the sandstone is gray, medium to thick bedded, crossbedded 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 ($Cu_2$ $(CO_3)(OH)_2$, copper carbonate hydroxide) and chrysocolla ($CuSiO_3$-$nH_2O$, 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 have partially replaced the host Bloomsburg bed. Thickness of the copper bearing Bloomsburg may be as much as 200 feet. The highest reported natural samples contained 3.25% copper, but the entire copper-bearing horizon has a much lower percentage.

Pahaquarry is thought to be an epigenetic deposit. The copper is believed to have been disseminated throughout the original beds as primary detrital grains. 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$^o$ C. Later, meteoric-water interaction aided the growth of the secondary minerals, malachite and chrysocolla.

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# APPENDIX A

## THE OLD MINE ROAD

This road log along the Old Mine Road in New Jersey—from Delaware Water Gap to Millbrook Village—is included because of the scenic beauty, historic significance, and geologic interest of this “ancient” transportation route. The log begins at the small parking area just south of the traffic light on the Old Mine Road, about 0.1 mile from the Millbrook Village exit of I-80, and follows the road northeast along the Delaware River a distance of 11.0 miles to Millbrook Village.

| **Miles** | | |
|---|---|---|
| Int. | Cum. | |
| | 0.0 | Parking area on left side before traffic light at entrance to Old Mine Road, north end of Delaware Water Gap. |
| 0.4 | 0.4 | One-way narrow road. Road has been rebuilt several times over the last century. Principal geologic hazards are slumping (due to undercutting of the lower slope by the Delaware River) and rock fall. Many striated Bloomsburg outcrops along route (show valley-parallel ice flow). Striae along the mountain’s crest show that ice moved southward across the mountain’s northeast-southwest topographic grain. |
| 0.6 | 1.0 | Trailhead of the Farview Hiking Trail. |
| 0.2 | 1.2 | Outcrop of Bloomsburg Red Beds, dipping northwestward, on right side of road. This road cut contains numerous examples of dorsal and ventral plates of the ostracoderms *Vernonaspis* and *Americaspis*. Because the outcrop lies within the National Park, collecting is strictly *verboten*. |
| 0.4 | 1.6 | Outcrop of Bloomsburg Red Beds on right. |
| 0.2 | 1.8 | Pass Shawnee Inn and Golf Resort across Delaware River on your left. |
| 0.5 | 2.3 | Enter Worthington State Forest. Much of this land was formerly owned by C. C. Worthington, who built the “Buckwood Inn” (now Shawnee Inn) in 1911 and had a game park here early in the last century. This is also the site of the village of Brotzmansville, which began in the late 1820’s with the construction of a gristmill by Jacob Brotzman. At its zenith, it consisted of grist and saw mills, post office, school and a few residences. The mills were built along several of the small tributaries that drained off the northwest flank of Kittatinny Mountain. |
| 1.0 | 3.3 | Douglass Parking lot, just past Worthington State Forest campgrounds on left. Campgrounds lie on a postglacial stream terrace. The long hillslope to the right is held up by the Bloomsburg Red Beds (dip slope overlain by thin till). |
| 0.8 | 4.1 | Tocks Island (downstream end). Site of the proposed Tocks Island dam (see LUNCH STOP, Day 2). Bloomsburg Red Beds crop out along the right side of the road. Numerous joints (sheeting) dip toward the road, and several rock falls have occurred along these during recent years (see Epstein, this guidebook p. 119). Locally preserved here are glacial striations that show a down-valley, late-glacial ice flow. |
| 0.6 | 4.7 | Tocks Island (upstream end). |
| 0.2 | 4.9 | Exit Worthington State Forest. |
| 0.6 | 5.5 | Pass over high-standing remnant of the Zion Church Valley train (about 100 feet above Delaware). Most of the glacial outwash in this part of the valley was eroded by meltwater and the Delaware River. |
| 0.8 | 6.3 | Dimmicks Ferry (located to the left across postglacial stream terrace). Initial ferry |

operations were probably begun in the 1820's by Moses Shoemaker. The ferry was sold to William Fisher in 1874 and purchased by Michael H. Dimmick in 1881, whose son Peter operated it until his death in 1937. The Dimmicks used two steel cables to transport the ferry across the river, an overhead cable during periods of high water, and a submerged cable during periods of low water. Ferries were a major means to cross the Delaware during the 19th and early part of the 20th centuries. Dimmicks Ferry was one of sixteen ferries to have been in operation between Delaware Water Gap and Milford, Pennsylvania.

| | | |
|---|---|---|
| 0.3 | 6.6 | Pass trailhead to Pahaquarry copper mine (see Monteverde, this guidebook, p. 150). |
| 0.3 | 6.9 | Pass Poxono Island access site on left, one of several boat launches operated by the National Park Service. |
| 1.1 | 8.0 | Pass Calno School on left, built in the 1850's. The low terraces in this area are postglacial age. Several Amerind occupation sites were located on the broad terraces west of the school. One was located just behind the school. |
| 0.3 | 8.3 | Climb onto outwash-fan terraces where Van Campens Brook enters the Delaware Valley. These terraces are as much as 60 feet higher than the postglacial terraces near Calno School. |
| 0.3 | 8.6 | Entrance to Depew Recreation site on left. |
| 0.1 | 8.7 | Cross Van Campens Brook. At this point, Old Mine Road leaves the Delaware River valley and follows Van Campens Brook to Millbrook Village, a climb of about 260 feet. |
| 0.2 | 8.9 | Outcrop of Bloomsburg Red Beds on left. |
| 0.5 | 9.4 | Striated (S56$^0$W and S78$^0$W) Bloomsburg Red Bed pavement behind farmhouse on left. |
| 1.1 | 10.5 | Entrance to Water Gate Recreation site on right. |
| 0.3 | 10.8 | Cross over Franklin Grove recessional moraine, barely a bump in the road. |
| 0.1 | 10.9 | Turn right onto Millbrook-Flatbrook Road. |
| 0.1 | 11.0 | Enter Millbrook Village. |

End of road log along the Old Mine Road.

# Appendix B1

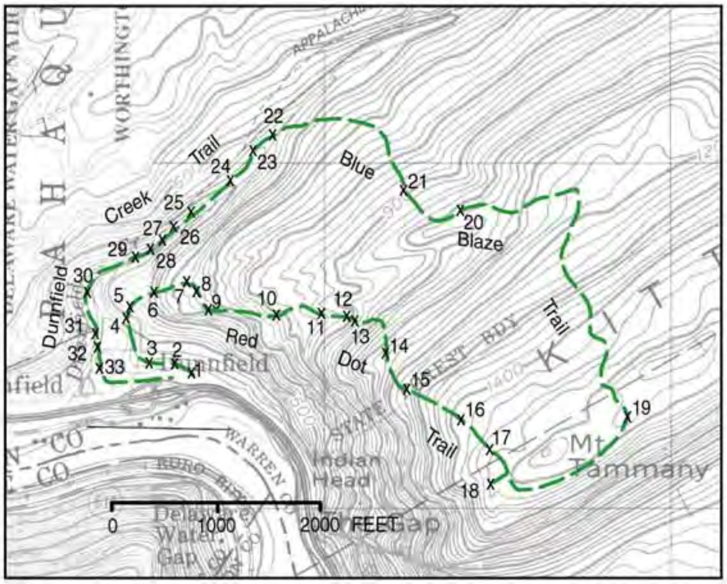
DELAWARE WATER GAP
National Recreation Area
Pennsylvania/New Jersey
National Park Service
U.S. Geological Survey

## GEOLOGIC FEATURES ALONG THE RED DOT-BLUE BLAZE-DUNNFIELD CREEK TRAILS DELAWARE WATER GAP NATIONAL RECREATION AREA

By
Jack B. Epstein

The Red Dot-Blue Blaze-Dunnfield Creek Trail circuit takes the hiker, in less than four miles, through many of the geologic phenomena, including a variety of rock types, landforms and glacial and structural features. The points of interest along the way provide insight into these natural elements that influenced the formation, history and composition of Delaware Water Gap.


The Red Dot Trail climbs to the top of Mt Tammany in New Jersey.
The sedimentary layers that make up the cliffs in the main part of the gap dip or slope to the left, but in Dunnfield Creek valley to the left the beds are horizontal.




Trail map showing stops of geologic interest

STOP 1: Near contact between the Shawangunk Formation and Bloomsburg Red Beds.
STOP 2: Eight-foot-long boulder with slickensides.
STOP 3: Glacial kame terrace on silt, sand and gravel.
STOP 4: Glacial striae.
STOP 5: Rotted limestone glacial erratic.
STOP 6: Rib of Bloomsburg bedrock.
STOP 7: Series of greenish-gray and red siltstone, sandstone and shale of the Bloomsburg.
STOP 8: Large erratic, Schoharie Formation.
STOP 9: Overlook of the Delaware Water Gap.
STOP 10: Red sandstone and siltstone of the Bloomsburg have been polished by the last glacier (20,000 years ago), producing glacial striae.
STOP 11: Springs.
STOP 12: Beginning of Shawangunk Formation on steep slope.
STOP 13: Talus.
STOP 14: Rib of quartzite with joints.
STOP 15: Glacial cobbles and glacial striae on Shawangunk.
STOP 16: Gentle slope underlain by some shale.
STOP 17: Forest fire.
STOP 18: Overlook, many sedimentary structures in the Shawangunk.
STOP 19: Blue Blaze Trail descends slope through laurel and blueberries.
STOP 20: Exposure of Bloomsburg bedrock with glacial striae.
STOP 21: Soil erosion by boots and rain exposing Bloomsburg bedrock with glacial striae.
STOP 22: Erosion has removed about three feet of glacial till.
STOP 23: Dunnfield Creek falls over flat beds of the Bloomsburg in bottom of syncline.
STOP 24: Three large erratic boulders of slightly cherty limestone.
STOP 25: Intersection with Appalachian trail. Several more boulders in creek.
STOP 26: Plunge pool formed where the creek drops over hard sandstone and gouges out a rounded pool in softer shales below.
STOP 27: The creek erodes along a joint surface here forming a 30-foot sluiceway.
STOP 28: Large boulders fallen from adjacent Bloomsburg outcrop have sharp edges compared to the rounded and eroded edges of erratics.
STOP 29: Cleavage present in horizontal shale layers but not in sandstone. Erratic in creek.
STOP 30: Beginning of the terrace deposit that was first seen at stop 3.
STOP 31: 25-foot-long limestone erratic limestone.
STOP 32: Bridge. Flat alluvial fan towards parking lot made up of rounded cobbles.
STOP 33: Parking lot. Note 6-foot boulder 40 feet above the creek in terrace to right.

GLOSSARY

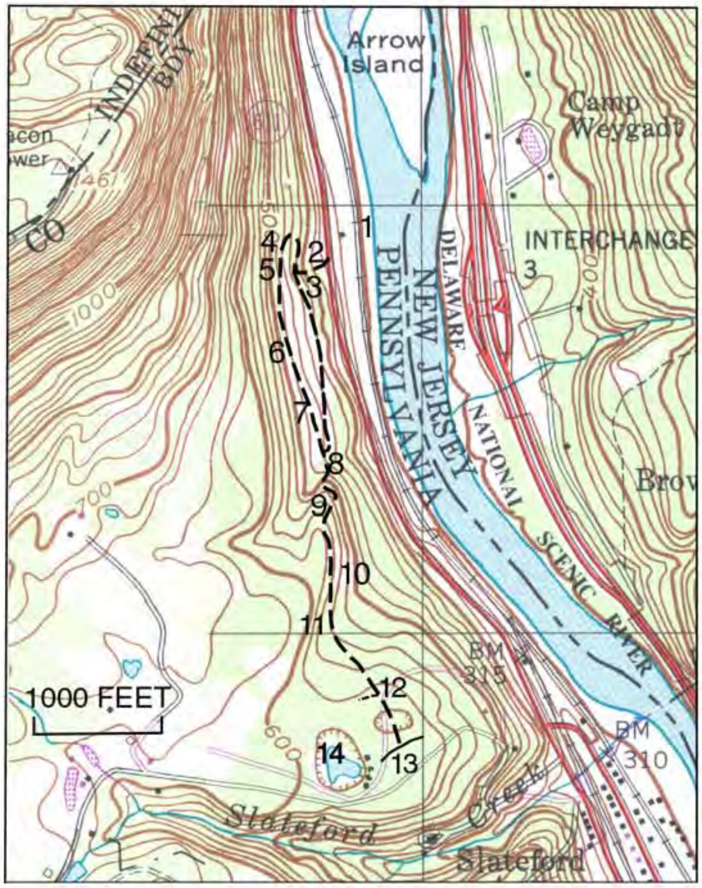
**Alluvial fan**: Gently sloping mass of sediment fanning out from a river mouth.
**Cleavag**e: Closely spaced fractures along which a rock may split.
**Erratic**: A rock that was carried some distance by a glacier from its place of origin.
**Kame terrace**: Flat-topped hill formed from sediment that was deposited along a valley wall by streams that flowed from an adjacent melting glacier.
**Slickensides**: Polished and striated rock surface caused by one rock mass sliding past another.
**Striae**: Narrow parallel scratches cut into a rock surface by rock debris embedded in the bottom of a moving glacier.
**Syncline**: U-shaped downfold of rock layers.
**Talus**: An apron of irregular rock fragments derived from and lying at the base of a cliff.
**Till**: Unsorted mixture of clay, sand, and boulders deposited beneath a glacier.

# Appendix B2

**DELAWARE WATER GAP**

**National Recreation Area**
**Pennsylvania/New Jersey**
**National Park Service**
**U.S. Geological Survey**

## GEOLOGIC FEATURES ALONG THE ARROW ISLAND TRAIL




*Trail map showing stops of geologic interest*

Stop 1: Parking lot. Delaware Water Gap; northwest-dipping beds; joints; talus; Arrow Island sand bar.
Stop 2: Round glacial erratics (red rocks; cherty siltstone) at start of trail; angular talus blocks to right.
Stop 3: Slate dump.
Stop 4: 20-foot talus boulder of the Shawangunk quartzite and conglomerate; sedimentary structures including cross bedding and channeling.
Stop 5: Slate quarry; Washington Brown quarry?
Stop 6: Waste pile of slate and small slate prospect.
Stop 7: Twenty-foot high slate pit in a ravine about 50 feet above the trail.
Stop 8: Creek near the junction of the yellow and white dot trails. Many large erratic boulders including red sandstone and an 8-foot long boulder of calcareous siltstone with some dark-gray chert. Downstream the creek has cut down through 30 feet+ of this glacial deposit.
Stop 9: Exposure of graywacke sandstone making up this topographic rib.
Stop 10: Another slate pit.
Stop 11: The bouldery nature of the glacial deposit does not make for good agricultural soil, but does supply boulders for this fence row.
Stop 12: Intersection with a cross country ski trail.
Stop 13: Parking lot and end of Arrow Island Trail.
Stop 14: Duck pond, a kettle hole.


Welcome to the Arrow Island Trail.

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

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

Stop 3. Foundation remnants below slate quarr

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

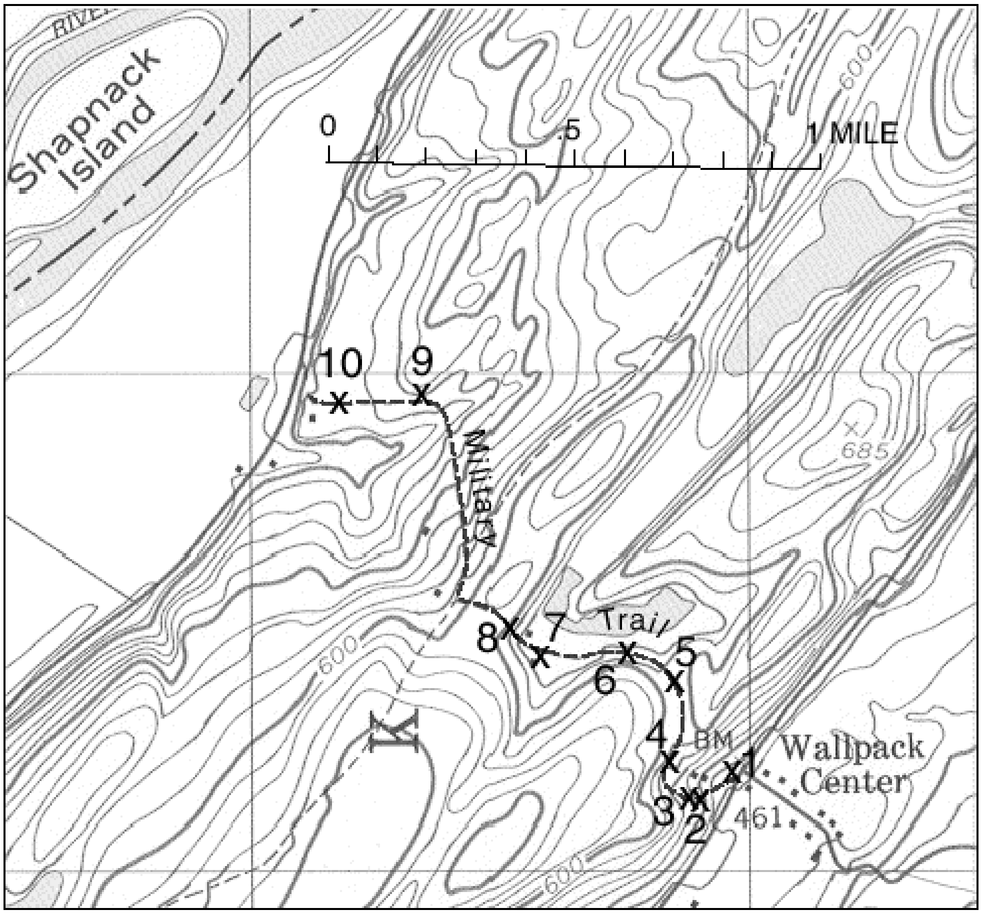
Stop 5. Original horizontal sediment layer (bedding; solid line) is now tilted. The rock breal along cleavage (dashed line).

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

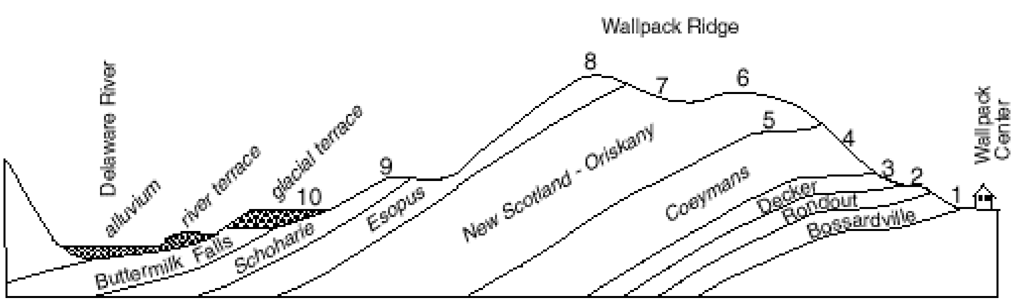
Stop 8. Rounded and polished glacial boulder north of creek.

# Appendix B3

## GEOLOGIC GUIDE TO THE MILITARY TRAIL
## WALLPACK CENTER, NEW JERSEY




Map of the Military Trail showing Stop Locations




Profile (termed a cross section by geologists) of Wallpack Ridge showing Stop Locations and Geologic Formations

## MILITARY TRAIL

The Military Trail was used in the 1750's during the French and Indian Wars to supply Fort Johns (Shapneck) at its northwestern end. The trail is a little more than one mile long. It is moderately steep during ascent and descent, rising about 150 feet to the top of Wallpack Ridge, with a rolling upper portion. Allow 2 to 3 hours for the round-trip. The stop locations are shown on the map on the front of this leaflet.

The trail head is on the west side of Wallpack- Flat Brook Road (County Route 615) across from the Wallpack Center Post Office.

STOP 1: Ten-foot-high outcrop of the Bossardville Limestone, a thin-bedded to platy limestone made up of calcium carbonate ($CaCO_3$) which fizzes when dilute hydrochloric acid is applied. The near-vertical face that parallels the trail is a joint with another joint set at right angles to it. The Bossardville Limestone has been used for agricultural lime in the past. Several kilns that were used to burn the limestone to produce lime are found elsewhere in the park. The beds in the limestone dip (tilted) to the northwest by 31 degrees. On a flat area just above the road the site of two former houses.

STOP 2: On the south side of the trail (to the left as you head uphill) and below the first sharp bend, is a pile of rocks which have a yellow weathering surface. The fresh blue-gray interior of the rock fizzes less rapidly in acid than the Bossardville Limestone. The rocks are dolomite ($MgCaCO_3$) of the Rondout Formation.

STOP 3: At the bend in the road is a one-foot-thick rib of tan sandstone of the Decker Formation. For the next 250 feet, lying below the deteriorating blacktop (isn't it nice that nature can and will reclaim the works of man?) are a wide variety of rocks (sandstones, shales, limestone, some of which are red, many foreign to this immediate area) which have been brought here by the last glacier which retreated from this area about 10,000 years ago.

STOP 4: Sticking up through the trail bed is limestone of the Coeymans Formation. Note that it's crystals are much larger that those of the Bossardville Limestone seen below. Nearly black nodules that are imbedded in this limestone are chert, or flint, which the early Indians of this region may have used for a variety of tools. Farther along the trail are more boulders of glacial till.

STOP 5: Small outcrop of limestone in the trail of the Coeymans Formation with the beds tilted gently to the southwest. This is a dip in the opposite direction of the beds below, showing that these rocks are slightly folded.

The trail continues for several hundred feet through a red cedar covered slope.

STOP 6: The slope on the west side of the trail with abundant ferns contains slightly calcareous shale (they fizz with acid) of the New Scotland Formation. Note that these shales produce much smaller fragments than the limestones below.

STOP 7: Foundation or rock wall on the west side of the trail. The large boulder of siltstone is from the Esopus Formation which probably slumped down from above.

STOP 8: The trail then ascends through glacial till comprising a wide variety of boulders, some rounded, one of which is about 5 feet long and of slightly calcareous siltstone of the Schoharie Formation that may have been carried by the glacier from lower down on the ridge, possibly from Stop 9. Fragments of siltstone of the Esopus Formation, both large boulders and small flakes, are seen further down the trail. These are mixed with a variety of boulders of glacial derivation. This glacial till becomes more abundant down slope.

STOP 9: Hidden behind cedars and other shrubbery is a rib of the Schoharie Formation. It is a siltstone, but differs from the Esopus by being denser and slightly calcareous. The Schoharie holds up a small knob along the trail at this point. Its beds dip 14 degrees to the northwest.

STOP 10: This flat terrace is composed of stratified sand and gravel, some of which can be seen in the bank below, derived from streams that originated from the melting glacier to the northeast up the Delaware Valley.

Excavations to the west are being performed by Pam Crabtree of New York University, uncovering evidence for Fort Johns (Shapnack).

A porta potty marks the end of the trail.

The cliffs across the Delaware River to the northwest are developed in siltstones at the base of the Pocono Plateau.

# Appendix C

***A FLOAT THROUGH TIME DOWN FLATBROOK BEND***
***THE GEOLOGY OF THE DELAWARE RIVER***
**Field Conference of Pennsyvania Geologists**
**Pre-Trip canoe trip, October 4, 2001**

Leaders:
Jack Epstein, US Geological Survey
Ron Witte, New Jersey Geological Survey
Don Monteverde, New Jersey Geological Survey
Jon Inners, Pennsylvania Geological Survey
Megan O'Malley, National Park Service

**1. Bushkill Access**. Overview of trip; introduction to geology

**2. Foxtown Member of the Buttermilk Falls Limestone**. Limestone, chert, crinoid fossils, fractures, stream terraces. Park upriver of outcrop on sandy beach.

**3. New Scotland Formation**. Fold structure, fossils. Park canoes along rock face. Observe massive coral reef in the Coeymans formation upon disembarking.

**4. Bossardville Limestone and Decker Formation**. Finely layered limestone and fossiliferous limestone. Park on beach downstream of outcrop.

**5. Riverbend campground glacial deposits.** Park at beach and hike to top of terrace.

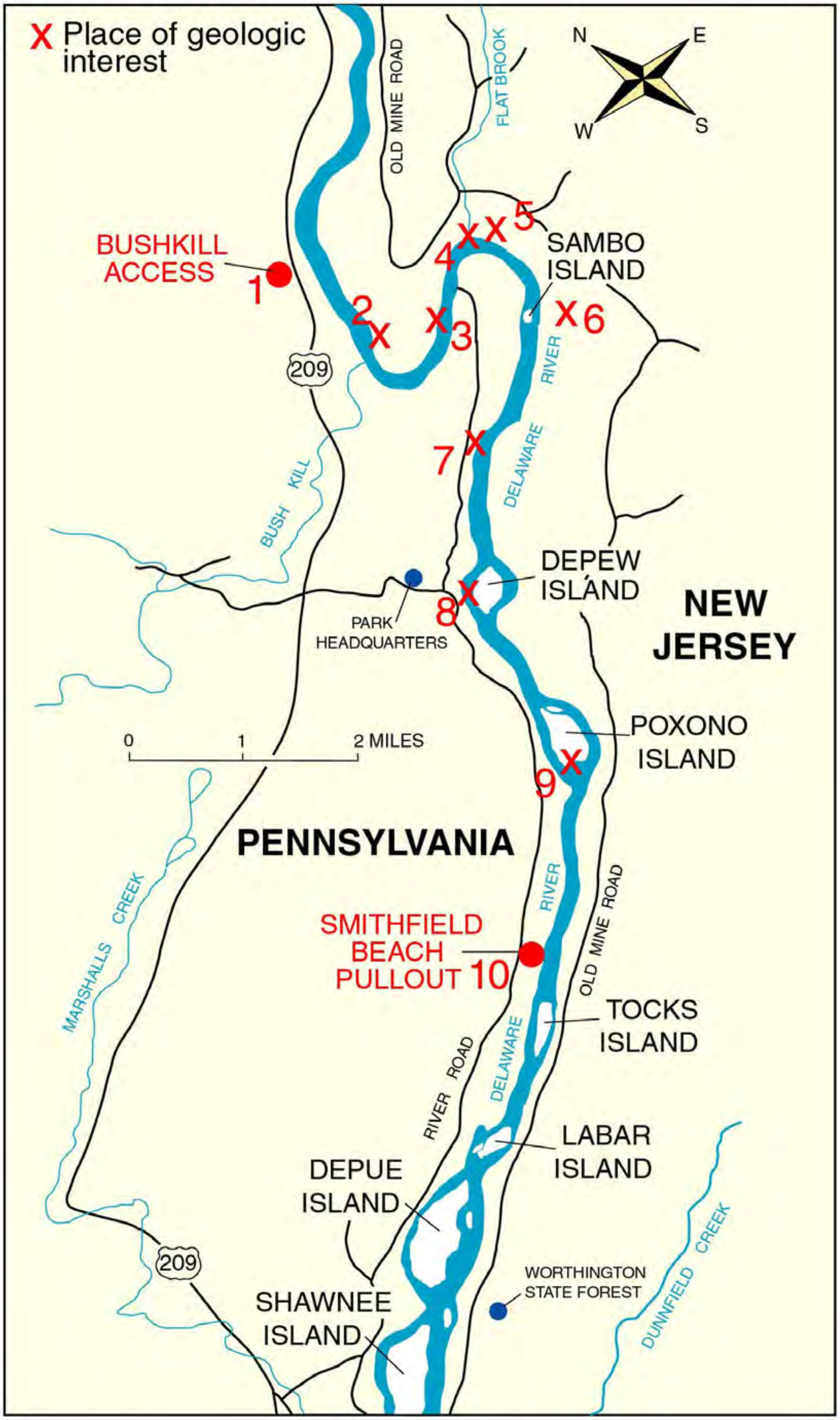
**6. Sambo Island landslide**. Bloomsburg Red Beds, till, glacial striae, erratics. Park canoes upstream of slide.

**7. Mystery formation--what is it?** Have we seen this rock before? If so, what is it? Park canoes downstream of outcrop.

**8. Hibachi rock; Bossardville Limestone/Decker Formation**. Mudcracks, algal laminations, micro-channels, corals and other critters, river jumping.

**9. Stream terrace**.

**10. Smithfield Beach. Tocks Island Discussion, pullout.**


X Place of geologic interest
OLD MINE ROAD
FLAT BROOK
N
E
W
S
BUSHKILL ACCESS
1
2
3
4
5
SAMBO ISLAND
6
209
RIVER
7
DELAWARE
BUSH KILL
DEPEW ISLAND
8
PARK HEADQUARTERS
NEW JERSEY
POXONO ISLAND
0
1
2 MILES
9
PENNSYLVANIA
MARSHALLS CREEK
RIVER
OLD MINE ROAD
SMITHFIELD BEACH PULLOUT 10
TOCKS ISLAND
RIVER ROAD
DELAWARE
LABAR ISLAND
DEPUE ISLAND
209
WORTHINGTON STATE FOREST
DUNNFIELD CREEK
SHAWNEE ISLAND

# Appendix D

Table D1

| Date | Time | Weather[1] | Temp A[2] | Temp B[2] | Temp C[2] | Wind Speed[3] | Wind Direction[4] |
|---|---|---|---|---|---|---|---|
| 09/07/00 | 9:25 AM | S | 56.4 | 58.1 | 56.2 | 128 | In |
| 09/07/00 | 10:48 AM | S | 68.6 | 69.7 | 68.8 | 0 | |
| 09/10/00 | 1:34 PM | C | 57.1 | 71.6 | 80.6 | 223 | Out |
| 09/16/00 | 9:32 AM | S,W | 61.6 | 60.8 | 60.5 | 116 | In |
| 09/23/00 | 1:10 PM | R | 53.8 | 58.2 | 65.1 | 92 | Variable - Out |
| 09/26/00 | 1:48 PM | R | 49.6 | 52.7 | 55.4 | 134 | In |
| 09/30/00 | 3:05 PM | S | 46.4 | 49.9 | 62.9 | 168 | Out |
| 10/03/00 | 1:27 PM | S | 49.1 | 60.4 | 78.6 | 284 | Out |
| 10/14/00 | 9:35 AM | S | 51.3 | 58.1 | 62.0 | 120 | Out |
| 10/17/00 | 10:23 AM | R | 54.0 | 54.2 | 58.6 | 0 | In |
| 10/20/00 | 1:15 PM | W | 52.5 | 62.1 | 65.0 | 140 | Out |
| 10/21/00 | 11:12 AM | S | 47.8 | 64.1 | 69.8 | 174 | Out |
| 10/24/00 | 9:35 AM | S | 45.8 | 55.3 | 54.7 | 78 | Out |
| 10/28/00 | 2:43 PM | S,W | 50.2 | 52.3 | 52.7 | 389 | In |
| 10/31/00 | 1:17 PM | S,W | 56.9 | 59.2 | 60.2 | 187 | Variable - In |
| 11/06/00 | 11:26 AM | S | 53.6 | 55.9 | 56.7 | 0 | |
| 11/07/00 | 10:32 AM | S,W | 43.8 | 54.7 | 50.1 | 116 | Out |
| 11/12/00 | 3:53 PM | S | 47.0 | 50.5 | 49.6 | 0 | Out |
| 11/18/00 | 2:16 PM | C,W | 40.6 | 44.7 | 41.5 | 92 | Variable - In |
| 11/21/00 | 2:25 PM | C,W | 37.2 | 40.3 | 41.2 | 87 | Variable - In |
| 11/28/00 | 10:36 AM | C | 33.0 | 40.6 | 51.6 | 158 | Out |
| 12/01/00 | 11:50 AM | S,W | 36.8 | 41.2 | 42.4 | 285 | In |
| 12/04/00 | 11:31 AM | S,W | 27.4 | 38.0 | 37.4 | 103 | Out |
| 12/09/00 | 11:25 AM | S,W | 32.5 | 37.5 | 40.2 | 198 | In |
| 12/27/00 | 11:47 AM | S,W | 21.2 | 32.1 | 31.3 | 110 | Out |
| 01/09/01 | 3:50 PM | S,W | 27.0 | 29.1 | 30.1 | 199 | In |
| 01/10/01 | 3:50 PM | PC | 25.7 | 28.7 | 38.3 | 129 | Out |
| 01/12/01 | 3:50 PM | S | 25.7 | 35.1 | 46.3 | 207 | Out |
| 01/15/01 | 3:49 PM | C,R | 25.5 | 32.6 | 47.0 | 147 | Out |
| 01/17/01 | 3:48 PM | C | 26.5 | 31.2 | 39.5 | 174 | Out |
| 01/22/01 | 3:51 PM | S | 23.5 | 28.3 | 40.1 | 114 | Out |
| 01/23/01 | 3:55 PM | S | 21.1 | 28.3 | 45.6 | 132 | Out |
| 01/24/01 | 3:50 PM | S | 22.1 | 29.6 | 41.5 | 166 | Out |
| 01/26/01 | 5:13 PM | S | 23.5 | 27.5 | 38.3 | 108 | Out |
| 01/29/01 | 3:48 PM | PC,W | 24.4 | 29.1 | 37.7 | 129 | Out |
| 01/31/01 | 3:53 PM | PC,W | 26.0 | 35.8 | 46.0 | 191 | Out |
| 02/06/01 | 3:45 PM | C | | | 41.0 | | |
| 02/12/01 | 3:53 PM | S,W | 23.3 | 34.9 | 38.2 | 101 | Variable - Out |
| 02/13/01 | 3:53 PM | S | 25.3 | 31.5 | 46.7 | 230 | Out |
| 02/14/01 | 3:50 PM | C,R | 25.8 | 30.9 | 41.7 | 183 | Out |
| 02/19/01 | 3:57 PM | PC,W | 25.1 | 39.2 | 44.9 | 198 | Out |
| 02/21/01 | 3:51 PM | S,W | 29.8 | 31.6 | 33.0 | 249 | In |

| Date | Time | Weather[1] | Temp A[2] | Temp B[2] | Temp C[2] | Wind Speed[3] | Wind Direction[4] |
|---|---|---|---|---|---|---|---|
| 02/26/01 | 3:50 PM | S,W | 40.8 | 43.2 | 44.6 | 209 | Variable - In |
| 02/28/01 | 3:50 PM | S,W | 32.1 | 35.6 | 36.1 | 195 | In |
| 03/12/01 | 3:56 PM | S | 31.6 | 37.1 | 49.7 | 201 | Out |
| 03/13/01 | 3:58 PM | C,R | 32.0 | 36.9 | 49.6 | 164 | Out |
| 03/19/01 | 3:57 PM | S,W | 33.9 | 50.8 | 51.0 | 210 | Variable - Out |
| 03/26/01 | 4:00 PM | PC,W | 33.5 | 35.9 | 35.7 | 231 | In |
| 03/27/01 | 3:51 PM | PC,W | 37.4 | 40.5 | 41.3 | 98 | In |
| 03/30/01 | 3:58 PM | C,R | 32.8 | 44.4 | 46.0 | 114 | Out |
| 04/02/01 | 3:50 PM | C,W | 35.8 | 46.4 | 44.6 | 70 | Variable - Out |
| 04/03/01 | 3:51 PM | C,W | 33.0 | 50.9 | 54.2 | 234 | Out |
| 04/04/01 | 3:50 PM | S,W | 32.9 | 53.8 | 56.0 | 242 | Out |
| 04/10/01 | 3:50 PM | C,W | 32.9 | 42.5 | 63.4 | 259 | Out |
| 04/11/01 | 3:58 PM | C | 32.9 | 39.9 | 56.9 | 241 | Out |
| 04/13/01 | 3:42 PM | S,W | 33.0 | 69.7 | 73.0 | 286 | Out |
| 04/16/01 | 4:30 PM | C,W | 33.6 | 46.4 | 58.1 | 215 | Out |
| 04/18/01 | 3:56 PM | PC,W | 46.2 | 49.1 | 48.2 | 138 | In |
| 04/20/01 | 3:49 PM | C | 33.8 | 42.0 | 63.1 | 283 | Out |
| 04/23/01 | 3:47 PM | PC | 34.4 | 47.9 | 85.1 | 345 | Out |
| 04/25/01 | 3:46 PM | PC | 37.0 | 43.3 | 64.7 | 190 | Out |
| 04/27/01 | 3:49 PM | PC,W | 37.0 | 71.6 | 72.3 | 144 | Variable - Out |
| 05/02/01 | 3:25 PM | PC | 36.4 | 47.8 | 86.2 | 360 | Out |
| 05/13/01 | 11:35 AM | S,W | 46.8 | 64.1 | 66.7 | 238 | Out |

[1] P, partly cloudy, S, sunny, C, cloudy, R, rainy, W, windy
[2]All temperature readings were recorded in the Fahrenheit scale.
Temp A - taken just inside the cave.
Temp B - taken at the top of the steps leading to and about 20 feet from the cave entrance.
Temp C - taken in the parking lot at the cave, about 50 feet from the cave entrance.
[3]Wind speed was recorded in feet per minute (FPM).
[4]Wind direction was recorded at the cave entrance.

Table D2. Summary of data at Cold Air Cave according to direction of wind flow

| Direction of wind flow | Number of readings | Temp A[2] | Temp B[2] | Temp C[2] | Wind Speed |
|---|---|---|---|---|---|
| In | 17 | 42.5 | 45.1 | 45.8 | 173 |
| Out | 44 | 34.3 | 44.9 | 54.3 | 175 |
| All | 64 | 36.8 | 45.5 | 52.1 | 170 |