2001: A DELAWARE RIVER ODYSSEY

66th Field Conference of Pennsylvania Geologists

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

Hosts
US Geological Survey
New Jersey Geological Survey
Pennsylvania Geological Survey
New York State Museum
National Park Service
Guidebook for the
66th Annual Field Conference of Pennsylvania Geologists

2001: A DELAWARE RIVER ODYSSEY

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

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North Park, Allegheny County
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.
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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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 Strellecki, 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.
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 awaaaaayyyyy 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.
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 progradation 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)

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>
<thead>
<tr>
<th>System</th>
<th>Series</th>
<th>Formation</th>
<th>Member</th>
<th>Description</th>
<th>Thickness (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper</td>
<td>Trimmers Rock</td>
<td>Millrift</td>
<td>Dark-gray to medium-dark gray siltstone, shale, and very fine-grained sandstone, coarsening upwards. Fossiliferous (brachiopods).</td>
<td>720-1,825</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Sloat Brook</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Middle</td>
<td>Mahantango</td>
<td>Sloat Brook</td>
<td>Medium-dark-gray siltstone and silty shale. Fossiliferous, biostromes (corals, brachiopods, pelecypods, bryozoans).</td>
<td>1,300-2,450</td>
<td></td>
</tr>
<tr>
<td>Middle</td>
<td>Marcellus</td>
<td>Brodhead Creek</td>
<td>Dark-gray, laminated to poorly bedded silty shale; depauperate brachiopods. Medium-dark gray shaly limestone.</td>
<td>800-950</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Stony Hollow</td>
<td>Medium-dark-gray to medium-gray, laminated to thin-bedded, shaly limestone, fossiliferous (brachiopods).</td>
<td>150</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Union Springs</td>
<td>Medium-dark-gray to dark-gray laminated shale; sheared along detachment.</td>
<td>50</td>
<td></td>
</tr>
<tr>
<td>Lower</td>
<td>Middle</td>
<td>Seneca</td>
<td>Fossiliferous cherty limestone. Contains TIOGA ash bed.</td>
<td>15</td>
<td></td>
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<tr>
<td></td>
<td>Onondaga (Buttermilk Falls)</td>
<td>Moorehouse (Stroudsburg)</td>
<td>Medium-gray limestone and argillaceous limestone with beds, pods and lenses of dark-gray chert. Fossiliferous (brachiopods, ostracodes), burrowed.</td>
<td>135</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Nedrow (McMichal)</td>
<td>Medium-dark-gray calcareous argillite with lenses of light-medium gray fossiliferous limestone.</td>
<td>40</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Edgecliff (Foxtown)</td>
<td>Medium-dark-gray calcareous siltstone and argillaceous limestone containing lenses of dark-gray chert. Fossiliferous, one-inch diameter crinoid “columns” in lower half.</td>
<td>80</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Schoharie</td>
<td>Medium-to medium-dark gray argillaceous calcareous siltstone. Fossiliferous (brachiopods, <em>Taonurus</em> burrows in lower half, vertical burrows in upper half).</td>
<td>100-150</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>Esopus</td>
<td>Medium- to dark-gray silty shale and shale to finely arenaceous siltstone. Poorly fossiliferous. Burrowed (<em>Taonurus</em>).</td>
<td>180-300</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower</td>
<td>Ridgeley</td>
<td>Port Ewen</td>
<td>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).</td>
<td>0-16</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Shriver Chert</td>
<td>Minisink Limestone</td>
<td>Medium-dark-gray siliceous calcareous shale and siltstone and beds, lenses, and pods of dark-gray chert and minor calcareous sandstone. Fossiliferous (brachiopods), burrowed.</td>
<td>50-85</td>
<td></td>
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<tr>
<td></td>
<td>Port Ewen Shale</td>
<td>New Scotland</td>
<td>Medium-dark-gray poorly fossiliferous, irregularly laminated calcareous shale and siltstone grading up to fossiliferous, buried, irregularly bedded calcareous siltstone and shale.</td>
<td>150</td>
<td></td>
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<tr>
<td></td>
<td>Minisink Limestone</td>
<td>Coeymans</td>
<td>Dark- to medium-gray argillaceous fossiliferous limestone.</td>
<td>11-14</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Stormville</td>
<td>Medium-gray, fine- to coarse-grained, biogenic limestone, fine-to medium-grained arenaceous limestone, fine- to coarse-grained, crossbedded and planar bedded calcareous sandstone and quartz-pebble conglomerate, with some dark-gray chert. Fossiliferous (brachiopods, crinoids).</td>
<td>0-20</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Shawnee Island</td>
<td>Shawnee Island: Medium-gray, argillaceous and arenaceous irregularly bedded fossiliferous and buried limestone with chert at top. Contains bioherms of medium-light-gray very coarse grained crudely bedded biogenic limestone with corals, stromatoporoids, and shelly fauna (<em>Gypidula</em>). Thacher: Dark-gray, unevenly bedded limestone with yellowish-gray shale partings.</td>
<td>0-60</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Thacher Mbr of Manlius Fm</td>
<td>Medium-dark gray argillaceous massive fossiliferous limestone (diversified fauna) with nodules and lenses of dark-gray chert.</td>
<td>0-60</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Kalkberg Limestone</td>
<td></td>
<td>0-35</td>
<td></td>
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<td></td>
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<td>Peters Valley</td>
<td>Medium-gray arenaceous limestone to light-medium-gray fine- to coarse-grained pebbly calcareous sandstone. Cross bedded, fossiliferous.</td>
<td>0-9</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Depue Limestone</td>
<td>Medium- to dark-gray arenaceous and argillaceous fossiliferous Limestone.</td>
<td>13-29</td>
<td></td>
</tr>
<tr>
<td>SILURIAN AND DEVONIAN</td>
<td>Up. Silurian &amp; Low. Devonian</td>
<td>Ronda</td>
<td>Mashipacong</td>
<td>Medium-dark- to light-gray shale, calcareous shale, and very fine- to medium-grained argillaceous limestone. Mudcracks, cut and fill.</td>
<td>8-15</td>
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<td></td>
<td></td>
<td>Whiteport Dolomite</td>
<td>Dark- to medium-gray mud-cracked laminated dolomite.</td>
<td>5-10</td>
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<tr>
<td></td>
<td></td>
<td>Duttonville</td>
<td>Dark- to medium-gray calcareous shale and argillaceous limestone.  Mud-cracked intervals and biostromal limestone beds.</td>
<td>10-20</td>
<td></td>
</tr>
<tr>
<td>Upper</td>
<td>Decker</td>
<td>Wallpack Center Clove Brook</td>
<td>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. Clove Brook: Medium-gray to medium-dark gray fossiliferous (crinoidal) limestone with light-olive-gray shale partings near top.</td>
<td>0-85</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Bossardville Limestone</td>
<td>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).</td>
<td>12-110</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Poxono Island</td>
<td>Light-olive-gray to green, calcareous and dolomitic, laminated, fissile to nonfissile shale, olive-green dolomite, sandstone, and siltstone.</td>
<td>500-800</td>
<td></td>
</tr>
<tr>
<td>SILURIAN</td>
<td>Middle &amp; Upper</td>
<td>Bloomsburg Red Beds</td>
<td>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.</td>
<td>1,500</td>
<td></td>
</tr>
<tr>
<td>Lower and Middle</td>
<td>Shawangunk (Members loose their identity several miles northeast of Delaware Water Gap)</td>
<td>Tammany</td>
<td>Gray, fine- to coarse-grained, partly crossbedded, pyritic conglomerate, evenly bedded quartzite, and about 2% dark-gray argillite.</td>
<td>800</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lizard Creek</td>
<td>Gray to olive-gray, fine- to coarse-grained, partly crossbedded, pyritic, thin- to thick-bedded quartzite interbedded with thin-to thick bedded, gray argillite.</td>
<td>275</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Minsi</td>
<td>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.</td>
<td>300</td>
<td></td>
</tr>
<tr>
<td>ORDOVICAN</td>
<td>Middle and Upper</td>
<td>Martinsburg</td>
<td>Pen Argyll</td>
<td>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.</td>
<td>3,000-6,000</td>
</tr>
<tr>
<td></td>
<td>Middle and Upper</td>
<td>Ramseyburg</td>
<td>Medium- to dark-gray claystone slate alternating with light- to medium-gray, thin- to thick-bedded graywacke and graywacke siltstone.</td>
<td>2,800</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bushkill</td>
<td>Dark- to medium-gray thin-bedded (beds do not exceed six inches thick), claystone slate with thin interbedded quartzose slate and graywacke siltstone and carbonaceous slate. Not exposed in Delaware Water Gap National Recreation Area.</td>
<td>4,000</td>
<td></td>
</tr>
</tbody>
</table>

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
“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
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 4A). 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 4B). 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° 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 finier 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 Formation.
   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.

**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.

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
Oriskany Group, and Coeymans Formation. Neritic deposits consist predominately 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...
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.
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°. Bedding and fold patterns in the Martinsburg within several thousand feet of the contact are discussed at STOP 3.

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 7º, 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-
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
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 microscars 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 12. Bedding thrust and ramp at base of sandstone in the Bloomsburg Red Beds along Old Mine Road in New Jersey.

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; Depv, Peters Valley Member of the Coeymans Formation; Dcd, Depue Limestone Member of the Coeymans Formation; Ds, Stormville Member of the Coeymans Formation; Des, 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; Dbsf, 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.
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.
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 15A 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 15A. 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 15B) 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 15A 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 15B. 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.
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°) than Alleghanian structures in southeastern New York.
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°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

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.
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
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° toward N32°E. The Martinsburg fold trends, as we see them now, are more northerly, by about 10° 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°.
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°) 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°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°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°W with an 8° 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

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 $\pm 3^\circ$.

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°C and a characteristic or C component with discrete unblocking temperatures between 660°C and 690°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 prepectonic 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 $R_f$-$\phi$ 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°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 $D = 355.2^\circ$, $I = -25.2^\circ$ ($k = 14.8$) which corresponds to a north paleopole position at $35.6^\circ$N, $110.0^\circ$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 12º-14º 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

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.

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.

**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.

**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 Etmothermys (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.

**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 (Hallinetae) 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 (Lamb’s 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 Palmerton Sandstone, marked at its base by a distinctive, meter-thick conglomerate bed (Ver Straeten, 1996a, 1996b).

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 Polygonathus costatus patulus conodont zone), is presently not recognized, but occurs between the base and top of the Edgecliff Member. Conodonts diagnostic of the succeeding Polygonathus costatus partitus conodont zone are found within the Nedrow Member; the top of the Nedrow marks the first occurrence of forms diagnostic of the Polygonathus 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 Fimbriospirifer 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, stylolind-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, Stylolina) 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 (Guericiphylum, 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
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
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 hightand 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 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. These 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 Genesee 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
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
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 *Atlanticocoelina, Coelosspira, 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)
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.


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*. 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.
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 Atlanticocoelidia 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).

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 Atlanticocoelidia 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.

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.
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.

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

Figure 33. Location map of Riccobono’s “Quarry in the Schoharie” along US 209 in Marshalls Creek.

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.

**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
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.

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°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.

The whitish-weathered siltstone ledge directly above the crevasse is smoothed and polished and exhibits glacial striae trending S60°W, parallel to the ridge (Figure 40). Striae on top of the ledge higher on the hill trend S28°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.
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.

SITE DESCRIPTION AND METHODS

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°); 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 Subzone.
GRAPTOLITE ZONE CORRELATIONS
(After Ross et al. 1982)

<table>
<thead>
<tr>
<th>Zones</th>
<th>Subzones</th>
</tr>
</thead>
<tbody>
<tr>
<td>13. Orthograptus amplexicaulus</td>
<td>Climacograptus spiniferus Orthograptus ruedemannii</td>
</tr>
<tr>
<td>12. Climacograptus bicornis</td>
<td>Corynoides americanus</td>
</tr>
<tr>
<td>11. Nemograptus gracilis</td>
<td></td>
</tr>
<tr>
<td>10. Glyptograptus cf. G.teretiusculus</td>
<td></td>
</tr>
<tr>
<td>9. Paraglossograptus etheridgei</td>
<td></td>
</tr>
<tr>
<td>8. Isograptus</td>
<td></td>
</tr>
<tr>
<td>7. Didymograptus bifidis</td>
<td></td>
</tr>
<tr>
<td>6. Didymograptus protobifidus</td>
<td></td>
</tr>
<tr>
<td>5. Tetragraptus fruticosus (3 and 4-branched)</td>
<td></td>
</tr>
<tr>
<td>4. Tetragraptus fruticosus (4-branched)</td>
<td></td>
</tr>
<tr>
<td>3. Tetragraptus approximatus</td>
<td></td>
</tr>
<tr>
<td>2. Clonograptus</td>
<td></td>
</tr>
</tbody>
</table>

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.

<table>
<thead>
<tr>
<th>Taxon</th>
<th>NJSM Catalogue Number</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Conularia</em> cf. <em>trentonensis</em> (Hall)</td>
<td>19577</td>
</tr>
<tr>
<td>cf. <em>Prasopora</em> sp.</td>
<td>20153</td>
</tr>
<tr>
<td><em>Plectorthis plicatella</em> (Hall)</td>
<td>20154</td>
</tr>
<tr>
<td><em>Orbiculoidea</em> sp.</td>
<td>20156</td>
</tr>
<tr>
<td>cf. <em>Sinuites</em> sp.</td>
<td>20168</td>
</tr>
<tr>
<td>cf. <em>Hormotoma</em> sp.</td>
<td>20167</td>
</tr>
<tr>
<td><em>Michelinoceras</em> sp.</td>
<td>20157</td>
</tr>
<tr>
<td><em>Flexicalymene</em> cf. <em>senaria</em> (Conrad)</td>
<td>20158</td>
</tr>
<tr>
<td>cf. <em>Pterygometopus</em> sp.</td>
<td>20159</td>
</tr>
<tr>
<td>Crinoidea, indeterminate</td>
<td>20160</td>
</tr>
<tr>
<td><em>Glyptograptus euglyphus</em> Lapworth</td>
<td>20161</td>
</tr>
<tr>
<td><em>Orthograptus amplexicaulus</em> (Hall)</td>
<td>20169</td>
</tr>
<tr>
<td><em>Climacograptus</em> cf. <em>mohawkensis</em> (Ruedemann)</td>
<td>20155</td>
</tr>
</tbody>
</table>
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.
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.

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

<table>
<thead>
<tr>
<th>Association</th>
<th>Facies</th>
<th>Flexible</th>
<th>UPGMA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sparse containing bivalves and brachiopods</td>
<td>1</td>
<td>A</td>
<td>C</td>
</tr>
<tr>
<td><em>Cystiphyloides americanum - Coenites</em> sp.</td>
<td>2</td>
<td>B-1</td>
<td>C</td>
</tr>
<tr>
<td><em>Helophyllum halli - Spiriferid brachiopods</em></td>
<td>3</td>
<td>C</td>
<td>C</td>
</tr>
<tr>
<td><em>Cystiphyloides americanum - Helophyllum halli</em></td>
<td>4</td>
<td>B-1</td>
<td>C</td>
</tr>
<tr>
<td><em>Helophyllum halli - Spiriferid brachiopods</em></td>
<td>5</td>
<td>B-1</td>
<td>C</td>
</tr>
<tr>
<td>Sparse containing bivalves and brachiopods</td>
<td>6</td>
<td>A</td>
<td>A</td>
</tr>
</tbody>
</table>

Figure 46. Brodhead Creek outcrop with associations, facies, and cluster analysis results.
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, Fimbri spirifer 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., Nuclooidea 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, Spinocystia 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., Proleptostrophia 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 audaculus*, 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.

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*, *Breviphereis pumilla*, *Heterophereis* sp., *Heliophyllum halli*, an indeterminate branching form of *H. halli*, *Eridophyllum archiaci*, *Cystiphylloides americanum*, *Favosites* sp. aff. *Favosites placenta*, *Favosites* sp. aff. *Favosites placenta*, *Favosites* sp. aff. *Favosites placenta*, *Favosites* sp. aff. *Favosites placenta*.

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.)
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 archaicai, 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., *Nucloida* sp., *Paleoneilo* sp., *Mediospirifer* sp., *Platylophax* 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*, *Fimbriopirifer* sp., *Elita* sp., *Atrypa* sp., *Cypricardella* sp., *Nucloidea* sp., *Phacops rana*, *Greenops boothi*, *Fenestrella* 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 *Mediospirifer* 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., *Protolampas* sp., and *Stereolasma rectum*. The association also includes rare *Phacops rana* and *Fenestrella* 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 than 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 hamiltonae and large Favosites sp. occur from 5 to 5.1 meters. Other common corals include Breviphereritis pumilla, Breviphereritis 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, Trachyypora 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 WISCONSINIAN 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 20th 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.
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**

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

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**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.
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.

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

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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.
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**

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 $^{26}\text{Al}$ 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 ± 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 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 ± 695 yr B.P. (GX-4279), and 18,570 ± 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 ± 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

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 glaciitectonized 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.

**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.
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 depressions. In places they are only partially enclosed forming small amphitheater-shaped depressions. 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 not 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.

**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

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 ductily.

![Lower till (B).](image1)

![Deformed lacustrine bed (in unit C).](image2)

![Proglacial outwash (A). Photo taken in lower pit (scale in 1 foot gradations).](image3)

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.

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. 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 the 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?

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 ($x10^2$ to $x10^3$ 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 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.
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

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
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 (McNott, 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 Remeter 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
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 debouched 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° 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.
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
miles (12.9 km) north of the Terminal Moraine indicates a minimum age of deglaciation at 18,570 ± 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 ± 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.
Table 3. Correlation of major ice margin positions and morphostratigraphic units in 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 morphostratigraphic 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 were comparable to other dated pollen sequences in northeastern Pennsylvania and New Jersey that additionally contained lower and older pollen. These basal deposits consistently contained pollen spectra characteristic of the herb pollen zone, an indication of tundra vegetation and a cold and wet climate. Cotter believed that organic sedimentation and lake formation had been delayed in Minisink Valley and on the Pocono Plateau. The older date of deglaciation also accords well with younger dates in the upper part of the Susquehanna drainage basin cited in Fleisher (1986) and Ozvath and Coates (1986).

Based on the morphosequence concept of Koteff and Pessl (1981), valley-train deposits show that the Minisink Valley lobe retreated to the northeast in a systematic manner, and chiefly by stagnation-zone retreat. Successive ice-retreat positions are marked by the heads of the Zion Church, Dingmans Ferry, Montague, and Tristates valley trains. Four end moraines, named Millbrook Village, Dingmans Ferry, Montague, and Millville further delineate ice-marginal positions, and both the Dingmans Ferry and Montague moraines are directly traceable eastward where they join the Ogdensburg-Culvers Gap, and Augusta moraines. Outwash-fan deposits provide additional information that supports a marginal retreat model. These deposits lie at the mouths of tributaries that feed the Minisink 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
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 define 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. Argillie 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.

<table>
<thead>
<tr>
<th>Years B.P.</th>
<th>Ecological Period</th>
<th>Cultural Period</th>
<th>Pollen Zone</th>
<th>Climate</th>
<th>Vegetation</th>
<th>Fluvial Activity</th>
<th>Radiocarbon Dates</th>
<th>Soil Stratigraphic Unit</th>
<th>Morphostratigraphic Unit</th>
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<tr>
<td></td>
<td>Sub-Atlantic</td>
<td>Woodland</td>
<td>Oak-mixed</td>
<td>Similar to modern conditions</td>
<td>Oak and chestnut forest</td>
<td>Flood plain stability Minor alluviation Formation of the modern flood plain</td>
<td></td>
<td>Ab1</td>
<td>T1</td>
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<td></td>
<td></td>
<td></td>
<td>Hardwoods</td>
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<td></td>
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<td>Ab2</td>
<td></td>
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<tr>
<td></td>
<td>Subboreal</td>
<td></td>
<td></td>
<td>Warmer and dryer, (very low ppt. rates)</td>
<td>Oak and hickory forest</td>
<td>Flood plain stability Minor alluviation</td>
<td></td>
<td>Ab3</td>
<td>T2</td>
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<td></td>
<td>Atlantic</td>
<td>Archaic</td>
<td>Oak-Hemlock</td>
<td>Warmer with near-modern conditions</td>
<td>Oak and hemlock forest; gradual decline of hemlock</td>
<td>Increased alluviation during the latter part of the Atlantic Period</td>
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<td>Ab5</td>
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<tr>
<td></td>
<td>Boreal</td>
<td></td>
<td>Pine</td>
<td>Warm and dry</td>
<td>Pine and birch to pine and oak forest</td>
<td>Channel of the Delaware River at or slightly above modern level.</td>
<td>(4) 9330</td>
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<td>(3) 10,750</td>
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<td></td>
<td>Paleolithic</td>
<td></td>
<td>Spruce</td>
<td>Cool and wet</td>
<td>Spruce and fir forest</td>
<td>Incision, abandonment of the T3 flood plain Initial formation of the T2 flood plain</td>
<td>(2) 12,160</td>
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<tr>
<td></td>
<td>Postglacial</td>
<td></td>
<td>Herb</td>
<td>Cold and wet</td>
<td>Open tundra to spruce parkland</td>
<td>Cut off of meltwater, shift from a braided to a nonincising/deterging stream. Lateral channel migration and flood plain formation.</td>
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<td>Qes</td>
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<td></td>
<td>Glacial</td>
<td></td>
<td></td>
<td>Coldest</td>
<td>Open tundra</td>
<td>Deposition of outwash Aggradation near the glacial margin, erosion down stream</td>
<td>(1) 18,570</td>
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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 ± 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 ± 300 yr B.P. (W-2994), and 10,750 ± 600 yr B.P. (W-3134), and the nearby upper Shawnee Island site (Stewart, 1991) which yielded a date of Early Archaic age (9380 ± 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.

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 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 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.
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
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.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Age</th>
<th>Stratigraphic Unit</th>
<th>Lithology and structure</th>
<th>Approximate Thickness (feet)</th>
</tr>
</thead>
<tbody>
<tr>
<td>IV</td>
<td>Middle-Upper Devonian</td>
<td>Catskill Formation&lt;br&gt;Trimmers Rock Formation&lt;br&gt;Mahanantango Formation&lt;br&gt;Marcellus Shale</td>
<td>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.</td>
<td>5,000+</td>
</tr>
<tr>
<td>III</td>
<td>Upper Silurian-Middle Devonian</td>
<td>Buttermilk Falls Limestone&lt;br&gt;Schoharie Formation&lt;br&gt;Esopus Formation&lt;br&gt;Oriskany Group&lt;br&gt;Helderburg Group&lt;br&gt;Rondout Formation&lt;br&gt;Decker Formation&lt;br&gt;Bossardville Limestone&lt;br&gt;Poxono Island Formation</td>
<td>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.</td>
<td>700-1,500</td>
</tr>
<tr>
<td>II</td>
<td>Middle and Upper Silurian</td>
<td>Bloomsburg Red Beds&lt;br&gt;Shawangunk Formation</td>
<td>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.</td>
<td>3,000</td>
</tr>
<tr>
<td>I</td>
<td>Upper Ordovician</td>
<td>Martinsburg Formation</td>
<td>Slate and greywacke in folds averaging 2,000 feet in wave length and 1,200 feet in amplitude.</td>
<td>1,000+</td>
</tr>
</tbody>
</table>

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.

**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.

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°), is composed of red siltstone, and is partly burrowed and mudcracked. The surface is highly polished and striated (Figure 70), glacial

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.
movement towards S40°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° 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,
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° E) and which dip steeply towards the road (50-70° 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°-53° northeast; Figure 78C,D, and E) and fracture cleavage (averaging N60°E, 65°SE; Figure 78D). 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°. 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° and 44° northwest (Figure 78D). 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 78C 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 78A 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°. 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.
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.

A. Longitudinal sheeting joints dipping 48° 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. Sheet 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 57° 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.

E. Irregular cross joints cutting steep longitudinal joint surface and separating rock into masses 2-8 feet wide.
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°E, 41°SE, and slickensides plunge 27°, S34°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°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° 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 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.

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.
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°E, 65°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).

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°, 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
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.
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.

The confining pressure against cross joints may be lessened by rapid erosion of rock along streams or by the

**ROCKFALL HAZARD AT LEHIGH GAP**
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 90A). PA 248 is cantilevered between two railroads as it enters the gap (Figure 90A), 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 90A).

The cross joints are irregular to roughly planar (Figure 90B, 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°E, and the cross joints trend about N20°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.
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.

CONCLUSION

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.
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).

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°/8°N and has widely-spaced joints oriented N69°/80°N and N4°/86°E. Bedding at Silverthread Falls is oriented N18°/10°N and has joints oriented N7°/90°, N60°/90°, and N44°W/79°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°E/66°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.
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.

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.

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 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°E/90° 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.

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 melting 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 (41º 02’ 51”N /75º 06’ 36”W, Bushkill quadrangle; Figure 101). This specimen is now called the Marshalls Creek mastodon after the area of entombment.

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

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.

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)

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 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.

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).

HISTORY OF MINING AT PAHAQUARRY

1750s

Historical records state that copper exploration in the Pahaquarry region began in the 1750’s. Between 1753-55, three men, John Reading Jr., Anthony Maxwell, and Martin Ryerson bought and 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 45º 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 contract 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 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.
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
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.

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
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.
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°E/42°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 (Cu₂S, copper sulfide) as the main copper-bearing mineral in the mine. Secondary copper minerals, malachite (Cu₂(CO₃)(OH)₂, copper carbonate hydroxide) and chrysocolla (CuSiO₃·nH₂O, 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° C. Later, meteoric-water interaction aided the growth of the secondary minerals, malachite and chrysocolla.
The Field Trip

And Awaaaaayyyyyy We Go

Arnold Palmer and Jackie Gleason at Shawnee in the early 1960’s. Photo courtesy of National Golf Classics, Hot Springs, Arkansas.
Route Map for Day 1 of the 2001 Field Conference of Pennsylvania Geologists
### ROAD LOG AND STOP DESCRIPTIONS

#### DAY 1

<table>
<thead>
<tr>
<th>Miles</th>
<th>Int.</th>
<th>Cum.</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>Leave from circle in front of Shawnee Inn. The Inn and golf course are located on postglacial stream terraces that reach a maximum elevation of 330 feet, about 35 feet above the mean annual elevation of the Delaware River. An Early Archaic occupation site excavated on Shawnee Island (Stewart, 1991) was dated at 9330 ± 545 yr B.P. (Uga-5488).</td>
</tr>
<tr>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
<td>Turn left onto River Road. The road passes over a sequence of Silurian shale and dolomite that is covered in many places by thin, late Wisconsinan till.</td>
</tr>
<tr>
<td>0.1</td>
<td>0.4</td>
<td>0.4</td>
<td>Limestone of the Shawnee Island Member of the Coeymans Formation on right (measured section 14-b of Epstein et al., 1967). The Stormville Member capping the top of the Coeymans Formation is seen in the private driveway to the right (Figure 112).</td>
</tr>
<tr>
<td>0.4</td>
<td>0.8</td>
<td>0.8</td>
<td>Top of hill, continue straight ahead. Buttermilk Falls Road on right. New Scotland Formation on left</td>
</tr>
<tr>
<td>0.3</td>
<td>1.1</td>
<td>1.1</td>
<td>New Scotland Formation on right</td>
</tr>
<tr>
<td>0.4</td>
<td>1.5</td>
<td>1.5</td>
<td>Minimart on left, and Smithfield School on right sit on a late Wisconsinan outwash terrace. The terrace lies at an elevation of 400 feet, about 100 feet above the Delaware. Based on its position near the mouth of Marshalls Creek, it was probably laid down by a meltwater stream flowing down the Marshalls Creek valley.</td>
</tr>
<tr>
<td>0.3</td>
<td>1.8</td>
<td>1.8</td>
<td>Gap View Road on right.</td>
</tr>
<tr>
<td>0.1</td>
<td>1.9</td>
<td>1.9</td>
<td>Village of Minisink Hills. Good exposures of Coeymans Formation through Esopus Formation along abandoned railroad grade to right. On the crest of the hill northeast of the railroad grade is an imposing ledge of cherty Ridgeley Sandstone (the “Indian Chair”) from which Amerinds extracted a good-quality flint (P. La Porte, personal communication, 2001). Bear left at intersection. Postglacial stream terraces of Brodhead Creek on right. Late Wisconsinan outwash terraces form the higher ground on the left.</td>
</tr>
<tr>
<td>0.3</td>
<td>2.2</td>
<td>2.2</td>
<td>Cross Brodhead Creek. About 1500 feet eastward is the Shawnee-Minisink Paleoindian site Island (McNett et al., 1977). It is located on postglacial stream terrace about 20 feet above the Delaware. Work here in the 1970’s revealed a very rich and diverse, stratified cultural assemblage of Woodland, Archaic, and Paleoindian components. Radiocarbon dating of organic material collected from a hearth about nine feet deep yielded a date of 10,590 ± 300 yr B.P. (W-2994). The hearth is located in cultural zone containing Paleoindian components (clovis point, scrapers, hammerstones).</td>
</tr>
<tr>
<td>0.1</td>
<td>2.3</td>
<td>2.3</td>
<td>Cross Norfolk Southern (originally, Delaware, Lackawanna &amp; Western) Railroad.</td>
</tr>
<tr>
<td>0.1</td>
<td>2.4</td>
<td>2.4</td>
<td>I-80 overpass.</td>
</tr>
</tbody>
</table>
0.2  2.6  Turn left onto US 611 toward village of Delaware Water Gap. We are at the lower end of Cherry Creek valley near its confluence with the Delaware River. During late Wisconsinan deglaciation, proglacial lakes formed in the ice-dammed (northeast-draining) valley. Several ice-contact deltas mark ice retreat (Epstein, 1969).

0.25  2.85  Water Gap Diner on left. Favorite meeting place for trip leaders, cohorts, and other associates of dubious character.

0.05  2.9  Traffic light. Turn left on US 611 South.

0.4  3.3  Northwest-dipping rocks of the Shawangunk Formation on right.

0.1  3.4  Crest of Cherry Valley anticline in the Shawangunk at top of road. The contact between the Shawangunk Formation and Bloomsburg Red Beds is conventionally placed at the base of the lowest red bed. However, at this locality this color change migrates up and down section by as much as 700 feet, making for a peculiar map pattern (Epstein, 1973). Enter upstream side of Delaware Water Gap.

0.1  3.5  Southeast-dipping rocks in the Bloomsburg Red Beds. US 611 traverses the Bloomsburg for the next 0.8 miles in a series of small undulating, low-amplitude folds. Note the southeast-dipping cleavage.

0.8  4.3  Contact between the Shawangunk and Bloomsburg dipping 35° NW.

0.7  5.0  Turn right into parking lot at Point of Gap. Disembark.

STOP 1. DELAWARE WATER GAP: GEOLOGIC OVERVIEW—STRATIGRAPHY, STRUCTURE, FORMATION OF THE GAP, AND GLACIAL GEOLOGY. Leader: Jack B. Epstein.

INTRODUCTION

Many of the parks within National Park System (NPS) owe their uniqueness to their geologic framework. Their scenery is the result of natural processes acting upon the variety of rocks that were deposited in diverse environments in the geologic past. Knowledge of these attributes, training of NPS personnel in their proper interpretation, development of a resource database, and communication of this information to the public are important priorities of the National Park Service. Bedrock and surficial geologic mapping by the federal and two state geologic surveys in the Delaware Water Gap National Recreation Area (DEWA) has been used to prepare a variety of products useful for the unit's mission of park management and service to the public. DEWA draws from several major population centers, totaling more than 30 million people within the heart of the northeast United States urban corridor and is presently the sixth most heavily visited NPS facility in the country. It includes a scenic and mostly undeveloped 40-mile stretch of the Delaware River between Port Jervis, New York, and the world-famous Delaware Water Gap in New Jersey and Pennsylvania (Figure 113). It straddles the Pocono
Plateau on the northwest, underlain by gently inclined Devonian sandstones and shales, and complexly folded Ordovician to Devonian rocks of the Valley and Ridge to the southeast. The stratigraphic sequence spans about 65 million years. Wisconsinan glacial erosion and deposition resulted in a varied scenery. The present Delaware River has cut through a silt and sand terrace that was occupied by American Indians about 11,000 years ago.

The application of our geologic efforts emphasizes scientific interpretation, land-use planning and management, points of scientific interest to be enhanced or protected (paleontologic, structural, geomorphic, stratigraphic, glacial, economic resources), landslide susceptibility, facility location and trail design, the park's GIS data base, scientific interpretation for both park personnel and the public, preparation of geologic exhibits, and general-interest publications including nature trail guides. Results of geologic investigations efforts can be effectively utilized by the Park Service only by making our data readily available and avoiding jargon. Hopefully, much of the information in this guidebook will serve this useful purpose. (For more of a “hands-on” view of the geology of the Water Gap, see Appendix B1—a guide to the Red Dot-Blue Blaze-Dunnfield Creek Trails on Mount Tammany in New Jersey.)

**STRATIGRAPHY**

Delaware Water Gap owes its notoriety to the depth to which the river has cut through Kittatinny Mountain. Exposures of 3,000 feet of Silurian clastic rocks are nearly continuous; the entire Shawangunk Formation, with its three members, and most of the Bloomsburg Red Beds are visible (see Figure 11). To the west, in central Pennsylvania, the Shawangunk merges into the Tuscarora Sandstone below and the Clinton Formation above. To the east, in New York State, as seen from the heights of High Point at STOP 6, the Shawangunk thins and just beyond it disappears. Eastward, the Bloomsburg likewise pinches out. The Bloomsburg has been erroneously called the High Falls Shale in the past. The High Falls of New York State is actually a facies of the Poxono Island Formation which overlies the Bloomsburg. For details, see Epstein, this guidebook, p. 1.

**STRUCTURE**

Shortly after the Delaware Water Gap National Recreation Area was established in 1965, an exhibit in the kiosk at the south end of the parking lot presented an interpretation of the structure in the gap. The plaques have since disappeared from the site as well as from most memories. Figure 114 brings back those memories.

This structural interpretation alludes to the fact that the Green Pond Conglomerate, the correlative of the Shawangunk, is exposed about 25 miles to the east in New Jersey. Hence, a way was needed to bring the rocks of the Shawangunk at Delaware Water Gap down again to mate with
the Green Pond rocks and a broad regional anticline was invoked. Satellitic folds that verge to the southeast would indicate that such an anticline does indeed lie to the southeast (Figure 115A). On the contrary, mapping along strike and down the plunge of the folds to the southwest on Kittatinny Mountain (Epstein, 1973), shows that an overturned syncline extends upwards from the rocks at Delaware Water Gap (Figure 115B). Because the terrain south of Kittatinny Mountain is replete with thrust faults, the structural relations between it and the Green Pond area is certainly much more complex than a simple regional anticline.

**The Arch of Cleavage**

Most pelitic rocks in the Delaware Water Gap area, regardless of age, have a secondary foliation, or slaty cleavage, which becomes more prominent northwestward and into higher stratigraphic units. A second slip cleavage, which crenulates the earlier-formed slaty cleavage, is locally developed in all units. Microscopic and field relations of the cleaved rocks suggest that slaty cleavage formed by pressure solution of more soluble minerals along anastomosing folia, leaving behind a residuum of carbonaceous matter and iron oxides. This was accompanied by mechanical reorientation of platy and elongate minerals and by some new mineral growth. Elongation of quartz and its removal from cleavage folia resulted from corrosion by pressure solution perpendicular to the cleavage direction. The cleavage folia are separated by more quartz-rich areas in which reorientation of platy minerals and dimensional alignment of prismatic minerals has not taken place, or is not as well developed.

Numerous lines of evidence point to the conclusion that cleavage developed after the rock was indurated. Plasticity increased during increased tectonism and the mobility (invasion of pelitic and sandy material along cleavage planes) may have been aided by silica derived from pressure solution and derived either from connate water squeezed out of the rocks during tectonic compaction or by the release of water from hydrous minerals during continued deformation. New growth of quartz, chlorite, muscovite, calcite, and probably albite in most rocks suggests formation of cleavage at and just below the limits of low-grade metamorphism (quartz-muscovite-albite-chlorite subfacies of the greenschist facies). Slip cleavage crenulates earlier foliations. Transposition of minerals into the new cleavage plane is common, and in this respect it is similar to slaty cleavage. It is also similar to slaty cleavage in that new minerals may grow parallel to the cleavage direction. To the southwest in Pennsylvania the slip cleavage appears higher in the Martinsburg, and in the Lehigh Gap area it is found in overlying formations, paralleling the increased development of the earlier slaty cleavage in younger units.
Intensity of slaty cleavage development increases to the southwest in Pennsylvania, and commercial slates appear higher in the section where, near Lehigh Gap, slate has been extracted from the Mahantango Formation, although now those operations have ceased. The slaty cleavage bears a geometric relationship to the folds in which it is found, fanning the folds by either opening or closing towards the anticlinal crest. In many places, particularly in the Martinsburg, but not exclusively, the slaty cleavage is cut by a second generation slip cleavage and the earlier slaty cleavage is rotated into arches by the folding process. At Delaware Water Gap, and at other localities near the contact with the competent rocks of the Shawangunk Formation, the slaty cleavage is also arched, but by a different process than by external rotation. Figure 116 is a generalized geologic map of the Delaware Water Gap area. Note that about 2,000 feet south of the Martinsburg-Shawangunk contact the cleavage dips to the southeast, but turns to the northwest as the contact is approached (Figure 116). This is similar to the structural situation seen at Yards Creek (STOP 3).

Drake et al. (1960) and Maxwell (1962) attributed this arching of the slaty cleavage to refolding during the Appalachian orogeny. However, the form of this cleavage fold in the Martinsburg is not reflected upwards into the overlying rocks. The contact between the Martinsburg and Shawangunk Formations is exposed at about a dozen localities between southeastern New York and Lehigh Gap, Pennsylvania, a distance of more than 100 miles. On the basis of observations at these localities and from data gathered during mapping along the contact, it is concluded that the arching of cleavage at Delaware Water Gap is due a strain-shadow mechanism in the trough of a syncline in the Shawangunk as shown in Figure 117 and initially described by Epstein and Epstein (1967, 1969).

In many small folds involving interbedded shale and siltstone which are cleaved and more competent rocks which are less cleaved, the slaty cleavage diverges around synclinal troughs and is either poorly developed or absent in the pressure-shadow area next to the trough (Figure 117B). This relationship is the same on a larger scale (Figure 117A), explaining the arching of cleavage at Delaware Water Gap and Yards Creek. It also explains the dying out of cleavage near the Martinsburg-Shawangunk contact elsewhere, such as at Lehigh Gap (Epstein et al., 1969). Similarly, in thin section, cleavage is seen to curve around clastic grains, small lenses of sandstone, or sand-filled burrows. The cleavage is most intensely developed (flattening is greatest) on top and bottom of these more competent...
clastic bodies and is poorly developed or absent in the areas of maximum extension to the sides of the grains in the areas of "pressure shadows."

A few additional comments about the relationships at the Taconic unconformity at Delaware Water Gap and the surrounding area are worth noting. Quartz, chert, and quartzite pebbles in the basal beds of the Shawangunk, in places more than 5 inches long, indicate that the Martinsburg was breached during Silurian time and that underlying stratigraphic units were exposed and supplied the pebbles (possibly chert from the Ordovician Beekmantown Group, quartzite from the Cambrian Hardyston Quartzite, and vein quartz from Precambrian rocks). The sharp lithologic break at the contact brings together rocks of vastly different origin—deep-water shales and turbidite sandstones of the Martinsburg are overlain by fluviatile-terrestrial deposits of the Shawangunk. Within the basal Shawangunk no fragments of shale from the underlying Martinsburg contain slaty cleavage that may have been produced during Taconic deformation. Rather, any cleavage that may be present conforms to the attitude of the regional cleavage in post-Ordovician rocks. The obvious conclusion is that no Taconic cleavage can be recognized in pebbles within Silurian rocks. Additionally, in a few localities folds have been mapped along the unconformity, such as at Yards Creek (STOP 3) and High Point (STOP 6). The fold axes pass from the Shawangunk into the Martinsburg Formation without deflection, showing that the folds are post-Taconic in age. Cleavage in the Martinsburg is parallel to the axial planes of the folds, or fans the folds (except for the arching of cleavage as described above), again showing that the cleavage is post-Taconic in age.
A STORY OF THE GAP

Anyone who maps in the Delaware Water Gap area is compelled to contemplate the history of formation of the gap and why it is where it is. The following thoughts are summarized from Epstein (1966, 1997).

Many different sediments were deposited in eastern Pennsylvania and northern New Jersey during the Paleozoic. Later, during consolidation, some of these became sandstones and conglomerates, resistant to subsequent erosion and formed mountains; others became shales, limestones and dolomite, less resistant to erosion and formed valleys. Following orogenic uplift of these diverse sediments during the late Paleozoic, the original divide of the Appalachian Mountains that formed lay somewhere to the south within the area of the present Piedmont or Valley and Ridge Province. During rifting and opening of the Atlantic ocean, that divide shifted westward towards its present position in the Appalachian Plateau because the steeper stream gradients towards the Atlantic Ocean created an erosional advantage over the lower gradient streams that flowed westward towards the continental interior. The manner of migration of that divide and how the streams cut through the resistant ridges are critical elements in any discussion of Appalachian geomorphic development. These subjects have been a source of considerable controversy for more than a century. In the area of this Field Conference, there are many wind and water gaps in Blue and Kittatinny Mountains. That ridge is held up by resistant quartzites and conglomerates of the Shawangunk Formation of Silurian age in Pennsylvania and New Jersey, extending into Shawangunk Mountain in New York (see Figure 9). The Shawangunk thins to the northeast and disappears above Ellenville, NY (Epstein, 1993). Viewed from a distance, these gaps or low sags interrupt the fairly flat ridge top that was termed the “Schooley peneplain” by Davis (1889) and popularized by Johnson (1931). Ideas on the origin of these gaps are critical factors in several hypotheses that discuss the geomorphic development of the Appalachians. Those hypotheses that favor down cutting (superposition) from an initial coastal plain cover (Johnson, 1931; Strahler, 1945) require that the location of the gaps be a matter of chance. Those hypotheses that suggest the present drainage divide was inherited from the pattern already established following the Alleghanian orogeny and controlled by the topography and structure prevalent at the time (Meyerhoff and Olmstead, 1936) or by headward erosion into zones of structural weakness (headward piracy, Thompson, 1949) require that there be evidence for structural weakness at the gap sites. Thus, an understanding of the structural configuration of these gaps is necessary for adequately discussing the drainage evolution of the Appalachians.

Sixteen gaps and cols in Blue, Kittatinny, and Shawangunk Mountains between Lehigh Gap in eastern Pennsylvania and Ellenville, New York, were examined (Epstein 1997). Most of the gaps are located at sites where there are structures that are not present between these sites. The general conclusion can be made that the gaps are located at sites of structural weakness. If this opinion is accepted, then those hypotheses which suggest that streams sought out weaknesses in the rock during headward erosion are favored.

The following are features that are found at gap sites: (1) dying out of folds along plunge within short distances; (2) narrow outcrop widths of resistant beds because of steep dips; (3) more intense folding locally than nearby; (4) abrupt change in strike owing to kinking along strike; (5) intense overturning of beds and resultant increase in shearing; and (6) cross faulting.

Delaware Water Gap

Delaware Water gap is often cited as the classic water gap in the Appalachian Mountains. Figure 116 portrays its geology. The Delaware River flows through the gap at an altitude of 300 feet. Kittatinny Mountain rises about 1,240 feet above the Delaware River on the New Jersey side, and it is nearly 100 feet lower on the Pennsylvania side. Also, the trend of the ridge crest lies about 700 feet
farther southeast on the Pennsylvania side than in New Jersey. The three members of the Shawangunk Formation match and are aligned at river level. In New Jersey, bedding rises uniformly to the top of the mountain with a dip of about 45° (Figure 118A), but in Pennsylvania the dip decreases about halfway up the mountain to about 25° (Figure 118B). Therefore, there must have been a kink in the rocks that formerly occupied the gap site (Figure 119) and as a consequence, the brittle rocks must have been weakened by fracturing in the flexure zone. The location of the gap is therefore interpreted to have been controlled by the local structure.

The overlying Bloomsburg Red Beds exhibit a series of folds just north of the gap that plunge out to the southwest within a short distance (Figure 116). Because similar tight folding is not seen in the Bloomsburg immediately beyond the gap site, the rocks are presumably more highly sheared here, and resistance to erosion is less than elsewhere along the ridge. Also, the outcrop width of the Shawangunk Formation is narrower at the gap site than to the northeast, where the Cherry Valley anticline and Dunnfield Creek syncline widens the exposure.

The Delaware River curves in a loop mimicking the curve of bedding in the southwest-plunging Cherry Valley anticline (Figure 116). This probably resulted when the river flowed at a higher altitude in a straight line towards the southern part of the gap. At that time, the river was cutting down through the Bloomsburg Red Beds, and when it intersected the more resistant Shawangunk quartzites and conglomerates, it migrated down the plunge of the anticline. Subsequently, the “meander” migrated downstream until it impacted the northwest-dipping rocks of the Shawangunk coming down off the main ridge. The projection of this proposed course of the river to the present top of the Shawangunk places the river at about 900 feet altitude, or about 600 feet above the present level, when it first encountered the Shawangunk.

Two additional gaps will be traversed on this field conference. The gap south of Millbrook (mile 38.3, Day 1) is an irregular gap in the ridge, not quite 200 feet deep. It is located at the site of at least two well-documented cross faults.

Culvers Gap area (STOP 4) is covered largely with glacial deposits and products of mass wasting, so outcrops are not as plentiful as could be desired. The bedrock geology of the gap has been
mapped by Monteverde (1992). It is 450 feet deep with a floor at about 910 feet in altitude, although
drift is fairly thick here and bedrock lies at about 750 to 800 feet. Two features may have aided in the
location of this gap: (1) it is located along the narrowest outcrop width of the Shawangunk Formation,
and (2) it is located immediately southwest of a considerable northeastward broadening of the
Shawangunk outcrop width (Figure 113).

All of the major gaps within the Delaware Water Gap
area and beyond are located at sites of geologic structures
that appear to have weakened the rocks at those sites. These
features are plunging folds, sharp flexures, cross fault, kinks
along strike, and narrow widths of outcrop of the resistant
Shawangunk rocks. There also is evidence that structural
control influenced the development of smaller gaps in ridges
to the north (Epstein, 1966, p. B83-B85). The strong
relationship between the position of the gaps and local
structure suggests that the concept of regional superposition
as applied by Johnson (1931) is invalid. Rather, hypotheses
are favored that maintain that the gaps are located in zones of
structural weakness, where erosion was most effective during
the course of stream competition along the ancestral drainage
divide. While the conclusions presented in this discussion
relate to structural control of gaps, the nature and timing of
stream development cannot be deduced. Of concern is
whether streams are in their original (post-Permian) position, whether they have been captured and
replaced by streams in front of the ridge or by tributaries behind the ridge, and what effect the structure
on both sides of the ridge may have had in the geomorphic evolution.

GLACIAL GEOLOGY

The latest (Wisconsinan) glacial advance into
eastern Pennsylvania and northern New Jersey resulted in
the deposition of a conspicuous terminal moraine which
crosses the Delaware River about 11 miles south of the
gap near Belvidere, NJ (Figure 120). The moraine then
trends northwestward to cross Blue Mountain about five
miles west of Delaware Gap, locally reaching heights of
more than 100 feet in places. As the glacier retreated from
its terminal position north of Blue Mountain, the
meltwater was dammed between the terminal moraine, the
surrounding hills, and the retreating ice front. A series of
stratified sand and gravel deposits were laid down in the
lake that formed, recording the sequential retreat of the
glacier. The lake has been named Lake Sciota, after the
classic delta and varved lake-bottom sediments that are
found there. The lake reached a depth of about 200 feet in
places. Initially, the outlet for the lake was over the
terminal moraine at Saylorsburg and the water flowed
west toward the Lehigh River. As the glacier retreated northeastward past the Delaware River, the
waters drained through the gap and the lake ceased to exist.
A variety of glacial deposits formed in the Delaware Water Gap area, composed of varying proportions of gravel, sand, silt, and clay. On the basis of texture, internal structure, bedding and sorting characteristics, and generally well preserved landforms, the deposits have been subdivided into till (ground, end, and terminal moraine) and stratified drift (delta, glacial-lake-bottom, kame, kame-terrace, and outwash deposits). Below the gap is an outwash terrace, more than 150 feet high on both sides of the river, comprising very coarse gravel with boulders exceeding eight feet long. This deposit may be seen at mileage 6.1 of the Day-1 road log.

Numerous striae, grooves, and roches moutonnee formed by Wisconsinan glacial erosion are found on bedrock surfaces in most parts of the area. Striae trends show that the ice was strongly deflected by underlying bedrock topography. Whereas the average direction of flow of the ice sheet in the immediate Delaware Water Gap area was about S20°W, the base of the ice traveled mostly more southwestward parallel to the valley bottoms and about due south over the ridge top.

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

There has been a long line of researchers of the glacial geology of the area around Delaware Water Gap, including White (1882), Lewis (1884), Salisbury (1902), Leverett (1934), Ward (1934, 1938), Happ (1938), Miller et al. (1939), Mackin (1941), Epstein (1969), Bucek (1971), Crowl (1971), Ridge (1983), Cotter et al. (1986), and Witte (2000).

Leave STOP 1, either by reboarding buses and continuing south about 0.3 mile on US 611—or, better yet, walk the short distance south to STOP 2.

0.3  5.3 Buses turn right into parking area on left side of highway.

STOP 2. COLD AIR CAVE.

Leaders: Mitzi Kaiura and Jack B. Epstein.

Simmering summers in eastern U.S. cities and enticement to vacationers to come to eastern Pennsylvania to admire the beauty of the Delaware Water Gap and enjoy the cool mountain breezes resulted in this area becoming a major resort during the 19th century. Major hotels were built and one, Kittatinny House, was located in the gap. At this point about 1,500 feet south of STOP 1, there is an unusual cave formed by the juxtaposed alignment of large talus boulders derived from conglomerate and sandstone blocks of the Shawangunk Formation (Figure 121). The roof of the cave is formed of large blocks approaching 30 feet long and look as if they are bedrock in place, but that is not the case—they are talus blocks. A large talus floe overlies the cave (Figure 122) and a large cleft in the cliff above (Figure 123) illustrates the potential for generating
large talus blocks during periods of freeze and thaw. The geology of the cave was originally described by Stone (1932) and its history of commercialization was discussed by Snyder (1989) who reported the length of the cave to be 70 feet. Presently, about 30 feet of the cave is accessible to normal-sized individuals. According to Snyder, the cave was discovered about 1870 when very cold temperatures were reported coming from the opening, approaching 30°F.

The cave became a tourist attraction sometime thereafter until about 1952. A building was erected at the entrance (Figure 124) and refreshments were served, cooled by the cold air from the cave.

When Stone (1932) visited the cave in 1931, he reported an air temperature of 38°F.

He wrote:

Cold air coming out on a warm morning makes fog. This cold air, which received its low temperature from the frosts of the previous winter, is stored by Nature in the voids of the rock floe. Cold air tends to settle, and, being held in the floe by the cover of soil and vegetation, it moves slowly down hill through the spaces between the blocks and emerges noticeably in summer at this opening.

Based on this description, it has generally been assumed that the flow of air was constant out of the cave, at least during the summer months. According to this belief, the cold air was residual from the winter months, stored within the talus floe above, and because it was heavier than the ambient warmer air, it constantly flowed down and out of the cave. Presumably, if this hypothesis is correct, the cold air could also be derived from storage within joints in the bedrock forming the cliffs above.

Beginning in September, 2000, temperature, wind speed, and wind direction was measured just inside the cave, at the mouth of the cave, and several tens of feet from the cave entrance. Weather conditions at the time of measurement were also noted. At the outset of the investigation it was recognized that the direction of wind movement varied, sometimes out of the cave, and other times into it. Results of the investigation have not been compiled as yet, but they show that the cave “breathes”, the temperature varies according to the season, and the average temperature is probably the average
subsurface ambient temperature. Readings for temperature, wind speed, wind direction, and weather conditions were recorded and are shown in Table D1 and are summarized in Table D2 in Appendix D.

Leave STOP 2, continuing south on US 611.

0.2  5.5 Small outcrop in the Martinsburg Formation on right. Note the gently northwest dipping (7°) cleavage due to fanning in the A pressure shadow of the syncline defined by the Shawangunk Formation. Bedding here dips 32° NW.

0.2  5.7 Arrow Island Overlook. Arrow Island, a streamlined bar is presently being modified by the Delaware River. This occurs due to a decrease in the velocity of the river as it emerges from the narrow confines of the gap to the north.

0.4  6.1 Very coarse gravel in late Wisconsinan outwash terrace to right exceeds 80 feet in thickness. The gravel was either laid down between the valley’s wall and stagnant ice forming a kame terrace, or it is the eroded remnant of a valley train that formerly filled the Delaware Valley.

0.2  6.3 Rock fence on right is composed of blocks of the Allentown Dolomite with abundant fine sedimentary structures.

0.3  6.6 Cross Slateford Creek. In 1805, the Pennsylvania Slate Company developed a slate quarry south of the water gap near Slateford Creek.

0.1  6.7 National Park Drive on right leads to Slateford Farm, an example of a National Historic Site maintained by the National Park Service.

0.5  7.2 Faulted and overturned beds in the Bushkill Member of the Martinsburg Formation in ravine to right.

0.1  7.3 Flat-lying slate in the Bushkill Member in a 100-foot-long abandoned quarry overlain by 10 feet of glacial drift in ravine to left. A dolomite concretion, characteristic of basal beds of the Martinsburg elsewhere, lies in the bottom of the quarry.

0.4  7.7 Crossing the concealed Portland fault, which juxtaposes Martinsburg against the Jacksonburg Limestone and rocks of the Beekmantown Group where it is exposed.

0.3  8.0 Enter Portland, Pennsylvania.

0.6  8.6 Traffic light. Continue straight.

0.05  8.65 Cross Jacoby Creek. Deglaciation of the Jacoby Creek valley resulted in the formation of several proglacial lakes that became progressively lower as the ice retreated from the northeast-draining valley and lower lake outlets were uncovered (Ridge, 1983).

0.15  8.8 Road goes underneath US 46.

0.1  8.9 Turn right following signs toward I-80.

0.2  9.1 Toll Booth. (No toll leaving Pennsylvania.)

0.2  9.3 Cross Delaware River into New Jersey. Good view of Delaware Water Gap to left.

0.2  9.5 Allentown Dolomite crops out on right. Continue straight.

0.1  9.6 Road signs for I-80 and NJ 94. Continue straight, following signs to NJ 94.

0.2  9.8 Cross the axis of the Ackerman anticline (Drake et al., 1969) in the Allentown Dolomite approximately at point where ramp bears off to I-80 West, but continue straight ahead on NJ 94.

0.2  10.0 Allentown Dolomite on right. We will be traveling up the Paulins Kill Valley, which is underlain by Cambrian and Ordovician carbonate rocks, such as the Allentown Dolomite, Beekmantown Group, and Jacksonburg Limestone. The valley is flanked by hills held up by slates and graywackes of the Martinsburg Formation, lying in fault contact with the carbonates on both sides of the valley.

0.3  10.3 Intersection with Stark Road (village of Warrington). Ice-contact deltas in the lower part of Paulins Kill valley delineate three ice-retreat positions (Witte, 2001; Ridge, 1983).
These deposits lie as much as 100 feet (30 m) above the Paulins Kill, and they form the bulk of meltwater deposits in Paulins Kill valley. These deltas were laid down in small proglacial lakes held in the south-draining valley by older outwash deposits downvalley (Figure 125), and possibly by ice in the Delaware Valley. This stepward style of deglaciation can be traced throughout the Paulins Kill valley. Based on the morphosequence concept of Koteff and Pessl (1981), Ridge (1983) and Witte (1988) have delineated 14 ice-retreatal positions in the valley. The meltwater-terrace deposits that cover parts of the valley floor were formed by meltwater emanating from these up-valley positions. The broad meltwater terraces in the vicinity of Vail and Walnut Valley lie well below the ice-contact deltas. Their lower positions in the valley reflect a lowering of local base level as older outwash down valley became further incised by meltwater draining from younger retreat positions upstream.

1.1 11.4 Pass through concrete tunnel (built 1909) beneath abandoned Delaware, Lackawanna & Western (DL&W) Railroad grade.

0.8 12.2 Road to Mt. Pleasant on left. Pass through village of Hainesburg. A nearly complete skeleton of Cervacles scotti Lydecker was recovered from a bog just southwest of the village.

0.2 12.4 Pass over Yards Creek. For next 1.5 miles, cross over a large ice-contact delta. The surface of the delta is kettled in many places.

0.1 12.5 Cemetery on left in late Wisconsinan outwash (ice-contact delta).

0.7 13.2 Knowlton – Blairstown Township line.

1.6 14.8 Turn left on Walnut Valley Road at sign identifying Yards Creek Pump-Storage Generating Station.

0.2 15.0 Beekmantown Group, upper part dipping northwest on left side of road. Just above this outcrop and around the bend the Jacksonburg Limestone is covered by thin till. Most of the large erratics are the Shawangunk Formation (whitish quartz-pebble conglomerate), with lesser red sandstone boulders of the Bloomsburg Red Beds.

0.9 15.9 Pass Frog Pond Road on left.

0.1 16.0 Cross over Yards Creek. Ramseyburg Member of the Martinsburg Formation exposed in creek bed to right. Here the Ramseyburg has a well-developed southeast dipping cleavage.

0.3 16.3 Many till stones beneath power lines to right. Such stony tracts are common in areas of thick till.

0.5 16.8 Ahead to left on Kittatinny Mountain are prominent cliffs of Shawangunk sandstone and conglomerate.
0.25 17.05 Pass Mount Vernon Road on right and enter Yards Creek Pump-Storage Project just past intersection of Walnut Valley Road and Mt. Vernon Road. Lower reservoir occupies till-dammed valley, the former site of a shallow glacial lake.

0.05 17.1 Stop at guard shack.

0.1 17.2 Visitors’ Center on the left. Continue straight.

0.1 17.3 Dam of lower reservoir to right. Bear left toward Upper Reservoir.

0.2 17.5 Cross Penstock With sharp eyes you can see the unconformable Martinsburg-Shawangunk contact on the cliff to left. Turn right uphill to picnic area.

0.2 17.7 Martinsburg exposed on left.

0.2 17.9 Turn left into picnic area and go to end of road.

0.2 18.1 Park buses. Disembark.


INTRODUCTION

The Yards Creek Pumped Storage Generating Station is owned by Jersey Central Power and Light and Public Service Electric and Gas Companies. It is a large hydrologic storage battery, taking advantage of the height of Kittatinny Mountain and utilizing a reversible pump-turbine. There is a storage reservoir at the top of the mountain and one at the bottom on the south, 737 feet below. A penstock and tunnel connects the two reservoirs (Figure 126). During peak hours, water from the upper reservoir flows through a turbine at the bottom, generating a maximum 400,000 kW of electricity. At night, when there is less demand for power, the generator reverses to become a pump lifting water back to the top of the mountain, thus storing energy for the next day. The operation is about 70 percent efficient, but the low-cost of the surplus off-peak power makes the process economically feasible. The engineering geology of the project area was summarized by Smith (1969).

When the project was initiated in 1963, at the time when the Tocks Island dam was planned on the Delaware River immediately to the north, it was anticipated that the impoundment behind the dam would serve as a additional storage area necessitating an increase in size of the reservoir on top of the mountain. This project was nullified because the dam was never built and was officially deauthorized in 1992.

REGIONAL GEOLOGY

Figure 127 shows the geology in the area. The power plant and lower part of the penstock are in the Ramseyburg Member of the Martinsburg Formation. The penstock continues into a tunnel near the top of the exposed Martinsburg and...
continues through the Minsi and Lizard Creek Members of the Shawangunk Formation and intersects the upper reservoir and the base of the Tammany Member (Figure 127). The penstock has cut through more than 1,000 stratigraphic feet of the Martinsburg, affording an unparalleled opportunity to study the changes in slaty cleavage that takes place as the contact with the more competent overlying quartzites and conglomerates of the Shawangunk are approached. Note that the uppermost member of the Martinsburg, the Pen Argyl, is missing, having been removed along the Taconic unconformity. The Pen Argyl-Ramseyburg contact disappears under the Shawangunk about one mile west of Delaware Water Gap (Epstein, 1973), eight miles southwest of Yards Creek. Erosion of the resistant rocks of the Shawangunk in a syncline-anticline couplet resulted in the offset of Kittatinny Mountain here. Some of the beds are overturned, less so than farther southwest and less so than to the northeast (compare with STOP 6). The Martinsburg is fairly well exposed and its structural characteristics can be seen in different parts of the folds. Of particular interest is the gradual decline and disappearance of cleavage as the unconformable contact with the overlying Shawangunk is approached. That contact lies about 70 feet above where the penstock enters the tunnel. Along with the decrease in cleavage development towards the contact is the gradual “arching” of the cleavage as its dip to the southeast away from the contact changes to a northwestward dip as the contact is approached (Figure 128).

![Geologic map and cross section of the Yards Creek-Tocks Island area, New Jersey.](image)

The lowest beds in the Martinsburg can be seen at the power plant under the transmission lines (Figure 128A). Here, cleavage dips slightly to the southeast and is fairly well developed, although not as well developed as in slate quarries farther from the Shawangunk contact. Higher, just above the picnic area, cleavage is more poorly developed, but is still a noticeable parting in many pelites (Figure 128B),
although is not readily apparent in some nearby beds (Figure 128C). Finally, within a few tens of feet of the contact with the Shawangunk, cleavage is not visibly present, although there may be indications to be seen in thin section.

The effects of the dying out of cleavage can be seen under the microscope (Figure 129). Many papers have described the petrographic characteristics of slaty cleavage (see Epstein and Epstein, 1967). Figure 129D is typical. Cleavage parting is controlled by dark folia that have resulted from the pressure solution of soluble minerals, such as quartz, with a residue of dark carbonaceous material and iron oxides. Interstratal quartz grains between the folia have smooth edges parallel to the cleavage direction due to solution and hackly terminations at right angles, partly due to solution and re-precipitation.
Chlorite and muscovite may grow along the cleavage direction in the pressure fringes of hard pyrite grains (Figure 129D-a). Large chlorite-muscovite grains, a magnitude larger than the groundmass minerals, with mineral cleavage at high angles to the slaty cleavage and commonly cutting across

Figure 129. Photomicrographs showing development of cleavage (short dashed line) at the expense of bedding mineral orientation (long dashed line) in the Martinsburg Formation. Bar scales are all 0.1 mm long. A-C are from outcrops of the Ramseyburg Member of the Martinsburg along the penstock at Yards Creek. D is from an abandoned slate quarry in the Pen Argyl Member of the Martinsburg Formation, 1.5 miles south of the Shawangunk contact at Lehigh Gap, in the abandoned David Williams slate quarry (Epstein et al., 1974, p. 348); it is typical of all well-developed slate throughout the Martinsburg.

A, 2.5 feet below the Shawangunk contact. Muscovite and chlorite grains are parallel to bedding. There are no presolved hackly edges on quartz grains. No cleavage is apparent.

B, 2.5 feet below the Shawangunk contact. Bedding is defined by muscovite grains dipping to the left (northwest in the outcrop). Cleavage is incipient and appears as dark carbonaceous folia with minor muscovite parallel to it.

C, 860 feet below the Shawangunk contact. Bedding is horizontal but cleavage development has reoriented most grains parallel to it. Quartz grains (white) have hackly edges due to pressure solution. Chlorite-muscovite grains that are much larger than the groundmass minerals and whose mineral cleavage is at a high angle to bedding, make an appearance suggesting that they may be prophyroblasts.

D, Well-developed slaty cleavage characteristic of most of the Martinsburg pelites in eastern Pennsylvania and northern New Jersey. Flattening of quartz grains (a), pressure-shadow mineral growth around pyrite grains (b) and growth of muscovite-chlorite along cleavage (c) is typical. Orientation of minerals parallel to bedding is not evident.

(Figure 129D-a). Chlorite and muscovite may grow along the cleavage direction in the pressure fringes of hard pyrite grains (Figure 129D-b). Large chlorite-muscovite grains, a magnitude larger than the groundmass minerals, with mineral cleavage at high angles to the slaty cleavage and commonly cutting across
cleavage folia, appear to be porphyroblasts (Figure 129D-c). As the Shawangunk contact is approached and cleavage dies out, these features gradually disappear. Quartz grains have their original shape. Pressure fringe growth around pyrite grains is lacking. Large chlorite-muscovite grains are missing. All this occurred after the folding of the Shawangunk, and by inference, after all of the overlying rocks which were probably about 20,000 feet thick. Clearly, the process of cleavage development occurred under conditions of heat and pressure (metamorphism), and based on conodont geothermometry, in the neighborhood of 300°C (Epstein, 1974).

As discussed at STOP 1, the arching of cleavage as the trough of a syncline in the competent rocks of the overlying Shawangunk is approached, which is accompanied by a dying of the cleavage, can be ascribed to a pressure-shadow mechanism. This can be seen at Delaware Water Gap (STOP 1), here at STOP 3, and elsewhere, such as at Lehigh Gap. Small-scale outcrop versions of this phenomenon, as well as microscopic, are numerous. The significant point is that it occurs at different levels within the Martinsburg (within the middle of the Ramseyburg Member here, at the top of that member at Delaware Water Gap, and high in the youngest Pen Argyl Member at Lehigh Gap). Thus, the controlling factor is the overlying Shawangunk, which had to be present at the time the cleavage developed. Since the Shawangunk is post-Taconic in age, the cleavage must be post-Taconic, i.e., Alleghanian (there is little evidence for significant Acadia (Devonian) deformation in this part of the Appalachians).

Nowhere in Eastern Pennsylvania and New Jersey is the contact between the Shawangunk and Martinsburg in the overturned limb of a syncline exposed to demonstrate the complementary arching of the cleavage as shown in Figure 117A, STOP 1. The contact between overturned beds of both formations was temporarily exposed, however, during construction of an additional tunnel for the northeast extension of the Pennsylvania Turnpike in 1989, two miles southwest of Lehigh Gap and 31 miles southwest of Delaware Water Gap (Epstein and Buis, 1991). At the overturned contact the Martinsburg dips 35° SE and the Shawangunk is 10° steeper. Cleavage dips 5° less than bedding in the Martinsburg, conforming to the model shown in Figure 117A, Stop 1.
of the same fold in which, up-plunge at Lehigh Gap, the contact is at the exact trough of the fold and cleavage dies out within 200 feet of the contact (Epstein and Epstein, 1969).

Yards Creek has one of the four known exposures of the Taconic unconformity between the Martinsburg and Shawangunk Formations. This is the best. The contact is about 70 feet above the top of the penstock as it enters the tunnel. The angle of discordance between the two formations is hardly measurable, and probably no more than a couple of degrees. An attempt to get bedding attitudes readings in both units at the contact resulted in N20°E/24°NW for the Martinsburg and N26°E/25°NW for the Shawangunk. The basal Shawangunk is sandstone and conglomerate with pebbles as much as 3/4 inch long with slightly irregular bedding planes. A bed in the Martinsburg rises to the southwest along the contact and is cut out within a few tens of feet under the lowest bed in the Shawangunk (Figure 130A).

At the contact, between the pelites of the Martinsburg and the feldspathic sandstone of the Shawangunk, there is a 1-2" thick yellowish-gray (5Y 8/1) to light-greenish-gray (5GY 7/1) clay. The color is in marked contrast between the dark-gray pelite of the Martinsburg and the grayish-orange weathered color of the basal Shawangunk. It is undoubtedly the result of leaching by ground water, which is common at this type of contact. However, the clay contains sheared, fragmented, contorted and disoriented clasts of pelite, showing that it is a fault gouge. The gouge is damp and fragile and difficult to photograph. Surfaces carefully cut with a sharp razor blade clearly show the brecciated shale fragments. Movement during folding might be normally expected with overriding beds moving up towards the crest of adjacent anticlines. However, elsewhere at the contact, such as at Lehigh Gap (Epstein and Epstein, 1967, 1969), slickensides show that movement has been in an opposite direction, to the northwest, conforming to the regional northwest translation of bedding-plane faults and vergence of folds in overlying Silurian and Devonian rocks.

The angular discordance at the unconformity, attributable to Taconic deformation of the Martinsburg prior to deposition of the Shawangunk, is only a few degrees here. Along the entire length of Kittatinny Mountain in northern New Jersey and into Shawangunk Mountain in southeastern New York (as least to near Ellenville, NY), and in the opposite direction to the southwest to Hawk Mountain in Pennsylvania, the angular discordance does not exceed 15 degrees, showing that this is a zone of low-amplitude Taconic folding. Figure 10 (p. 15) shows the regional picture.
Group, lower part. Immediately south lies a southeast-dipping thrust fault that traverses almost the entire Paulins Kill carbonate valley.

0.2 23.45 Turn left off NJ 94 onto Main Street, Blairstown. The town is built on a meltwater-stream terrace and slightly higher deltatic outwash.

0.25 23.7 Turn left onto Warren County 602 in Blairstown and go up hill.

0.2 23.9 Pass Blair Lake on left.

0.1 24.0 Cross over Blair Creek. Follow Warren County 602 (Blairstown to Millbrook Road) up Blair Creek valley.

0.1 24.1 Exposure of the Allentown Dolomite on the left. The trilobite faunas at this site have been extensively studied, beginning with Weller (1903) and later Howell (1945). The faunas place this section of the Allentown within the upper Trempealeauan of Late Cambrian time.

0.2 24.3 Small exposure of the Allentown Dolomite on the left. The rocks here dip towards the northwest. Continuing along the route will traverse the section oblique and crosses younger units.

0.5 24.8 Conklin lime kiln. The Allentown Dolomite ridge forms the backdrop of the lime kiln.

0.5 25.3 Late Wisconsinan outwash on right, covering parts of the floor of Blairs Creek valley.

0.3 25.6 Intersection of Millbrook and Spring Valley Roads at Hardwick Center. Bear left and continue up hill. Pass through faint morainal topography (Franklin Grove recessional moraine) and begin climb out of carbonate valley.

0.1 25.7 Small weathered crop of the Beekmantown Group, upper part. This dolomite is characteristically coarse grained and highly fetid. Just above this exposure there is a slight flattening of the topography where the Jacksonburg Limestone traverses. Outcrops were found during foundation construction of several new homes to the west. The contact between these two units marks the Beekmantown unconformity that can be traced southward and upsection into the Knox unconformity in the southern Appalachians. It marks the end of Lower Paleozoic carbonate margin and the beginning of the Taconic Orogeny and subsequent foredeep infilled by the shale, now slate, and graywacke of the Martinsburg Formation.

1.3 27.0 Entrance to Blairstown Center of Princeton University on right.

0.4 27.4 Road to Camp Mason on right. Large stone rows of glacially-shaped boulders and cobbles. Most of the boulders are quartzite and quartz-pebble conglomerate, quarried and plucked from the Shawangunk Formation (the same rock that holds up Kittatinny Mountain).

0.8 28.2 Cross over Jacksonburg Creek, well-formed morainal topography (Franklin Grove moraine) to the left. Note boulder-lag terrace in valley along creek.

0.2 28.4 Well-formed morainal topography on left (Franklin Grove moraine). Climb southeast face of Kittatinny Mountain. Ahead is a good view of the mountain’s steep, southeast-facing escarpment. The lower part of slope is covered with thick talus.

0.7 29.1 Enter Delaware Water Gap National Recreation Area (DEWA). Pass shale-chip rubble pit on left. Material chiefly derived from frost-shattered Ramseyburg Member of the Martinsburg Formation, which here lies in the south limb of a syncline. Cleavage is poorly developed to absent in the thin-bedded shale and graywacke. The Taconic unconformity (contact between the Martinsburg and overlying Shawangunk) is 100 feet above road level. The Martinsburg was quarried here for road material.

0.4 29.5 The last exposure of the Ramseyburg immediately below the Taconic unconformity on the right (north). A 6-foot-thick graywacke bed in the Ramseyburg Member of the
Martinsburg dips gently to the southwest in the trough of a syncline. As at STOP 3, this graywacke parallels the form of folds in the overlying Shawangunk and disappears under the Shawangunk to the northeast.

0.3  29.8 Cross over Appalachian Trail and descend northwest flank of Kittatinny Mountain. Two cross faults were mapped in steeply dipping beds of the Shawangunk immediately to the east. Good view of the Pocono Plateau straight ahead on a clear day.

0.1  29.9 First exposure of the Shawangunk Formation above the Taconic unconformity. This crop exposes a quartz pebble conglomerate that dips westward.

0.9  30.8 Cross over Van Campens Brook and enter Millbrook Village (restoration of a 19th century village). Millbrook began to develop in 1832 when Abram Garis built a gristmill on Van Campens Brook, followed by a Methodist Church in 1839. By 1840, a general store and blacksmith shop were added, as was a post office in 1848. In the 1850’s, a wheelwright shop, a shoemaker shop, hotel, and a tavern were constructed. The village grew out of a necessity to serve the local farms—and, later, travelers who were using the nearby ferries.

0.1  30.9 Intersection with Old Mine Road. (A road log along the historic Old Mine Road from the Delaware Water Gap to this point is Appendix A of this guidebook.) Turn right onto Millbrook – Flatbrook Road.

0.6  31.5 Gated park road on left before pond.

0.1  31.6 Striated Bloomsburg pavement (S28°W ) near low earthen dam. Shows cross-strike ice flow.

0.2  31.8 Bloomsburg outcrop on right.

0.25  32.05 Warren County – Sussex County boundary. Several striated outcrops of Bloomsburg Red Beds (S10° to 20°W) along road on right.

0.35  32.4 Pass intersection (on right) with Blue Mountain Lakes Road. The House at this intersection, now a general store operated by the Park Service, was built sometime between 1840 and 1880. It was purchased by Samuel Garris in 1904 and operated as a hunting lodge, called the Flatbrookville Hotel. In 1926, Andrew and Nelda Salama (White Russians who fled the Russian Revolution) purchased the property and ran the place as a communal summer house, named Salamovka, for New York artists and theatrical friends (Clemensen, 1996).

0.5  32.9 Pass access road to Riverbend Campgrounds on left. (The campgrounds was one of the stops on the pre-conference canoe trip. See the guidebook for this trip for a discussion of late Wisconsinan deglaciation, postglacial terraces, local glacial stratigraphy exposed along bluff, formation of Wallpack Bend [consequent drainage, superimposed meander, structural control ?].)

0.1  33.0 Cross Flat Brook. At stop sign turn right onto Wallpack-Flat Brook Road (Sussex County 615 North). (Note old iron truss bridge to right.) Wallpack Valley is a strike-controlled valley underlain by the Poxono Island Formation, a weak dolomite with sandy interbeds, and the Bossardville Limestone. Across the road is a small outcrop of Bossardsville Limestone that dips southeast. This orientation defines a small, tight syncline. The Poxono Island Formation underlies the Flat Brook and is thought to be a detachment zone by some authors (Epstein) that separates two different rock units, the siltstone and sandstones of the Bloomsburg Red Beds below from the Silurian and Devonian, dominantly carbonate rocks, above. Unfortunately, the Poxono Island is very rarely exposed in New Jersey. Only a few feet have been exposed in three locations east of the Delaware.
Just downstream from the intersection is the village of Flatbrookville. By 1872, it contained two blacksmith shops, wheelwright shop, three general stores, grist mill, saw mill, a cooper shop, hotel, Dutch Reformed Church, and about 20 houses. The village thrived due to its proximity to Decker’s Ferry, a well-traveled route across the river, located about one-half mile downstream at Wallpack Bend (Clemensen, 1996).

0.3 33.3 Cemented, collapsed cobble-pebble gravel outwash overlying Bossardville Limestone on left.

0.4 33.7 Pass USGS Gaging Station on right.

0.45 34.15 Pass small, abandoned, sand and gravel pit on right. Six feet of planar-bedded, matrix supported cobble-pebble-gravel overlying eastward dipping foreset strata of cobble-pebble-gravel and pebble gravel. The outwash here is possibly the remnant of an ice-contact, valley-fill delta.

0.35 34.5 On left side of road, fossiliferous limestone and sandy limestone of the Shawnee Island and Peters Valley Members of the Coeymans Formation are exposed (measured section 10 of Epstein et al., 1967).*

*This locality has both personal and geologic significance to one of us.

Epstein reflects: In 1962, my geologic partner, Anita, and I visited this outcrop in search of ostracodes in anticipation that it would become one of the research localities for her paleontologic studies in Silurian and Devonian rocks in eastern Pennsylvania, leading to a PhD dissertation at Ohio State University. At this time I was mapping the area to the west for the USGS. I broke off a piece of the limestone (which we later named the Shawnee Island Member of the Coeymans Formation). A tiny black shape in the rock prompted me to say, “Anita, look—a conodont.”

“Impossible, you dummy,” said she (her typical vernacular). “Conodonts are too small to be seen with the naked eye!”

“It sure looks like a conodont,” said I, “a big black one.”

It indeed was a conodont, a large Icriodus woschmidti, typical of the Coeymans.

Anita subsequently changed her dissertation from ostracodes to conodonts because conodonts are excessively better for zonation, have a worldwide distribution, and are found in a variety of rock types.

For the next year or two we pondered as to why these conodonts were pitch black, whereas those in the mid-continent in Ohio were pristine amber. Reasoning that the color may reflect depth and duration of burial, and initially using poorly calibrated ovens, we determined that conodonts are marvelous geothermometers. With increasing temperature and over time, they changed from their amber color, through shades of brown, to black as their organic matter was altered. The temperature range spanned the temperature of hydrocarbon generation and destruction, encroaching on lower green schist metamorphic grade (and beyond as further
alteration was noted in metamorphic rocks).

During a phone discussion regarding environmental-geologic matters, I mentioned our analytical technique and results to a USGS geologist interested in oil and gas exploration. He recognized the importance of our findings and the initial disclosure of the discovery appeared in a Geological Society of America abstract (Epstein, Epstein, and Harris, 1964). We collaborated to produce a comprehensive report (Epstein, Epstein, and Harris, 1967) describing the technique in detail. This was followed by a series of papers and maps as the technique was further applied and as one Epstein became unbetrothed and rebetrothed (e.g., Harris, Harris, and Epstein, 1978). The technique has since become an indispensable tool for oil and gas exploration and for other applications.

Three years after the initial conodont discovery, this locality was included in the Delaware Water Gap National Recreation Area. The spot is the birthplace of a geothermometer technique that has saved the petroleum industry countless millions of dollars in exploration costs by indicating areas whose temperature regime is promising for production and indicating those areas that may be worthless for such exploration.

So the oil companies are making millions—and Jack is still running around banging away at rocks. Maybe Anita was right!

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Figure 132. Mudcrack polygons on bedding surfaces of the Bossardville Limestone at Haneys Mills. These and other structures indicate subaerial exposure on tidal flats during Silurian time. Note that the polygons have been compressed during folding. The average length/width ratio of the polygons is about 1.5/1.
Bossardsville Limestone exposure displaying several small folds, and coarse, crystalline calcite veins.

Pass Wallpack Inn on right (“We feed the deer and the people too.”) What they really mean is this: “We feed the deer to fatten them up, so that we may prepare a mighty fine venison sauerbraten.”

DEWA ranger station on left. The basal section of the Bossardsville Limestone occurs on the west side of the road. Henry Herpers of the New Jersey Geological Survey observed the upper few feet of the Poxono Island Formation during road repair in the 1950’s.

Wallpack Valley Environmental Education Center on right. More Bossardsville Limestone occurs along the left side of the road.

DEWA ranger station on left. The basal section of the Bossardsville Limestone occurs on the west side of the road. Henry Herpers of the New Jersey Geological Survey observed the upper few feet of the Poxono Island Formation during road repair in the 1950’s.

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DEWA ranger station on left. The basal section of the Bossardsville Limestone occurs on the west side of the road. Henry Herpers of the New Jersey Geological Survey observed the upper few feet of the Poxono Island Formation during road repair in the 1950’s.
Enter Culvers Gap. Upper Shore Road on left. Continue south on US 206 and cross over the Appalachian Trail, which is located at the intersection of Upper Shore Road and US 206.

Pass drop-off point for STOP 4 on left. We will pass through the gap and turn the buses around near Culvers Lake. By doing this trip participants will not have to cross the highway (where there’s a good chance someone will wind up as road kill), and we’ll have a good view of Culvers Cap from the Kittatinny Valley side.

Culvers Lake on left.

Turn around at McKeowns-at-the-Lake restaurant on left.

Just past Lower Shore Road, on the right, is the drop-off point for STOP 4. Pull off highway into small parking area. Buses will continue to the Culvers Gap Parking Lot on the west side of the gap. STOP 4 will consist of 3 parts. For part “a”, we will examine the materials that make up the moraine. Part “b” will consist of a short walk (5 min) to the outer margin of the moraine where we will gather for the part “c” on the crest of the frontal ridge of the moraine. Here we will discuss the morphology of the moraine, its formation (push moraine, dump moraine, stagnation moraine ?), and historical context. Due to the large number of trip participants, proximity of the drop-off point to the highway, and limited space at STOP 4a, there will be no formal discussion until we meet again at STOP 4c. For those who wish to forgo the first two parts, the discussion area may be accessed from the Culvers Gap Parking Lot. See directions below.

Directions for buses to Culvers Gap Parking Lot:

Turn right onto Upper Shore Road (Sussex County 636).

Turn left onto Sunrise Mountain Road, followed by a quick left into Culvers Gap Parking Lot. Park buses and wait for weary and very hungry field trippers. For those that want to find STOP 4c, return to intersection of Sunrise Mountain Road and Upper Shore Road. Cross Upper Shore Road and climb small ridge. Await bevy of geologists.

Leader: Ron W. Witte.

Location

STOP 4 is located in the Culvers Gap, NJ-PA, 7.5-minute quadrangle in the Culvers Gap section of Stokes State Forest, Sussex County, New Jersey (Figure 133). At one time, a well-worn trail through the gap linked the many Lenapi hunting and fishing camps around Culvers Lake, Lake Owassa, and nearby Swartswood Lake with the Delaware River.

Geologic Setting

Culvers Gap is a preglacial wind gap (Figure 134) that lies in the glaciated section of the Valley and Ridge physiographic province. It forms a prominent pass through Kittatinny Mountain, linking Kittatinny Valley to the east with Wallpack and Minisink Valleys to the west. The mountain is held up by the Shawangunk Formation, a tough and very resistant quartzite and quartz-pebble conglomerate of Silurian age. The floor of the gap is covered by the Ogdensburg-Culvers Gap moraine, a recessional moraine of late Wisconsinan age, laid down at the margin of the Kittatinny Valley lobe. Based on nearby well records, the gap’s bedrock floor lies between 750 and 800 feet (230 and 245 m) above MSL. Salisbury (1902, p. 350) appears to have been the first to use the term Ogdensburg-Culvers Gap moraine in describing a discontinuous morainal belt that traced a course westward through Kittatinny Valley from Ogdensburg to Culvers Gap. Salisbury also noted that morainal topography in Minisink Valley near Fisher School House (located just north of Dingmans Ferry) and in Wallpack Valley near Layton, might have been coeval to the feature at Culvers Gap. Minard (1961) traced the moraine over Kittatinny Mountain, and Witte (1997a) further refined its course and placed it within the morphostratigraphic framework that he developed for the Kittatinny and Minisink Valley ice lobes. (See Witte, this guidebook, p. 81, for more information on the distribution of end moraines, their history, and how they are formed.)

End Moraines

End moraines in northwestern New Jersey form conspicuous, cross-valley ridges that mark former, stable, ice-marginal positions of the Laurentide ice sheet. These features are of late Wisconsinan age, consist chiefly of till, have a varied morphology, and follow looping courses through the Kittatinny and Minisink Valleys. Their courses show that the margins of the Kittatinny and Minisink Valley lobes were distinctly lobate at both a regional and local scale. End moraines include
End moraines consist of poorly consolidated, bouldery, silty-sandy to sandy till with minor beds of water-laid sand, silt, and gravel (Figure 135). This material is distinctly different from the more compact, and less stony ground moraine or till that lays near the moraine. Additionally, stratified drift is not a major constituent, even in places where the moraine crosses river valleys or former glacial lake basins. Outcrops of morainal materials are rare due to the difficulty of digging the bouldery drift and more importantly its lack of economic value. The best outcrops are places where the moraine was removed to expose economic deposits of sand and gravel.

STOP 4a: Composition of the Moraine
At STOP 4a (Figure 133) we’ll examine the materials that make up the end moraine in Culvers Gap. The outcrop in the west wall of the borrow pit exposes about 7 feet of till. The parent material consists of poorly sorted, 10YR4/4 (dark-yellowish brown), slightly to moderately compact, silty sand. As to gravel content, 7-10 percent of the clasts are greater than 1 inch in size, and 25-30 percent are less than 1 inch. The till’s matrix has a granular to slightly prismatic structure, and contains indistinct layers and lenses (less than 2 inches thick) of coarse sand and very fine gravel. Most of the clasts are subrounded, although subangular and rounded shapes are present. In places, some layers are crudely graded, their bases marked by a concentration of larger clasts. The gravelly and granular texture of the morainal till, its indistinct and graded layering and sandy interbeds suggest that this material consists of flow till and ablation till. The character of this outcrop and several others observed by the author suggest that morainal materials have had a complex history of sedimentation related to the release of ice-entrained debris by melting at the glacier’s margin.

The high percentage of clasts and fragments derived from the Martinsburg Formation (mineralogy of the morainal till is listed below) and the course of the moraine (Figure 135), clearly shows that the debris here was derived from sources in Kittatinny Valley. The Kittatinny Mountain rocks (Shawangunk Formation and Bloomsburg Red Beds) also make up a large percentage of the moraine. These lithotypes are found in glacial drift in Kittatinny Valley, especially along especially the western side of the valley. The reason for this is that during the late Wisconsinan maximum, ice flow was directed across Kittatinny Mountain into Kittatinny Valley (Witte, 1997a). Their abundance in the
End moraine in Culvers Gap shows that the moraine is largely made up of reworked till.

**Lithology and mineralogy of Ogdensburg-Culvers Gap moraine in Culvers Gap**

*(sample depth = 77 inches)*

*(Granules and sand mineralogy based on visual estimates provided by F.L. Muller, New Jersey Geological Survey)*

Texture: 20% gravel, 35% sand, 45% silt and clay.

Pebbles (1-3 inches): 3% dolostone (Kittatinny Supergroup), 41% slate and graywacke (Martinsburg Fm.), 25% lithic, quartz sandstone (Martinsburg Fm.), 29% quartzite and quartz-pebble conglomerate (Shawangunk Fm.), 2% red sandstone (Bloomsburg Red Beds), 1% vein quartz.

Granules: 20% red sandstone, 35 % gray sandstone - quartzite, 45 % gray siltstone and shale.

Sand (light fraction): 5% red sandstone, 15% gray siltstone and shale, 10% medium- to fine-grained yellow-stained sandstone-quartzite, 15% light gray to white shale, 40% quartz, 15% clay and silt aggregates.

Sand (heavies): 5% goethitic ironstone, 5% muscovite, 1% ilmenite, 1% garnet, 2% magnetite, 5% aluminosilicates, 1% rutile, 7% zircon, 1% pyroboles, 60% quartz, 5% clay and silt aggregates.

**Questions:** 1) If this material was largely deposited as debris flows, is it really till? 2) What is the source of the morainal sediment, and does their exist a correlation between the thickness and extent of till immediately northward (up-ice) from the moraines and their size? 3) How is debris concentrated at the glacier’s terminus?

**STOP 4b: Morphology of the Moraine**

Head to STOP 4b (Figure 133) by following the trail marked with blaze-orange ribbons (northwest direction). This short walk should take about five to ten minutes. The traverse will take us from the moraine’s inner margin to its outer margin, a climb of about 80 feet (25 m). The main purpose here is to observe the moraines morphology and note how it changes from knob and kettle to ridge and swale. Unfortunately, due to the time of year vegetation will mask some of the morainal landforms (you should have been here earlier this year).

The first part of the traverse will take us through a stand of hemlocks and across the inner morainal margin. Here knob-and-kettle topography is well formed and many of the kettles contain year-round water. The overall appearance of the topography here is one of collapse. The second part of the traverse will take us through the outer morainal margin. It starts approximately, where the arboreal cover changes to a mixture of chiefly white, red and chestnut oak, hickory, and silver birch. Here knob-and-kettle gives way to ridge-and-swale topography. Ridge crests in this area follow curvilinear traces that parallel the course of the moraine. Swales here are elongated and typically parallel the adjacent ridges. Continue along trail and gather on the crest of a large ridge at STOP 4b.

STOP 4b lies on the crest of a large morainal ridge located along the front (outer margin of the moraine). Ridge and swale topography is well formed here. Behind the ridge (eastward), the moraine quickly drops off about eighty feet, with topography changing to knob and kettle.

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 topographic setting that to the casual observer 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, end moraines consist of several types of topographic elements that can be mapped, and characterized (see Witte, this guidebook, p. 81, more detailed discussion on their morphology and origin).

A main topographic feature of end moraines in this part of New Jersey are their sets of nested ridges that parallel the moraine’s course (Figure 136).

MPR’s, like the one we standing on, 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 (610 m) long. However, most are less than 500 feet (152 m) long. Many appear to have been formerly continuous, but they may have been disconnected by collapse during the melting of buried ice. Ridge crests follow straight to slightly curved traces that parallel the outer border of the moraine. In places, they form nested sets that exhibit a remarkable degree of parallelism, suggesting they were built at several ice-margin positions. The topographic profiles of these moraine-parallel ridges (MPR’s) 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. They are either push ridges, formed where the advancing ice had bulldozed ice-marginal sediment; ramps, formed by stacked sets of imbricate thrusts; or colluvial ramparts, laid down where the glacier margin remained stationary, shedding an apron of debris off its terminus.

Morainal knolls consist of low, rounded 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 are typically 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 terminus of the glacier. Over time, the icy substrate melted, letting down its sediment load on the land; the thicker areas of sediment now form the higher parts of the moraine.

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. 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 margin of the glacier.

Typically, the innermost parts of the morainal segments have fewer ridges, fewer elongated depressions, and exhibit knob-and-kettle rather than ridge-and-trough topography. The morphology expressed by the Augusta moraine (Figure 136) 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 supply most of the sediment that makes up the moraine rather than nearby glacially eroded bedrock.

**Questions:**
1) How are moraine-parallel ridges formed? Are they formed by active ice, or are they a product of mass wasting at the terminus of a stagnant glacier margin?  
2) Why are the moraines larger and more continuous on Kittatinny Mountain where they lie next to thick till?  
3) Do end moraines represent places where the glacier margin remained in a relatively stable position for hundreds of years?
Lake.

0.2 53.1 Cross Stony Brook. Bloomsburg outcrop in streambed below bridge on left.
0.1 53.2 Stony Lake Parking Lot. Disembark. After picking up your lunches and drinks, head over to the picnic area in the nearby hemlock grove or out to the shore of the lake.

LUNCH STOP. STONY LAKE AND STOKES STATE FOREST: A POTPOURRI OF LOCAL HISTORICAL TIDBITS.
Hosts: Ron W. Witte and Donald H. Monteverde.

Welcome to Stony Lake Day-Use Area, Stokes State Forest, New Jersey (Figure 137). After picking up your lunches and drinks, head over to the picnic area in the nearby hemlock grove or out to the shore of the lake. After lunch, take a few minutes to look over the sites (sights?, whatever) described below. (Information on the sawmill, silver mine, and Stokes State Forest is from A Guide to Stokes State Forest, 39 p., Division of Parks and Forestry, New Jersey.)

PARK HISTORY

In the early 1800’s, William Snook operated a sawmill on Stony Brook, downstream from the bog that is now Stony Lake. Early on, the mill cut a steady quantity of railroad ties, supplying the Railroad Barons with the timber they needed to expand their empires. In 1872, a new mill was built, replacing the old up and down style saw with a circular one. The mill, a family-run business that spanned seven generations, continued operation until 1950.

In 1907, New Jersey Forest and Park Commission purchased 5432 acres of land in northwestern New Jersey and named it Stokes State Forest in honor of Governor Edward Stokes. He had donated the first 500 acres. Subsequent land acquisitions have increased the size of the park to 15,482 acres. It borders DEWA on the south and High Point State Park to the north, collectively forming a large tract of public land that covers most of Kittatinny Mountain. The state forest is maintained for public recreation, and wildlife and water conservation. In its early days, timber production was also overseen. In the 1930’s, two Civilian Conservation Corps (C.C.C.) work camps were set up in the park to improve trails and roads, cut lumber, plant seedlings, and make overall improvements. One such improvement was the creation of Stony Lake, formed after a 12-foot high, concrete arch dam was built across Stony Brook.

MASTODON DISCOVERY

A fortuitous discovery in 1939 by a C.C.C. worker led to the recovery of several mastodon...
molars, and part of a jaw bone and front leg. The remains were uncovered while excavating the north bank of Shotwell Pond, located about one mile west of Stony Lake. Initially, the find was kept a secret out of fear of “being ridiculed by those who might consider the bones to be those of domestic farm animals.” Only after the specimens were taken to the American Museum of Natural History in New York and positively identified as *Mammut americanus*, did the C.C.C. worker acknowledge his find. They are on display at the Stokes State Forest Visitors Center.

**LOCAL GEOLOGY**

The Stony Lake dam was constructed on the Bloomsburg Red Beds on the former site of a small waterfall on Stony Brook. An abandoned plunge pool and narrow-rock channel may be observed just below the dam. The plunge pool, notching of the falls and formation of lower cascades were probably late Wisconsinan age, cut by meltwater during deglaciation about 18,000 years ago. A cross sectional Bloomsburg cut exists just below the dam on the stream’s southwest bank. The medium-bedded siltstone and fine sandstone, often displaying fining-upwards cycles, offers a good look of the Bloomsburg’s regional lithology both here and regionally (Figure 138; Prave et al., 1989). Finer grained beds have a much stronger developed cleavage (Figure 138). Variable cleavage development can be used to highlight the fining upward cycles. The well-developed southeast-dipping regional cleavage is well illustrated at this stream cut. More exposure immediately below the dam just northwest of the stream exhibits a change to a gentler northwest bedding dip. Very gentle southeast-dipping bedding outlines an opposing fold limb of an anticline. The dam blocks further tracing of this fold. Quartz slickenlines on the northwest-dipping bedding surface suggest tops to the southeast. This movement direction indicates flexural-slip movement, similar to trends seen elsewhere in the outcrop belt.

The hummocky area below the hemlock grove is underlain by a gravelly, poorly compacted, light-brown to yellowish-brown, silty sandy till, the common surface till for this part of the park. These small (typically less than a few acres) hummocky areas are found throughout the park. In places topography is rock-controlled, in other places topography is constructional, the drift possibly an accumulation of ablation till.

**SNOOK SILVER MINE**

According to state forest history, John Snook (son of William Snook) found traces of silver in a veined Bloomsburg outcrop. So in 1875 he single handedly began to develop the small prospect. He hammered, drilled and blasted until eventually a rectangular shaft 4 feet wide by 10 feet long and 32 feet deep was excavated (Figure 139). For his own safety he decided to enlist his own small children as the blasting crew. They entered the shaft and ignited the blasting fuse. We have no knowledge if anyone was injured on the job.

![Figure 138. Photograph looking westward across small stream just north of the Stony Lake dam. Several fining-upwards cycles of Bloomsburg Red Beds are evident in the photograph. Variable cleavage development can be used to highlight the fining-upward cycles. The upper clayey siltstone beds portray a stronger developed cleavage than the lower sandstone layers.](image-url)
As the prospect was not profitable for full time employment, Snook often flooded the shaft when he was not actively working. History suggests the flooding was intentional to deter trespassers from working the prospect in Snook’s absence. A hand powered pitcher pump allowed Snook to drain the shaft upon his return. There is a question on Snook’s flooding of the shaft, as the opening is only several feet above a small stream. During spring runoff the opening would occasionally flood. Together with a wooden A-frame structure that suspended a bucket with a pulley further excavation could continue. State forest history does not list the total profit from the prospect except that Snook received $75.00 per ton of ore. Snook was known to be somewhat of a braggadocio. There is a belief that the price of the ore was used by Snook to raise the value of his real estate (W. Foley, personal communication). There is no note of this mine in New Jersey Geological Survey historical records. Figure 139 shows all that remains of Snook’s silver mine.

Figure 139. The gated opening to Snook’s Silver Mine. Bedrock is exposed at the surface and down for approximately 15 feet. The opening lies only several feet above a small intermittent stream just off the photograph in the foreground. View is looking northeastward.

Leave LUNCH STOP, following Kittle Road back to Coursen Road.

0.3 53.5 Turn left onto Coursen Road, heading back to US 206.
2.0 55.5 Turn right onto US 206 North.
1.65 57.15 Cross Big Flat Brook. On the west side of the river is the remnant of a valley-train laid down from the Culvers Gap margin (located about 1 mile upstream). Most of this material has been eroded by later meltwater, and the postglacial Big Flat Brook. Climb out of the Big Flat Brook valley. The upland topography is formed by thick to thin till overlying large strike-controlled ridges held up by the Bloomsburg Red Beds.
1.65 58.8 Cross Dingmans Ferry moraine and descend into Little Flat Brook valley.
0.9 59.7 Cross Little Flat Brook. The lowland here is a strike valley that principally follows the very weak Poxono Island Formation (Monteverde, 1992). The valley floor is covered by collapsed and dissected outwash that was laid in a small proglacial lake dammed by the Dingmans Ferry moraine (Witte, 1997a). Some poorly drained areas appear to outline former stagnant ice blocks, an indication of stagnant ice at the margin of the Minisink Valley lobe. The low ridge on your left (west) is Wallpack Ridge. On your right (east) is the broad alluvial plain of Little Flat Brook. The low terraces that rise above the flood plain are stream-terrace deposits.
1.0 60.7 Pass Hainesville General Store on the left.
1.2 61.9 Cross over drainage divide between Little Flat Brook and White Brook. The surrounding uplands chiefly consist of undulating strike-controlled bedrock ridges covered with thin till.
1.2 63.1 Intersection of US 206 and Clove Road (Sussex County 653). Bear right at Y intersection. The level plain here is underlain by outwash laid down at the margin of the Minisink Valley lobe (Witte, 1997b). The low relief ridge directly ahead (north) is the
Montague recessional moraine.

0.2 63.3 Cross over Montague recessional moraine.

0.35 63.65 Intersection with the Deckertown Turnpike (Sussex County 650).

0.9 64.55 Cross Shimers Brook. Crop of New Scotland in stream.

0.65 65.2 Entrance to High Point Country Club on right. The area around the country club is underlain by deltaic outwash. Lakes and wetlands occupy ice-block depressions. The New Scotland Formation crops out across the street behind the Montague Municipal Building.

0.3 65.5 Turn right into small shopping center (Montague Mini-Mall) and park behind buildings near self-storage area.

STOP 5. MONTAGUE MINI-MALL FOSSIL SITE: FLANK OF A CORALLINE BIOHERM IN THE COEYMANS FORMATION.
Leader: Donald H. Monteverde.

This site is located within the Milford, PA-NJ, 7 1/2-minute quadrangle behind the Montague Mini-Mall on Clove Road (County Route 653, Montague Township, Sussex County, New Jersey (Figure 140). Zitrone Construction Co. owns the storage facility and permission should be obtained before entering the site. Currently, New Jersey Green Acres Program (NJGAP) is in discussion with Zitrone about purchasing the limestone outcrop. This is part of a new program to purchase geologically significant plots so they will not be lost to development. Staff of the New Jersey Division of Parks and Forestry has a long-term goal to create “Geologic Heritage Trail” that locates and interprets this and other geologically significant sites for the public. When NJGAP purchases geological sites, they become public lands under the laws of New Jersey. The state’s initiatives in such preservation are in line with a very active movement throughout Europe to preserve significant geological sites, headed by both government agencies and private organizations. A broad range of geological sites have been preserved as geological parks and/or nature preserves since the 1940s, from Hutton's Siccar Point in Scotland and international Global Stratotype Section and Point to exposures of regional interest. Many sites continue to be added in Europe, and activity is spreading globally. An overview of the Geological and Landscape Conservation Movement is presented in O'Halloran et al. (1994). Perhaps we should follow the Europeans’ lead and preserve geological sites of interest.

This outcrop is rapidly weathering and has receded several feet since first being exposed in the 1980’s. Weathering is accentuated along northwestern moderately dipping fractures and veins.

**PLEASE DO NOT DAMAGE THIS EXPOSURE WITH HAMMERS OR EVEN DIRTY LOOKS. WE WISH TO MAINTAIN THE INTERGRETY OF THIS OUTCROP. WE PREFER YOU TAKE
Shawangunk and Bloomsburg sediments. Sourced by northwest draining streams, these sediments represent fluvial, estuarine, lagoonal, tidal flat and offshore bar and beach environments. Poxono Island and Bossardville sediments continue a marine transgression and mark initiation of carbonate deposition. They indicate a brackish supratidal and intertidal flats depositional environment (Epstein et al., 1967; Barnett, 1970; Epstein, 1986). Locally developed biostromal reef complexes (corals? +
stromatoporoids?) developed on a carbonate shelf that experienced some siliciclastic influx during Decker deposition. Relative sea level fell as witnessed by the restricted lagoonal and tidal flat paleoenvironments of the Rondout (Herpers, 1951; Epstein et al., 1967; Epstein and Epstein, 1969).

The Rondout marks the end of the Silurian and the beginning of the Devonian (Denkler and Harris, 1988). Subsequent marine transgression ensued as the lower Helderberg Group marks a shift from supratidal, through lagoonal and intertidal deposits of the Manlius Limestone (Epstein et al., 1967; Barnett, 1970; Smosna, 1989) down to high-energy shallow water conditions of a carbonate shelf/shoal and associated patch reefs of the Coeymans Formation. Successively deeper shelf settings followed through the Kalkberg Formation into argillaceous limestones of the New Scotland Formation. In this region the overlying Alsen Formation records a minor regression and redeposition of fossiliferous cherty limestone. The same regression is marked in eastern NY by crinoidal grainstones of the Becraft Limestone. Deepening sea level returned back to deeper subtidal and controlled the Port Ewen Shale deposition (Barnett, 1970). This marked the end of the Helderberg Group deposition.

Helderberg Group rocks are the basis of this STOP. Epstein et al. (1967) characterized the variations of these units along strike from eastern Pennsylvania through northwestern New Jersey into southern New York (Figure 142). The different members highlight the regionally shifting paleoenvironmental changes along the carbonate margin. (Description of the different limestone units follows the terminology originally devised by Dunham [1962], and subsequently modified by Embry and Klovan, [1971]. Refer to Figure 143 for a definition of terms.) Only the Shawnee Island Member of the Coeymans Formation is exposed at this STOP; the underlying Thacher Member of the Manlius Limestone crops out in the woods to the east. The overlying units, the Kalkberg and New Scotland Formations, are covered nearby but are exposed along strike to the northwest and southwest.

**PATCH REEFS AND FLANK PALEOENVIRONMENTS**

Along its outcrop belt from eastern Pennsylvania through western New Jersey, eastern New York and westward into central New York, the Coeymans Formation contains at least 14 isolated reef

![Figure 142. Fence diagram portraying the relationship of the different members of the Coeymans Formation from eastern Pennsylvania through western New Jersey and into southern New York. Four patch reefs are identified within the Shawnee Island Member. Numbered measured sections are from Epstein et al. (1967) and lettered sections from Spinks (1967).](image-url)
deposits (Oliver, 1960; Rickard, 1962; Epstein et al., 1967; Isaacson and Curran, 1981; Precht, 1982, 1984, 1989; Smosna, 1989). Nine-reefs have been documented in the Deansboro Member of the Coeymans Formation in the Syracuse, New York, region and the remaining are in the Shawnee Island Member in northeastern Pennsylvania and northwestern New Jersey (Figure 142). Each Shawnee Island build up is a patch reef with a central core ranging up to 525x230 feet (160x70 m) and 50 feet (15 m) thick, accompanied by associated flank beds (Figure 144) (Epstein et al., 1967; Precht, 1982, 1984, 1989; Finks and Raffoni, 1989; Raffoni and Finks, 1989).

**Flank Beds**

Reef flank facies consist of bedded bioclastic reef debris that thin away from the reef proper (James and Bourque, 1992) (Figure 144). Wilson (1975) differentiates two separate facies, that of flank beds and talus. Bioclastic debris is the exclusive sediment of flank beds while talus also contains lithoclastic material. Lithoclastic material consists of partially lithified micritic material ripped up from underlying beds. Talus facies deposits are rare compared to flank bed facies occurrence (Wilson, 1975). Tucker and Wright (1990) suggest that these flank deposits only form near wave base and are therefore grainstone dominated. High-energy waves and currents washed over these skeletal debris deposits and removed all micritic material (Smosna, 1989).

Reef flanks, as seen here, contain a high content of reefal debris material. This material lies broken and fragmented without any accompanying micritic or lithoclastic material. Fragmental hemispherical *Favosites* and crinoids constitute the greatest percentage of debris (Figure 145). Scarcely stromatoporoids can also be seen on the exposure. Rare brachiopods are present though I have yet to

Figure 143. Carbonate classification as devised by Dunham (1962) and later modified by Embry and Klovan (1971). It should be noted that bafflestone, bindstone and framestone are common applied to reef facies while floatstone and rudstone are more indicative of flank deposits.
find any examples. Crinoids commonly weather in relief and are extremely fragmented. Single specimens often can be plucked directly from the outcrop. The absence of lithoclasts and presence of fragmented and rotated skeletal debris characterize this exposure as a reef flank depositional environment.

Relative abundance of the different bioclastic material varies over the outcrop. Corals exist across the outcrop but they dominate in the southern beds. Crinoids occur as dispersed debris and also concentrated as bed load. They are moderately sorted in the absence of coralline material. Stromatoporoids are least abundant debris across the exposure.

A general grain size trend exists across central beds of the exposure. Coarser debris, dominated by fragmented hemispherical corals forming a rudstone texture, occurs towards the south while finer material, smaller coralline debris and a general grainstone texture, towards the north. The lowest bed maintains a fairly uniform lithology across the outcrop starting as wackestone to packstone and changing upwards through thin lenses of fragmental crinoids into a bioclastic rudstone.

The character of the limestone beds indicates a reef flank paleoenvironment. The absence of micritic material and lithoclasts suggests an environment above wave base, in strong currents but not reef proximal enough to be considered talus. The corresponding reef could be south of this exposure as indicated by this outcrop’s apparent coarser grain size in that direction. Finks (oral communication, 2001) suggests that this site is proximal to a patch reef, currently not exposed.

### Local Reefs

One of the Shawnee Island reef examples occurs in Montague Township, only 1.5 miles south-southeast of this stop. First documented by Epstein et al., (1967), it has subsequently been studied by Precht (1982, 1984, 1989), Finks and Raffoni (1989) and Raffoni and Finks (1989). Precht (1989) described four growth zones that follow the successive reef growth stages of stabilization, colonization, diversification and domination (Figure 146) as described by Finks and Raffoni (1989). The four growth stages of Precht (1989) are:

<table>
<thead>
<tr>
<th>STRUCTURE</th>
<th>STAGE</th>
<th>LIMESTONE</th>
<th>DIVERSITY</th>
<th>SHAPE</th>
</tr>
</thead>
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<tr>
<td>REEF</td>
<td>Climax</td>
<td>Framestone</td>
<td>High</td>
<td>Domal Lamellar</td>
</tr>
<tr>
<td></td>
<td>Diversification</td>
<td>Framestone Bindstone</td>
<td>High</td>
<td>Massive Lamellar Branching Encrusting</td>
</tr>
<tr>
<td></td>
<td>Domination</td>
<td>Bindstone Framestone</td>
<td>Low</td>
<td>Laminate Encrusting</td>
</tr>
<tr>
<td>MOUND</td>
<td>Colonization</td>
<td>Bafflestone Floatstone</td>
<td>Low</td>
<td>Branching Lamellar</td>
</tr>
<tr>
<td></td>
<td>Stabilization</td>
<td>Grainstone Rudstone</td>
<td>Low</td>
<td>Skeletal Debris</td>
</tr>
</tbody>
</table>

Figure 144. General arrangement of a Devonian stromatoporoid patch reef including flank beds that can contain extensive crinoidal debris as seen at STOP 5. Reefs may build multiple cores and flank beds as they prograde and/or construct vertically towards sea level. Windward and leeward sides have been identified at the Montague reef (Finks and Raffoni, 1989). Shapes tend to be elongate and trend parallel to depth contours. Modified from Wilson (1975).

Figure 145. Photograph of reef debris within these flanks deposits. Crinoid fragments occur dispersed throughout the photograph and outcrop. Rare stromatoporoid can be found as seen here. At this spot on the exposure the matrix material is almost exclusively composed of fragments of the hemispherical coral Favosites. US quarter-dollar for scale. James and Bourque (1992). The four growth stages of Precht (1989) are:
1. Deposition of massive bedded crinoidal packstones to grainstones that stabilized the substrate.
2. Large domal and branching forms of the tabulate coral *Favosites* colonize the substrate, initiating reef formation.
3. Development and diversification of reef core where bafflestones, floatstones, and framestones predominate. Reef growth includes branching tabulate corals (*Cladopora* and *Favosites*), domal and planar tabulate corals (*Favosites*), rugose corals, and domal and laminar stromatoporoids.
4. Massive stromatoporoids overgrown by laminar and encrusting stromatoporoids and encrusting algae predominate and develop into bindstones and framestones under shallow water conditions. Diversity of tabulate corals and stromatoporoids diminishes drastically as compared to lower stages.

Finks and Raffoni (1989) and Precht (1989) suggest that deeper, more open water existed north of the Montague reef. The reefs formed and prograded across gently northward-sloping sea floors in 10-20 m water depth (Smosna, 1989; Precht, 1989) (Figure 148). Reefs built up to sea level and higher energy flow conditions (Epstein et al., 1969; Finks and Raffoni, 1989; Precht, 1989) (Figure 148).
Leave STOP 5, exiting Mini Mall and turning right onto Clove Road.

0.4 65.9 Descend ice-contact slope of valley-head delta, entering the upper end of Mill Brook valley. Mill Brook Valley drains northeastward. It is a prolongation of Wallpack Valley, continuing along the same strike belt of Silurian and Devonian rocks.

0.6 66.5 Abandoned sand and gravel pit on right. Collapsed foreset beds of sand, pebbly sand, pebble gravel, and cobbles gravel are interlayered with flow till. The many small sand-and-gravel pits in this valley all revealed collapsed foreset beds of varying texture. In most places, the deltaic outwash is found along the sides of the valley, suggesting deposition against stagnant ice that occupied the center of the valley—the classic view of a kame terrace. Reconstructed longitudinal profiles of the deposits show that their elevation was controlled by spillways across the valley-head delta and cols through Wallpack Ridge. This shows that deglaciation occurred by stagnation-zone retreat rather than by valley-bottom stagnation.

0.2 66.7 Road cut in “kame terrace”.

0.3 67.0 New Mashipacong Road.

0.3 67.3 Abandoned sand and gravel pit on right.

0.4 67.7 Abandoned sand and gravel pit on right.

0.2 67.9 Westfall Farm on left.

0.2 68.1 Abandoned sand and gravel pit on left.

0.6 68.7 The lime kiln on left sits on outwash.

0.5 69.2 Cross Clove Brook.

0.4 69.6 Abandoned sand and gravel pit on right.

0.7 70.3 At stop sign, turn right on NJ 23 South.

0.75 71.05 Stratified drift exposed in road cut to left. Based on the position of the outwash here, it appears to be kame (ice-hole filling).

0.2 71.25 Enter High Point State Park and climb northwest flank of Kittatinny Mountain.

0.05 71.3 Bloomsburg Red beds on left side of road. The hill here is underlain by south-dipping Bloomsburg Red Beds.

0.15 71.45 Borrow pit in till on left

0.4 71.85 Boulder field on right.

Figure 148. Photograph of reef debris within these flanks deposits. Crinoid fragments occur dispersed throughout the photograph and outcrop. Rare stromatoporoid can be found as seen here. At this location of the exposure the matrix material is almost exclusively composed of fragments of the hemispherical coral *Favosites*. US quarter dollar for scale.
STOP 6. HIGH POINT: OVERVIEW OF BEDROCK GEOLOGY, GEOMORPHOLOGY, AND THE CULVERS GAP RIVER.
Leaders: Jack B. Epstein, Donald H. Monteverde, and Ron W. Witte.

INTRODUCTION

This stop is located in High Point State Park (Figure 149) along the northeastern most extent of Kittatinny Mountain in New Jersey. High Point affords its visitors exceptional vistas….To the southwest is the curving ridgeline of Kittatinny Mountain. The small notch near the ridge’s midpoint is Culvers Gap (STOP 4). Eastward lies Kittatinny Valley with the New Jersey Highlands in the background. To the northeast, Kittatinny Mountain gives way to a broad upland that forms the Shawangunk Mountains. The Catskill Mountains, which can be seen only on a very clear day, form an area of high, jagged relief west of “the Shawangunks.” Westward lies the uneven Pocono Plateau, an area underlain by gently northwest-dipping shale, siltstone, and sandstone. Extensive erosion of these rocks over millions of years has created a rugged landscape that remains largely uncultivated. The Delaware River flows
southeastward off the Pocono Plateau to Minisink Valley where it makes a right-hand turn, continuing its course to the southwest. The slopes to the south and east of High Point house graywacke and slates of the Martinsburg Formation. Sequestered within the Martinsburg is the Paulins Kill Valley, underlain by Cambro-Ordovician carbonate rocks. Beyond are well-defined ridges in the Hudson and New Jersey Highlands consisting of Precambrian metamorphic rocks, upon which are the 1,000-foot-high slopes of the Vernon Ski area. Perhaps, after our long day in the field, it is a time for quiet contemplation, rather than a lengthy and loud discussion about the past.

**Park History**

In 1890 Charles Saint John, Jr. purchased a 1700-acre tract of land from the Rutherford family that encompassed the summit of Kittatinny Mountain around the future site of High Point Monument. St. John built the High Point Inn on a ridge overlooking Lake Marcia. The Inn opened for business in 1890 as a summer resort. For a time, the Inn was a popular place, but in 1908, it was closed because it was not profitable. Colonel Anthony Kuser and his twin brother John purchased the property in 1910. In 1911, Anthony Kuser’s father-in-law, John F. Dryden, purchased an additional 7,000 acres from the Rutherford family giving the Kuser’s a combined estate of about 10,400 acres. In 1923, the Kuser’s decided to donate their land for the creation of High Point State Park in memory of John Dryden. The park was originally administered by High Point Park Commission until their responsibilities were centralized in the State Park Service in 1945. During the 1930’s, the Civilian Conservation Corps (CCC) operated two camps at High Point. CCC workers improved roads, constructed trails, picnic pavilions, and camping sites. They built the old Iris Inn, which is now the park office, and created Steeny Kill Lake by constructing a dam across Clove Brook. In 1965, the cedar swamp was set aside as The John Dryden Kuser Memorial Natural Area in memory of the former New Jersey State Senator and son of Colonel and Mrs. Anthony Kuser.

The High Point Monument was built on the highest point in New Jersey (1803 feet, or 549.5 m, above sea level) and was dedicated in memory of New Jersey’s wartime heroes. The construction of the monument, which was funded by Colonel and Mrs. Kuser, began in 1928 and was completed in 1930. The 220-foot high monolith is faced with New Hampshire granite and local quartzite. For those used to the rigors of climbing, steps lead from the monument’s base to the top of the structure. Presently, the monument is closed for restoration. (Information about the history of High Point State Park is from New Jersey Department of Environmental Protection [1994].)

**BEDROCK GEOLOGY**

**Local**

At High Point, we are standing on the Shawangunk Formation of probable Middle Silurian age (Figure 150). The Taconic unconformity just below the Shawangunk dominates the local geology. The trip has already traversed this unconformity twice in the last 1.5 miles, but unfortunately the contact is not exposed. I-84 construction uncovered the contact (Figure 151), approximately 3 miles north and displayed a thin clayey fault gouge similar to that observed at Yards Creek (STOP 3) separating the Martinsburg from the overlying Shawangunk. Epstein recorded a slight angular divergence between the two units. Another exposed contact originally described in New Jersey Geological Survey historical notes depicts a classic erosional unconformity where Shawangunk coarse clastics bevel the Martinsburg without apparent structural deformation (Figure 152). Again only a slight angular discordance separates the Shawangunk and Martinsburg Formations. Colluvium covers the Martinsburg beneath the contact.
The Taconic unconformity represents a major break in the rock record. It separates two diverse units that represent distinctly different paleoenvironments. Beneath the unconformity, the Martinsburg sediments represent deposition in a deep foreland basin. Turbidite deposition occurred both under northwestward-directed downslope conditions as well as parallel to the southwest-northeast trending basin axis (McBride, 1962). The Martinsburg foreland basin displays a pronounced along-strike morphology with deeper environments represented by the commercial slate of the Penn Argyl Member in eastern Pennsylvania and the thick sandstone beds of the locally occurring High Point Member signifying more proximal position (Drake and Epstein, 1967; Drake, 1991). McBride (1962) showed that paleoflow in New Jersey parallels the bathymetric basin axis towards the southwest whereas Martinsburg sediments, located farther westward, were carried northeast. This indicates a deep basin depocenter in eastern Pennsylvania, around the Penn Argyl outcrop belt. The youngest Martinsburg is late Middle Ordovician in age (Parris and Cruikshank, 1992) with deposition synorganic in the Taconic orogeny. The Shawangunk resides on the other side of the unconformity. This clastic wedge developed after filling and uplift of the Martinsburg foreland basin. A mountain range to the east built by the progressive Taconic northwest verging fold and thrusting supplied clastic detritus transported westward in fluvial systems until final deposition under braided stream conditions (Epstein and Epstein 1972, Epstein, 1993). A sedimentological problem, common throughout the Appalachians, is why there is a dearth of pebbles derived from rocks overlying the Precambrian. Sediment provenance shows the source as an uplifted and unroofed Grenville terrain (Gray and Zeitler, 1997).

Figure 150 – Block diagram of the High Point State Park region. The resistant Shawangunk Formation underlies Kittatinny Ridge crest. Bloomsburg forms the northwest mountain slope while the Martinsburg forms low lying topography to the east. Thicker bedded sandstone units within the Martinsburg hold up the low ridges. Depression of the Martinsburg foreland basin displays a pronounced along-strike morphology with deeper environments represented by the commercial slate of the Penn Argyl Member in eastern Pennsylvania and the thick sandstone beds of the locally occurring High Point Member signifying more proximal position (Drake and Epstein, 1967; Drake, 1991). McBride (1962) showed that paleoflow in New Jersey parallels the bathymetric basin axis towards the southwest whereas Martinsburg sediments, located farther westward, were carried northeast. This indicates a deep basin depocenter in eastern Pennsylvania, around the Penn Argyl outcrop belt. The youngest Martinsburg is late Middle Ordovician in age (Parris and Cruikshank, 1992) with deposition synorganic in the Taconic orogeny. The Shawangunk resides on the other side of the unconformity. This clastic wedge developed after filling and uplift of the Martinsburg foreland basin. A mountain range to the east built by the progressive Taconic northwest verging fold and thrusting supplied clastic detritus transported westward in fluvial systems until final deposition under braided stream conditions (Epstein and Epstein 1972, Epstein, 1993). A sedimentological problem, common throughout the Appalachians, is why there is a dearth of pebbles derived from rocks overlying the Precambrian. Sediment provenance shows the source as an uplifted and unroofed Grenville terrain (Gray and Zeitler, 1997).
High Point rests on the Shawangunk Formation. Looking southwestward, the unconformity, projecting halfway across Lake Marcia, is exposed along a doubly plunging anticline that folds both the underlying Martinsburg and overlying Shawangunk (Figures 153 and 154). The anticline continues through the Shawangunk west of the High Point monument before plunging out. The glacially polished rocks exposed here at the ridge crest form the southeast dipping limb of the anticline (N38°E/55°SE). The corresponding syncline folding both the Martinsburg and Shawangunk lies to the east.

Shawangunk folds are doubly plunging, discontinuous structures that commonly form en-echelon. Locally, they develop as upright, open structures except just west of the major ridge where overturning is evident. Gently southeasterly inclined fold hinges mark these structures that could be related to blind thrusting in the basal Shawangunk and Martinsburg. Seismic sections along I-84 depict northwestward-directed thrust faulting within the Martinsburg as the Shawangunk contact is approached. These structures have been truncated by southeast-directed backthrusts (Herman et al., 1996).

Figure 151 – Taconic unconformity exposed in the Interstate Route 84 construction cut. A thin fault zone exists along the contact. There is a slight angular divergence between the Shawangunk Formation bedding above and the Martinsburg Formation below.

Figure 152 – Taconic unconformity exposed in Stokes State Forest, several miles south of Stop 6 in the Branchville quadrangle. The contact shows an angular unconformity that down cuts to the north. Dotted line marks the unconformity and solid line the Martinsburg bedding. No deformation was evident along this contact. The basal Shawangunk is a thick bedded, quartz pebble conglomerate. This basal unit portrays this characteristic rough weathering profile all along the Kittatinny Mountain front. Only the upper two feet of Martinsburg is exposed.
Looking to the northeast, a broad arch can be seen in the Shawangunk at Ellenville, New York (Figure 10, p. 15). Because of the excellent exposures at this locality, we are able to separate Taconic from Alleghanian structures, as described by Epstein and Lyttle (this guidebook, p. 22). They determined that only gently upright folds of Taconic age are found adjacent to the Shawangunk contact and northwest of Ruedemann’s line. Similar folds in the Martinsburg occur at the unconformity in the High Point region. Here the Martinsburg outcrops form a uniformly northwest-dipping panel except in close proximity to the Shawangunk. Small folds affect both the Shawangunk and Martinsburg beds (Figure 154). These folds plunge out over short distances and commonly continue for less than one mile. Several igneous bodies intruded along bedding cut the Martinsburg. These intrusive bodies relate to the Beemerville Intrusive Complex and related diatremes exposed just below the Taconic unconformity in the Branchville quadrangle to the south. Taconic structures, however, become increasingly more complex to the southeast.

**Regional**

The regional geologic relations are shown in Figure 10 (p. 15). The structure of Blue-Kittatinny-Shawangunk Mountains changes markedly from the complex folding and faulting in the southwest, such as at Hawk Mountain and Lehigh Gap, through overturned folds at Delaware Water Gap (STOP 1) and Yards Creek (STOP 3), to more upright and open folds at High Point. This deformational trend in the Shawangunk is fairly uniform and persistent along the ridge. Immediately west of Culvers Gap and High Point, the Bloomsburg through Marcellus outcrop belt widens to nearly eight miles in the vicinity of Montague, NJ. These rocks, between Wallpack Center and Port Jervis, gently dip northwestward forming a monoclinal panel (Monteverde, 1992; Monteverde and Epstein, in prep). Spink (1967, 1969, 1972) hypothesized that the decreased structural intensity relates to strain shadow effects of a large intrusion, the Beemerville Intrusive Complex, at and near the Shawangunk-Martinsburg contact.

Several lines of reasoning indicate the possibility of the Beemerville Intrusive Complex acting as a buffer to tectonic strain propagation into the foreland during the Alleghanian orogeny (Figure 155). The propagation of both penetrative and mechanical strains has been affected by the intrusion. Cleavage developed within Upper Silurian through Upper Devonian sediments displays similar regional trends. Southwest of Wallpack Center, NJ through Stroudsburg, PA. Lower Devonian sediments exhibit an intense southeast-dipping, closely spaced, regional cleavage. In select areas the well-developed cleavage in the Esopus is a defining characteristic used in bedrock mapping (Alvord and Drake, 1971). Between Wallpack Center and the New York State line, cleavage is almost completely absent in
Devonian units (see discussion at STOP 10, Day 2). Locally a spaced cleavage may occur but it is not penetrative. Compare the well-developed cleavage within the Onondaga and Marcellus at Stops 8 and 9 with the less penetrative pencil cleavage in the Mahantango at STOPS 10 and 12. The regional geologic map accompanying STOP 5 does not display a strong regional cleavage in the post-Bloomsburg sediments. Reduced cleavage development projects at least up into the Mahantango Formation.

Fold formation and trend analysis show the mechanical strain interference pattern created by the relatively stiff vertical wall of the Beemerville. There is an absence of any folding within the Upper Silurian through Upper Devonian sediments west of the Beemerville. Along strike on Wallpack Ridge in New Jersey and Pennsylvania the post Bloomsburg through Onondaga sediments delineate tight, upright to northwest directed overturned folds. The folds locally become subdued and more open as they propagate upwards through the Oriskany to the Marcellus. Similar steep to overturned bedding occurs to the north along Trilobite Mountain—over the border in New York. These same sedimentary units only display very gentle (8-30° NW) uniform dips within the strain shadow. Along the southern strain-shadow edge, Monteverde (1992) mapped several fold axial trends that diverged by up to 20° from the northeast-southwest regional orientation into a more west-southwest direction. The fold pattern suggests that the Silurian-Devonian units deflected around the Beemerville Intrusive Complex. A similar though muted trend appears with the Martinsburg fold pattern east of the pluton.

The Beemerville’s regional extent is much larger than depicted on geologic maps (Drake et al., 1996). Ghatge et al. (1992) and Jagel (1990) used gravity and magnetic methods to model the intrusive complex’s subsurface expression southwestward. They depict the pluton continuing almost to Culvers Lake and thickening to approximately 2 miles wide at 7000 feet depth. This large unfoliated plutonic body could disrupt the otherwise gently folded pre-Alleghanian structural configuration along which strain translated towards the foreland. Regional cross section analysis offset the syenite along a basal
decollement rooted in basement but not within 4 miles below the present erosion surface (Herman et al., 1996).

Some questions remain concerning the strain shadow hypothesis. Directly above the Beemerville lie the Shawangunk and Bloomsburg. Folds within the Shawangunk remain relatively unchanged along Kittatinny Mountain. Overturned beds can be seen immediately southwest of the exposed pluton suggesting the igneous body did not buffer the progressing strain. But there are fewer overturned panels close to the pluton than farther along strike to the southwest. The mapped Shawangunk outcrop belt does double its width behind the intrusion, similar to the Upper Silurian and Devonian units. The most dramatic outcrop widening occurs in the Bloomsburg. Pleistocene glacial sediments blanket most of the Bloomsburg so their structural signature cannot be interpreted.

**EVOLUTION OF DRAINAGE AND THE CULVERS GAP RIVER:**

*Figure 155 – Regional structural interpretation of the Kittatinny Mountain region of New Jersey. Map depicts the outcrop location of the Beemerville Intrusive Suite shown in orange as well as the subsurface extension in blue, identified by Ghatge et al., 1992. The outcrop width of the Shawangunk and Bloomsburg increases markedly behind the intrusion. Fold trend deflects in the Silurian Devonian and also the Ordovician sedimentary units along the southern area of the intrusion. Base map is a compilation of 16 DEM quadrangle consisting of Blairstown (BL), Branchville (BV), Bushkill (BK), Culvers Gap (CG), East Stroudsburg (ESB), Edgemere (EM), Flatbrookville (FV), Lake Maskenozha (LM), Milford (MF), Ponds Eddy (PE), Port Jervis North (PJN), Port Jervis South (PJS), Stroudsburg (SB), and Twelve Mile Pond (TP). DEM’s have a 5x vertical distortion and a southeast looking illumination.*

A view from the Schooley peneplain, or a view from the summit of a very long, monoclinal ridge, held up by a very resistant rock that weathers so slowly over time that it will stand out in greater relief amongst the softer rocks on its flanks.

*The scientist. He will spend thirty years in building up a mountain range of facts with the intent to prove a certain theory; then he is so happy in his achievement that as a rule he overlooks the main chief fact of all – that his accumulation proves an entirely different thing.*

Mark Twain, “The Bee”
Standing on the summit of Kittatinny Mountain (Figure 156), we would be remiss not to provoke a discussion on the evolution of drainage in this part of the Appalachians. Given the enormous body of literature on the geomorphic evolution of the Appalachians, it is not the intent of the authors’ to discuss in detail the many hypotheses and controversies that surround this complex subject, but rather only attend to the later geomorphic history of Culvers Gap. Although, many will argue that these two are not mutually exclusive. The following discussion will focus on the course of the late Culvers Gap River, and the timing and abandonment of Culvers Gap.

Upon examination of the topography of Culvers Gap (Figure 157), there should be little doubt that: 1) a river (named the Culvers Gap River in this report), smaller than the modern Delaware River, had at some time flowed through the gap; 2) the gap was largely the work of the Culvers Gap River over a long period, although erosion during three or more glaciations have further widened and deepened it since it was abandoned; and 3) at some point the Culvers Gap River was captured by the Delaware River in what is now Minisink Valley. There are several scenarios by which this capture may have been facilitated. These will all be discussed in the following sections. In addition, the idea that the Culvers Gap River may have been displaced by an early glaciation will also be discussed.

Map and Field Observations

Culvers Gap is a prominent wind gap in northwestern New Jersey that lies in the Delaware River drainage basin. The gap is approximately 23 miles northeast of Delaware Water Gap, and it forms a prominent pass through Kittatinny Mountain that joins Kittatinny and Minisink Valleys. The Ogdensburg-Culvers Gap moraine is a recessional moraine of late Wisconsinan age, and it forms a plug of thick drift in the gap (see STOP 4). Based on nearby well records, the elevation of its bedrock floor has been estimated to lie between 750 feet (230 m) and 800 feet (245 m) above sea level, which makes it the lowest wind gap through Kittatinny Mountain. Glacial erosion has undoubtedly lowered the gap’s floor since it was abandoned. However, all the gaps mentioned in this report along Kittatinny Mountain have experienced a similar history of weathering and erosion with the exception of the number of...
glaciations, or they lie well above the height of Culvers Gap. An exception may be Wind Gap (Figure 158), which lies south of the late Wisconsinan limit, but inside the Illinoian glacial limit. Its rock floor lies at an elevation of 980 feet (299 m) above sea level. It is arguable that the erosional effects of an additional glaciation may have lowered the floor of Culvers Gap below Wind Gap. However, it seems very unlikely, given the difference in their elevations, and the resistant nature of the Shawangunk Formation to glacial erosion. Because Culvers Gap also lies in the Delaware River drainage basin, its abandonment presumably represents the last major adjustment of drainage in this part of the basin.

Descending from the Catskill Mountains, the West Branch and East Branch of the Delaware River, follow a southwesterly course onto the Pocono Plateau. After a distance of approximately 50 miles (80 km), both branches turn to the southeast and join each other near the town of Hancock, New York. From here the river flows approximately 60 miles (100 km) southeastward towards the town of Port Jervis, New York, along a slightly meandering course that is deeply incised in the gently, northwest-dipping country rock. At Port Jervis, it enters Minisink Valley where it makes a sweeping right-hand turn and flows southwestward following the trend of the Marcellus Shale in a course that parallels the southwest-trending monoclinal structure bedrock structure. At Wallpack Bend, the river follows a large, structurally controlled meander that cuts through Wallpack Ridge. From here it turns back to the southwest following a similarly parallel course as the stretch north of the bend. At Delaware Water Gap, a distance of approximately 35 miles (55 km) from Port Jervis, the river turns back to the southeast through an “S”-shaped bend and enters Kittatinny Valley. From here it flows a distance of approximately 80 miles (130 km) to the fall line at Trenton, New Jersey, generally following a southeastward course that cuts across the strata that
underlies Kittatinny Valley, New Jersey Highlands, and the Piedmont. Water gaps south of Delaware Water Gap include those at Marble Mountain, Riegelsville (1 and 2), Lambertville, and Titusville.

In many places the river is cut down as much as 200 feet (60 m) in rock and it lies well below the highest parts of the Pocono Plateau, Kittatinny Mountain, and the New Jersey Highlands. From Port Jervis, New York, to Foul Rift, New Jersey, the river flows over thick deposits of glaciolacustrine and glaciofluvial sediments laid down during the late Wisconsinan glaciation.

The strongly developed northeast to southwest topographic grain of the study area is aligned with the fold axes and fault traces formed in the Paleozoic during the Taconic and Alleghanian orogenies (Figure 155). The Delaware River, from its headwaters in the Catskill Mountain to the fall line at Trenton, New Jersey, flows either in a southeasterly direction crosscutting the strike of the country rock or in a southwesterly direction parallel to bedrock strike. As previously shown by Epstein (1966), most of the gaps through Kittatinny Mountain are located in areas where there are structural weaknesses. Culvers Gap is positioned across the narrowest part of resistant Shawangunk Formation, and it lies near a structural flexure where immediately to the northwest the outcrop width of the Shawangunk broadens by as much as 300 percent (Figure 155).

Projection of the Culvers Gap River southeastward in Kittatinny Valley shows that it would now fall near the drainage divide between the Delaware and Hudson Rivers (Figure 158). This shows that there has been a complete topographic reversal in the valley since the time the gap was abandoned. The lowest parts of the modern divide lie about 400 feet (120 m) above sea level. This unsophisticated analysis shows that there has been a minimum of 350 feet (105 m) of erosional lowering in Kittatinny Valley since the capture of the Culvers Gap River.

Further examination of the topography (Figure 1580) shows that some gaps in Kittatinny Mountain are aligned in a southeast direction with gaps cut through the New Jersey Highlands and Reading Prong. The most conspicuous of these is the alignment of Wind Gap with Marble Mountain Gap and the two gaps at Riegelsville, and Delaware Water Gap with Pequest Gap, Oxford Gap, and Glen Gardner Gap. These alignments were cited by Johnson (1931) as proof of regional superposition. In contrast, Epstein (1966)
showed that most of the gaps through Kittatinny Mountain, and especially the larger ones, are positioned in areas of structural weakness. In addition, many Highlands and Reading Prong gaps (Figure 158) lie along lines of structural weakness. Additionally, the absence of paleo-drainage northwest of Culvers Gap on the Pocono Plateau and the very small drainage area northwest of Delaware Water Gap would also argue against regional superposition.

**Geomorphologic History (Summary of Views)**

The following summary limits discussion to the history of the Culvers Gap area. For those interested in the “big picture”, excellent reviews on Appalachian geomorphic history and origin of wind and water gaps are found in Morisawa (1989) and Clark (1989). Water gaps and wind gaps provide direct evidence for transverse drainage, and wind gaps further show that the courses of large streams have changed over time. The study of gaps can generally be summarized into two main camps of thought. The first includes all theories that suggest the location of gaps through resistant ridges of bedrock was fortuitous, either the result of: 1) superposition as envisioned by Johnson (1931), or 2) incision of uplifted, regional erosional plains (peneplains), where most of the relief had been formerly reduced to base-level (Davis, 1889). During subsequent uplift, streams remained in their gaps, their consequent courses becoming antecedent. The second group includes the more recent investigations of Myerhoff and Olmstead (1936), Thompson (1939), and Oberlander (1985). These studies have explained the origin of gaps in terms of headward erosion and stream adjustment to bedrock structure, former location of Appalachia, and migration of the Atlantic-Mississippian drainage divide. The location of gaps is not by chance, because they crosscut resistant rocks in areas that show evidence of structural weakness.

Davis (1889) suggested the present drainage system in eastern Pennsylvania and northwestern New Jersey started to form during the Permian Period. The master stream at this time, named the Anthracite River, drained toward the continental interior chiefly along longitudinal courses that developed on southwest- to northeastward-trending folds. Transverse west to northwesterly courses formed where structural sags, faulting, or extensive jointing weakened the country rock. The younger, southeastward flowing rivers started to form during the Triassic Period during continental rifting and subsequent formation of the Atlantic Ocean. (Though Davis did not explain the Triassic rifting in exactly these terms!) These streams, because they had shorter courses to the newly formed Atlantic Ocean, were more aggressive than the tributaries of the Anthracite River, and over time, they captured parts of it. The headward erosion of the Atlantic streams, especially their encroachment across belts of resistant rock, was aided by the previous overall lowering of the landscape to a gently sloping plain of low relief called the Schooley peneplain. This erosional surface was thought by Davis to be now represented by the crest of Kittatinny Mountain, and accordant hilltops in the New Jersey Highlands. Later additional periods of uplift and erosion led to the incision of the Schooley surface and the development of lower peneplains. During these subsequent Davisian cycles, streams became further adjusted to structure, and capture occurred where conditions favored one stream over another. In post-Schooley time subsequent tributaries of the Delaware River, working their way along belts of weaker rock, dismembered nearby rivers and become the master stream. Although, Davis does not directly discuss Culvers Gap or its river it seems implied in his work that the Culvers Gap River was originally part of the ancient Raritan or Passaic Rivers.

Johnson (1931) suggested all the major streams in the study area were initially flowing in a southeastward direction on a cover of unconsolidated deposits of Cretaceous age that had previously been laid down on a pre-Schooley erosional surface called the Fall Zone peneplain. Following a lengthy interval of erosion the underlying folded and faulted rocks of the Appalachians and Highlands were exposed. The water gaps and wind gaps through Kittatinny Mountain and the Highlands are places where the ancestral streams first encountered resistant rocks and cut into them. During successive
cycles of erosion and uplift, which included the formation of the Schooley peneplain, parts of these streams became adjusted to the underlying bedrock structure where as other parts remained antecedent. The present drainage was the result of a series of stream captures by subsequent tributary streams working their way headward in a northeastward or southwestward direction along belts of weak rock.

Ruedemann (1932, 1949) proposed that the original drainage developed during the Permian or Triassic, when consequent streams drained southwestward from the Catskill Mountains, while Myerhoff and Olmstead (1936) suggested that the consequent drainage developed southeastward on the more gently-sloped limbs of overturned folds. In contrast, Thompson (1939) suggested that the divide originally lay more to the southeast and through headward erosion and stream piracy had moved back to the northwest to its present position. Thompson further suggested that the gaps in the Appalachians of Pennsylvania were positioned along areas of weakness and argued there was no compelling evidence to show they had originated as a product of superposition. Oberlander (1985) and Hoskins (1987) also cited evidence to suggest that the course of the Susquehanna River above Harrisburg was controlled by fluvial adjustment to bedrock structure. Inherent to all these studies is that gaps are found in areas of structural weakness (joints, faults, change in the attitude of bedding). Work by Epstein (1966) has further shown that most of the gaps in the study area are preferentially in areas of structural weakness.

Campbell and Bascom (1933) showed that Culvers Gap River was a tributary of the Proto-Raritan River, and it flowed through the New Jersey Highlands via Andover to Ledgewood. The gap was abandoned during the early Pleistocene when its river became displaced presumably because the gap was blocked by ice or drift. Meltwater cut a new course down Minisink Valley to Delaware Water Gap where it joined a small river that drained the Brodhead Valley. Campbell and Bascom cited the youthful appearance of the Delaware Valley between Easton, Pennsylvania, and Trenton, New Jersey, as evidence for the young age of the modern Delaware River. Happ (1938) in his study on the geomorphic history of Minisink Valley supported Johnson’s earlier view that the southeastward course of the Delaware River above Port Jervis was proof of superimposed drainage, and all the prominent gaps in Kittatinny Mountain were initially the result of southeastward-flowing consequents. Over time, these drainage systems became dismembered as streams adjusted to structure and capture occurred where one stream had found an advantage over another. Mackin (1939) accepted Johnson’s theory of superposition and outlined a scenario where the Wind Gap River was captured by the Delaware River.

Late History of Culvers Gap River

The Culvers Gap River was part of the Ancient-Raritan River drainage basin during the late stages of its history. Before its abandonment of Culvers Gap, the river followed a course through the Pequest Valley and crossed the New Jersey Highlands through the Oxford and Glen Gardner Gaps into the Raritan lowland. Alternatively, the river may have crossed the Highlands south of Andover into the Musconetcong Valley, although the Cranberry Lake Gap seems a little too high to have been used at the time Culvers Gap was abandoned. From here, it followed the valley downstream to Glen Gardner Gap. Possibly, the Culvers Gap River may have flowed along this path at an earlier time, with the shift to the Oxford Gap facilitated by a capture in Kittatinny Valley by a subsequent stream working its way up the Pequest Valley. The part of the Delaware River north of Belvidere may have also been a tributary of the Culvers Gap River via the Pequest Gap, based on its small drainage area northwest of Delaware Water Gap, and youthful appearance of the modern Delaware River drainage basin south of Riegelsville. If one accepts the course of the Culvers Gap River as it has been outlined here, it may have supplied the quartzite, quartz, and chert that make up the Beacon Hill Formation. A projection of the Bridgeton surface inland to Glen Gardner Gap tentatively supports the view that the Culvers Gap River was part of the ancient Raritan drainage system.

At some point during the Late Miocene or Early Pliocene, and possibly driven by base-level
lowering and incision during the growth of the Antarctic ice cap in the Miocene and flexural uplift due to offshore sediment loading (Pazzaglia and Gardner, 2000), a tributary of the Delaware River captured the Culvers Gap River in Minisink Valley. Apparently over time, the narrow width and structural weakness of Proterozoic rocks in the Riegelsville Gaps, and structural weakness of the Shawangunk Formation in Delaware Water Gap gave the Delaware River an advantage over the Culvers Gap River and its more northerly through the New Jersey Highlands. The previous capture of Wind Gap River by the Delaware may have also hastened the demise of the Culvers Gap River. The hypothesis that an early Pleistocene glaciation may have displaced the Culvers Gap River from its course (Campbell and Bascom, 1933) is untenable because the gap lies well above the projected gradient of the base of the Pennsauken Formation; a fluvial deposit of Pliocene age (Stanford, 1993). In addition, erosion of a bedrock-floored drainage divide in Minisink or Wallpack Valleys by glacial lake drainage is highly unlikely (Witte, 1997d).

The evolution of the Delaware River and demise of the Culvers Gap River appears to have been the result of a series of stream captures that redirected the course of the master stream to the southwest. Water and wind gaps cut by these rivers are found over resistant rocks that exhibit structural weakness and have narrow outcrop widths perpendicular to drainage. The captures postulated in this report were by southwest draining subsequent streams that had cut their way headward along belts of weaker sedimentary rocks. The variation in the maturation of the Delaware River drainage basin further supports its expansion through capture, and suggests that the ancient Delaware River initially had a more difficult time expanding its drainage basin inland. A possible explanation for this latency is the nature of the rocks that lie on the southeast side of the Highlands. The rocks in the Raritan lowland are less resistant to erosion than the basalt and argillite that lie along the modern Delaware River.

Figure 159. Generalized direction of ice movement in northern New Jersey during the late Wisconsinan. Lines represent regional ice-flow movement at the base of the ice sheet. Flow directions are based on striae, drumlins, dispersal of erratics, and till provenance. Shaded areas represent major uplands. A shows direction of ice flow when the glacier margin was at the Terminal Moraine. Field data in the Kittatinny Valley area indicates ice flowed southward across the valley's southwest-trending regional topographic grain. B shows direction of ice flow during deglaciation. Flow lines in Kittatinny and Minisink Valleys and surrounding uplands are oriented in a southwest direction with well developed lobate ice flow at the glaciers margin. The change in regional ice flow to a southwest direction appears to be related to thinning of the ice sheet at its margin, and reorganization of ice flow around the Catskill Mountains, and in the Hudson-Wallkill Valley. Data from Ridge (1983), Stanford and Harper (1985), Witte (1988), Sevon et al. (1989), Stone et al. (1989), and unpublished field maps on file at the New Jersey Geological Survey, Trenton, New Jersey.
Local Glacial History

The late Wisconsinan advance of ice into northwestern New Jersey was marked by lobes of ice initially moving southwestward down Minisink and Kittatinny Valleys. Over time, ice became thicker, negating the influence of local topography on ice flow. Gradually, ice flow turned south, cutting across the region’s southwesterly topographic grain (Figure 159). Drumlins on Kittatinny Mountain (Figure 160), till provenance, erratic dispersal patterns, and striations (Witte, 1997c) support this southerly direction of ice flow and show that ice flowed over Kittatinny Mountain into Kittatinny Valley during the glacial maximum.

During deglaciation from the late Wisconsinan terminal moraine (Figure 159), the edge of the ice sheet thinned and its flow became more directed by the southwesterly trend of the larger valleys. The Kittatinny and Minisink Valley ice lobes retreated gradually to the northeast. However, at times the edge of the ice lobes did maintain a constant position, and in a few instances readvanced a few miles. During periods in which the glacier margin maintained its position, neither retreating nor advancing, outwash deposits built up at and beyond the glacier margin in stream valleys and glacial lakes in Minisink, Wallpack, and Kittatinny Valleys (Figure X-13). Additionally, end moraines were deposited at the glacier’s terminus. Both the heads of the outwash deposits and the moraines mark the former edge of the ice sheet and define its shape both geometrically and temporally.

The Augusta margin, which runs through High Point State Park (Figure 161), is delineated by the Augusta and Montague moraines and ice-contact deltas located along the eastern side of Kittatinny Mountain. The continuity and size of the moraines, and the size and extent of meltwater deposits laid down in nearby valleys down from the Augusta margin shows that the glacier’s terminus probably maintained a nearly constant position for over one hundred years.

Figure 161. Late Wisconsinan ice-recession margins and glacial lakes in the upper parts of Kittatinny and Wallkill Valleys, New Jersey and New York. Data modified from Witte (1997a), Connally et al. (1989), Stanford and Harper (1985), and Ridge (1983).
There are excellent examples of striations on the glacially polished rock surface northeast of the monument. They show that ice flowed S 80°W to due west. In places weathering has removed striations and roughened up the polished rock surface. This suggests that parts of the outcrop were once covered by thin till which lessened the effects of weathering. Later, the till was stripped by erosion. Crescentic marks (Figure 162) are also well exposed, and indicate a similar direction of ice flow. Westerly ice flow atop Kittatinny Mountain may at first seem enigmatic. However, these glacial markings were made near the glacier’s margin where ice flow was controlled by the lobate geometry of the Kittatinny Valley lobe. Because ice was thicker along the central axis of Kittatinny Valley flow lines diverged outward near the glacier’s margin. Divergent flow is recorded by striations in northwestern New Jersey (Figure 160) and the occurrence of Martinsburg erratics. These sandstone boulders, some as much as 10 feet (3 m) in diameter were carried upwards and out of Kittatinny Valley to the top of Kittatinny Mountain, where they rest as much as 900 feet (275 m) above the valley floor.

Leave STOP 6, returning to park entrance on NJ 23.

1.45  76.95  Bear right at sign for Port Jervis, and at stop sign turn right on NJ 23 North.
0.35  77.3   Turn left onto Sawmill Road.
0.1   77.4   Intersection with Ridge Road on right. Continue straight.
1.0   78.4   Low Shawangunk ridge on right.
0.6   79.0   Dam at south end of Sawmill Pond on left.
0.1   79.1   Cross over Sawmill Brook.
0.35  79.45  Beaver Pond on right. The many swamps and poorly-drained areas are typical of constructional glaciated landscapes. Thick till deposited in drumlins, ground moraine, and recessional moraine has completely buried the preglacial surface concealing former stream courses. Upon deglaciation, surface water, which had at one time flowed off the mountain in a well defined network of streams, became momentarily trapped in the many depressions, swales, and glacial lakes and ponds that made up this landscape. Over time streams expanded their drainage areas by utilizing meltwater channels, cutting down through spillways, working their way headward through the thick drift, and by joining other streams.

1.1  80.55  Cross over gas pipeline.
0.15  80.7   Intersection with Ridge Road on right. Continue straight.
1.1   81.8   At stop sign turn right onto Deckertown Turnpike.
0.1   81.9   Cross over Big Flat Brook.

Thick till exposed in road cut.

Cross over Parker Brook. Outlet stream of Mashipacong Pond.

Road into Stokes State Forest on left.

Cross over Forked Brook.

Pass over crest of drumlin. Drumlins are found throughout Kittatinny Valley and on Kittatinny Mountain in two different settings. The first setting consists of multiple drumlins that lie in two areas underlain by very thick and widespread till. The first of these areas is a 1-to-2 mile wide belt that extends from the Delaware River to Culvers Gap. The second lies on Kittatinny Mountain north of Culvers Gap and extends northeastward toward New York. Well records and seismic refraction data (Witte and Stanford 1995; unpublished data on file at the N.J. Geological Survey) indicate that the overburden here is typically greater than 100 feet thick, and most of the drumlins do not have bedrock cores. The second setting consists of isolated drumlins located amongst areas of thin till. Well records show that many of these do have a bedrock core. In some places, drumlins have cores that consist of weathered pre-Wisconsinan material (Stanford and Harper 1985).

The direction of the long axes of most drumlins shows they were molded by ice that flowed across the regions southwest topographic grain. Drumlins located in the large swarm southwest of Culvers Gap have long axes that range from S25°W to S40°W, and drumlins on Kittatinny Mountain have long axes that range from due south to S15°W. A few drumlins have a compound shape, suggesting that they may have been molded by multiple directions of ice flow during the late Wisconsinan at different times.

Cross over unnamed brook.

Leave High Point State Park.

Four Corners. Intersection with New Road (Sussex County 675). Continue straight.

Low hills to right are ice-hole fillings (“a kame, by any other name is still a hill of sand and gravel”) that chiefly consist of deltaic foreset beds. This is a classic case of “inverted topography.”

Outcrop of Coeymans limestone on the left.

New Scotland outcrop on the left.

At stop sign, continue straight. Intersection with Clove Road.

At stop sign, cross Montague-Millville road and continue straight ahead on US 206 North toward Pennsylvania.

Milford Toll Bridge. Cross Delaware River and enter Pennsylvania. The floor of the valley, here know as the Minisink Valley, is covered by the head-of-outwash of a valley-train deposit laid down from the Augusta margin (Montague moraine). Except for a few small segments near the edge of the valley walls the moraine has presumably been eroded by meltwater in this area. The Delaware lies about 100 feet below the outwash plain and flows along a deeply incised straight and narrow channel. These deposits had been previously mapped by Crowl (1971) as kame terraces.

Toll booth ($1.00 for passenger vehicles). Note it only costs money to leave New Jersey.

Intersection of US 206 and US 209. Turn left (south) on US 209, following signs for
Dingmans Ferry, Bushkill, and Stroudsburg. The highway here follows the Montague valley-train terrace.

Enter the Delaware Water Gap National Recreation Area (technically we really haven’t left it) and begin descent onto the eroded surface of the valley train. Meltwater-terrace deposits in the valley generally lie more than 50 feet above the Delaware. Below this, you’ll find the postglacial alluvial terraces and the modern flood plain. In a few places the upper alluvial terrace (Qt3) has been eroded, revealing meltwater-terrace deposits. (See Witte, this guidebook, for a detailed description of postglacial terraces.)

Behind the trees to right is an abandoned borrow pit in shale-chip rubble derived from cliffs of Mahantango shale and siltstone above. This is first of many such pits you will see on the way back to Shawnee and on Day 2. The shale-chip (sharpstone) colluvium that forms a thick apron along the base of the steep slope was highly prized as road base during the 19th and early 20th century (Henn, 1975).

To right is another abandoned Mahantango shale pit. This pit and others in the immediate area were featured in Hoskins et al. (1983, p. 150-153) as the fossil collecting site for Pike County. More than 20 genera of fossils, including brachiopods, bivalves, trilobites, and cephalopods, could easily be found in the rubble here. Unfortunately, the pits are in the National Recreation area and the Park Service immediately shut down collecting after publication of the book. Much of the rubble material that was present here in the early ‘80’s has since been hauled away for various purposes.

The Mahantango cliff above, known appropriately as “The Cliff” (Figure 164) rises more than 400 feet above the valley floor. In this vicinity, on August 27, 1867, fourteen-year-old John Brink slipped and plummeted 160 feet from the top of “The Cliff”—and miraculously survived, thanks to being snagged in a tree at the end of his fall. His most serious injury was a fractured skull, which was somehow repaired with a silver dollar. Later in life, he became a Justice of the Peace in Milford, and was known as “Squire Brink” (Hen, 1975; Leiser, 2001).

Another Mahantango shale pit to right. The valley floor on the left is covered by postglacial alluvial terraces. The higher terrace is an eroded Qt3. The lower terraces and channels are part of the main Holocene terrace (Qt2) that covers large parts of the valley floor. The low channel near the road is a common feature that forms on the landward...
side of the terrace, pinned against the valley wall and the natural backslope of the terrace, and further eroded during large floods. Over time the Delaware may erode westward and deposit more proximal overbank sediment (levee facies) over the back channel sediment. This stratigraphy was recently revealed in a large excavation for restroom facilities downstream near Bushkill, where 12 feet of overbank sand and silt is underlain by 9 feet of organic-rich silt and clay.

0.4 91.6 Cross Raymondskill Creek, locale of the last two STOPS (11 and 12) of Day 2. (Details of route from here to mile 117.9 are on Day 2 road log.)

4.9 96.5 Traffic light at intersection with PA 739. Continue straight ahead.

11.7 108.2 Blinking light in Bushkill.

0.3 108.5 National Park Visitors’ Center to left.

0.3 108.8 Bear left onto Community Drive (Middle Smithfield Township 633).

0.05 108.85 Along Sand Hill Creek to right is an old, but well-preserved cobblestone wall, perhaps part of a former millrace.

0.05 108.9 To left at culvert is a cut in terrace material, sand at the base and cobble gravel above.

0.3 109.2 Low cut in intensely cleaved Onondaga Limestone to left at bend in road.

0.2 109.4 Schoonover Community House (Bushkill Outreach) to right. Just beyond on right are several low cuts in the Onondaga.

0.1 109.5 To the left is one of the numerous fens in this area.

0.05 109.55 In front of the house to the right is a handsome cobblestone wall. A little farther on are several low cuts in intensely cleaved Schoharie siltstone.

0.45 110.0 Stop sign. Turn left on River Road (SR 2028).

0.1 110.1 Esopus outcrop to right.

0.1 110.2 National Park Headquarters to left. Behind the building is another large fen.

0.1 110.3 Cut through cleaved Esopus siltstone.

0.05 110.35 Another fen to left. This is “fentastic”!

0.05 110.4 Ridge of Oriskany on right, followed by a hill held up by the Shawnee Island Member of the Coeymans Formation (measured section 12 of Epstein et al., 1967). The New Scotland Formation and Port Ewen Shale are covered between the Oriskany and Coeymans exposures.

0.1 110.5 Historical marker to the right reads:

DUCTH SETTLERS. First European settlers in this region were the Dutch, who came over the Old Mine Road, traveling from the Hudson to the Delaware. Crossing at Walpack Bend, they then used this road, oldest in Monroe County.

0.2 110.7 Outcrop of Decker Formation on right.

0.3 111.0 Ruins of lime kiln on right. The source of the limestone was a nearby quarry in the Bossardville Formation.

0.1 111.1 Parking area for John Turn Farm on left. Park exhibit reads:

The story of the John Turn farm and its owners illustrates the changing land-use and livelihoods in the Delaware Valley. John Turn, a carpenter and cabinetmaker, began farming here in 1815, building the main house in 1832. Turn and his wife, Julia Ann Shoemaker, had five children who helped with the farm chores. The eldest son, John Jr., after operating a raft on the Delaware River to Philadelphia, returned in 1844, and with his wife, Ency Dupue, continued farming here with their nine children. Of John Turn, Jr.'s seven sons, Samuel owned Turn's General Store in Bushkill, while Melchoir Depue built Turn Villa (Ridge View House), a summer resort about a mile away on today's Route 209. The youngest son, Charles, a prominent busines man in Stroudsburg.
converted the farm from subsistence agriculture to dairying. In 1945, the farm itself became a place for recreation when the Turn family sold the property to the Evangelical Lutheran Church. The church operated Camp Ministerium, a summer youth facility, for 30 years. In 1975, the land became part of Delaware Water Gap National Recreation Area.

0.1 111.2 Sharp bend in road. The road will now follow the base of Wallpack Ridge for the next two miles roughly paralleling the contact between the very poorly exposed Poxono Island Formation and Bloomsburg Red Beds.

0.4 111.6 Bloomsburg outcrop in gully on right.

0.9 112.5 Postglacial stream-terraces on left.

0.3 112.8 Zion Church Road on right. Rare exposures of Poxono Island Formation in ravine on right. Church and cemetery sit on outwash.

0.3 113.1 Hialeah Air Park on right. Flying area for model plane enthusiasts. Valley floor here is covered by broad terraces of Holocene age.

0.3 113.4 Saprolitic Poxono Island Formation in gully on right.

0.4 113.8 Entrance to Smithfield Beach on left. Small quarry in the Bossardville Formation on right.

0.4 114.2 Outwash underlies fields on left (part of the Zion Church valley train).

0.2 114.4 Historical Marker to right reads: SMITHFIELD CHURCH. Believed to be the first church built by Dutch Settlers in Pennsylvania, it stood below the road toward the river. The log structure housed the Dutch Reform congregation until they moved to the Old Stone Church in Shawnee ca. 1752.

0.6 115.0 More outwash on left (Zion Church valley train).

0.1 115.1 Massive outcrop on right is a reef within the Shawnee Island Member of the Coeymans Formation (Figure 165). The reef was visited during the 1967 Field Conference (Epstein and Epstein, 1967, p. 69-70) and the reef-off reef facies was figured by Epstein et al. (1967, p. 24)

Figure 165. A, Massive limestone that houses a stromotoperoid-coral Devonian reef in the Shawnee Island Member of the Coeymans Formation at milage 70.2. B, Scanned image of polished section showing corals (c) encrusted with stromatoporoids (s) as the major reef component. Occassional brachiopods (r) and debris of seashells and echinoderm fragments (d) make up the interstitial matrix.

0.4 115.5 Entrance to Hialeah Picnic area on left. The terrace here consists of outwash. Downvalley, lower terraces are postglacial age.

0.5 116.0 Outcrop of Decker Formation in ravine on right.
Exit Delaware Water Gap National Recreation Area.

Historical Marker to right reads:
*NICHOLAS DEPUY. First known settler in this region, 1727. His home, stockaded and garrisoned, became Fort Depuy of the French & Indian War, after 1755.*

Quarry in Bossardville on right.

Bear left at fork in road, village of Shawnee. Shawnee playhouse, a marvelous summer diversion, is on left.

Small water falls over deep mudcracks in southeast-dipping, overturned beds of the Bossardville Limestone to right (Figures 7, page 12).

Turn left into Shawnee on Delaware entrance.

Front of Shawnee Inn. End of Day-1 field trip.
Route Map for Day 2 of the 2001 Field Conference of Pennsylvania Geologists
**DAY 2**

<table>
<thead>
<tr>
<th>Miles</th>
<th>Int.</th>
<th>Cum.</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>Leave from circle in front of Shawnee Inn.</td>
</tr>
<tr>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
<td>Turn left onto River Road (SR 2028) at entrance to Shawnee Inn.</td>
</tr>
<tr>
<td>0.5</td>
<td>0.8</td>
<td>0.8</td>
<td>Intersection with Buttermilk Falls Road. Continue straight ahead.</td>
</tr>
<tr>
<td>1.5</td>
<td>2.3</td>
<td>2.3</td>
<td>Cross Brodhead Creek.</td>
</tr>
<tr>
<td>0.2</td>
<td>2.5</td>
<td>2.5</td>
<td>Pass under I-80.</td>
</tr>
<tr>
<td>0.1</td>
<td>2.6</td>
<td>2.6</td>
<td>Stop sign. Turn right onto ramp to I-80 West.</td>
</tr>
<tr>
<td>0.5</td>
<td>3.1</td>
<td>3.1</td>
<td>Merge with I-80 West.</td>
</tr>
<tr>
<td>0.1</td>
<td>3.2</td>
<td>3.2</td>
<td>Deep road cut exposing a shallow syncline in the Port Ewen Shale (base), Shriver Chert, and Ridgeley Sandstone. At the top of the cut are two big joint blocks of Ridgeley sandstone held in place over the highway by five steel cables. The eastern block is separating from the cliff face by an open vertical joint striking N60°E.</td>
</tr>
<tr>
<td>0.1</td>
<td>3.3</td>
<td>3.3</td>
<td>Cross Brodhead Creek. The Ridgeley forms a high natural cliff to the south.</td>
</tr>
<tr>
<td>0.3</td>
<td>3.6</td>
<td>3.6</td>
<td>On the right along the on-ramp of Exit 309 the Edgecliff Member of the Onondaga Limestone is exposed. The road then cuts through late Wisconsinan glacial till.</td>
</tr>
<tr>
<td>1.0</td>
<td>4.6</td>
<td>4.6</td>
<td>Bear right onto ramp at Exit 308 (East Stroudsburg).</td>
</tr>
<tr>
<td>0.2</td>
<td>4.8</td>
<td>4.8</td>
<td>Stop sign. Turn left onto Prospect Street.</td>
</tr>
<tr>
<td>0.1</td>
<td>4.9</td>
<td>4.9</td>
<td>Turn right onto Forge Road immediately after crossing I-80 (opposite cemetery).</td>
</tr>
<tr>
<td>0.4</td>
<td>5.3</td>
<td>5.3</td>
<td>Stop sign. Turn left toward East Stroudsburg sewage treatment plant, then turn left onto dirt road just before gate. (Buses proceed right down Lincoln Avenue 0.1 mi, then turn left into East Stroudsburg recycling center where they park. After STOP is completed, buses return for pick up at sewage treatment plant. )</td>
</tr>
</tbody>
</table>

**STOP 7. EAST STROUDSBURG RAILROAD CUT: SCHOHARIE FORMATION AND ONONDAGA LIMESTONE—STRATIGRAPHY AND STRUCTURE.**

Leaders: Charles A. Ver Straeten, Jon D. Inners, and Jack B. Epstein.

This oft-described anticlinal rock cut on the Norfolk Southern Railroad at East Stroudsburg (Figure 166) provides both the most complete exposure of the Early/Middle Devonian-age Onondaga Limestone in the Water Gap region and an instructive local “snapshot” of the complex geology of Godfrey Ridge (see Epstein, 1989). Originally blasted out along the route of the old Delaware, Lackawanna & Western Railroad (from Scranton, PA, to Hoboken, NJ) in the early 1850’s, both its stratigraphy and structure were initially described by I. C. White of the Second Pennsylvania Geological Survey (White, 1882, p. 265-266). E. M. Kindle (1912) and Bradford Willard (1936, 1939) also made reference to the cut, the former noting several typical “Onondaga” fossils and the latter introducing the name Buttermilk Falls for

Figure 166. Location map for STOP 7.
the limestone here referred to as Onondaga. However, it was a series of much more recent studies that provided most of the information for this STOP description. Epstein (1970, 1984) described in detail the structure and stratigraphy of the cut; Inners (1975) studied its stratigraphy and paleontology; and Ver Straeten (1996a) made a detailed stratigraphic analysis of the Onondaga Limestone and correlated individual members and beds with units in the Onondaga of New York State. All three independently measured the Onondaga section on the south side of the cut, the measurements being made before the upper 27 ft (8.2 m) of the formation were removed by construction of Forge Road across the railroad at the northwest end of the cut.

The 267 ft (81.5 m) that comprise the Onondaga Limestone, or Formation, in the Stroudsburg area represents the area of greatest thickness for Onondaga and equivalent strata known in the basin, with the exception of a more clastic-dominated and volcanic tuff-rich section adjacent to Massanutten Mountain, in northern Virginia (Dennison, personal communication; Rickard, 1989; Ver Straeten, unpublished data). This is part of a regional trend of greater thickness and accommodation space through Lower to Middle Devonian strata (Rickard, 1975, 1989), focused in a Tri-States area sub-basin of the larger Appalachian foreland basin. (NOTE: The measurements used in this STOP description are those of Inners, 1975; both Epstein, 1984, and Ver Straeten, 1996a, record 272 ft/82.9 m.)

STRATIGRAPHY

Nomenclature

A number of distinctive marker beds occur through the strata in the railroad cut at East Stroudsburg and throughout the Stroudsburg area. These include K-bentonite beds, a coral-rich biostrome, shale-rich intervals, a pair of dark gray to black shales split by a thin limestone bed, and other unique beds. Detailed study of the Onondaga Limestone and equivalent strata across the Appalachian Basin by Ver Straeten (1996a, b; this guidebook, p. 35) shows that these same marker beds, along with member-level and finer stratigraphic subdivisions are widely correlatable (see Figure 26, p. 42). Correlation of the East Stroudsburg railroad cut with the Onondaga Limestone in its type area indicates that the Buttermilk Falls and the Onondaga Limestones of eastern Pennsylvania and New York, respectively, are exact equivalents; the four member-level subdivisions of each also represent almost exact correlatives, excepting the boundary between the upper two members, in each area.

Recognizing the synonymy of the units, and following the North American and International stratigraphic codes, Ver Straeten and Brett (in preparation) will propose that the older nomenclature (Onondaga Limestone, or Formation, of Hall, 1839; Edgecliff, Nedrow, Moorehouse, and Seneca Members) be applied to the strata in eastern Pennsylvania. The terms Buttermilk Falls Limestone of Willard (1936); Foxtown, McMichael, Stroudsburg Members of Epstein (1984); and Echo Lake Member of Inners (1975) will be abandoned. This nomenclature is followed in this guidebook.

Biostratigraphy and Age

Different North American stage-level terms (e.g., Onesquethaw Stage of Cooper et al., 1942; Southwood and Cazenovia stages of Rickard, 1975) have been applied in the past to the Onondaga Limestone and associated strata in the Appalachian Basin. Recognizing the increasingly better correlation of Appalachian Basin and global Devonian strata, the use of stages (Emsian and Eifelian) utilized by the International Subcommission on Devonian Stratigraphy is recommended and applied here. The global GSSP ("Global Stratotype Section and Point") for the Emsian-Eifelian boundary is in the Eifel Hills of southwestern Germany.

The base of the Onondaga Limestone, or Formation, in New York was long considered to represent the Lower-Middle Devonian (Emsian-Eifelian stages) boundary. However, recent definition
of the global Emsian-Eifelian boundary by the Subcommission on Devonian Stratigraphy has raised doubts as to its position in Eastern North America. At present the position of the boundary is poorly defined, occurring somewhere between the base and top of the Edgecliff Member. Conodont work is presently underway in an attempt to resolve this issue. Higher up in the Onondaga, Klapper (1971) reports two conodont zones within the post-Edgecliff Onondaga, the *patulus* and *costatus* zones. The latter includes the upper part of the Nedrow, and the whole of the Moorehouse and Seneca Members. So, at present, the Edgecliff Member is considered to be uppermost Emsian and/or lowermost Eifelian; the Nedrow to Seneca members are of early Eifelian age.

The Tioga B K-bentonite, at the base of the Seneca Member, was dated by Roden et al. (1990) at 390.0 ± 0.5 Ma, using U-Pb dating techniques on monazite crystals. A more recent date of 391.4 ± 1.8 Ma is reported by Tucker et al. (1998) for a K-bentonite from the Tioga "Middle Coarse Zone" of Dennison (1961) from Wytheville, VA. New work by Ver Straeten (2001, in preparation), however, shows that the Middle Coarse Zone underlies and is older than the Tioga A-G zone (see below).

### Facies

Figure 167 displays an idealized onshore-offshore facies transect for the Onondaga Limestone and equivalents across the Appalachian Basin. The predominance of facies 4-7 (fossiliferous packstones, chert-rich limestones, wackestones, and calcisilt facies) in the railroad cut and the Stroudsburg area is characteristic of an intermediate water depth position in the basin. High subsidence rates in the Stroudsburg/Tristates area relative to other parts of the basin is indicated by greater thickness of Onondaga strata. However, regional maintenance of water depth trends directly correlative to a basinwide relative sea level curve indicates high sediment productivity rates and steady state infilling that kept pace with the formation of excess accommodation space.
**Detailed Description of Railroad Cut**

**Schoharie Formation**, 58 ft/17 m +

**Saugerties-Aquetuck Members, undivided**, 58 ft/17.7m +

The thick-bedded, partly calcareous, medium-dark-gray, burrowed, sandy to argillaceous siltstone to very fine sandstone in the core of the overturned anticline constitutes approximately the entire thickness of the “upper” Schoharie in the Field Conference area. As at other nearby localities (particularly on U.S. 209 near Buttermilk Falls [mile 8.5 of road log]), beds in the upper 15-20 ft (4.6-6.0 m) of the formation contain considerable fine sand—a reflection of the southwestward lateral change of the upper part of the Schoharie Formation into the middle part of the Palmerton Sandstone (Epstein, 1970, 1984; Inners, 1975; Ver Straeten, 1996a, b). In fact, the uppermost bed of the Schoharie in the railroad cut and other local outcrops consists of a 3 ft/0.9 m-thick, very resistant, calcareous, fine-to medium-grained sandstone. The contact is abrupt (Figure 168), and clearly defined on both sides of the cut.

**Onondaga Limestone**, 267.5 ft/81.5 m

**Edgecliff Member**, 82 ft/25 m (= Foxtown Member of Epstein, 1984.)

**Unit 1 (0 to 50 ft/0 to 15.3 m)**

The lower part of the Edgecliff Member consists primarily of thin- to medium-bedded (in irregular to lenticular beds 1 in. to 2 ft thick), medium-gray (N5) to medium-dark-gray (N4), medium-light-gray (N6) weathering, fine- to medium-grained, fossiliferous limestone, interbedded with medium-dark-gray (N4), calcareous, evenly bedded shale and siltstone in beds 1 in. to 1 ft thick and grayish-black (N2) to dark-gray chert. Chert occurs in irregular nodules 0.5 to 6 in. long. The most conspicuous fossils are large crinoid “columnals” (probably holdfast fragments) up to 1.5 in. in diameter (Figure 169); solitary rugose corals are also relatively common, especially in the lower 2.5 ft/0.8 m where they may represent the equivalent of the non-cherty, coral bioherm facies that is found in basal Edgecliff strata all along the New York outcrop belt. Ostracodes are abundant in thin sections of Unit 1.
limestones. The base of the unit is marked by a 1-ft thick, medium- to very coarse-grained limestone in abrupt contact with the Schoharie Formation (Epstein, 1984, p. 30).

**Unit 2 (50 to 82 ft/15.3 to 25.0 m)**

The upper 32 ft/9.8m of the Edgecliff is somewhat thinner bedded (beds up to 1 ft/0.3 m thick) and conspicuously more cherty (at least 50 percent) than the lower part. Limestone occurs mostly in pods 2 to 6 in. in diameter, while the shale and siltstone beds are 1 to 2 in. thick. Large crinoid “columnals” appear to be absent, but ostracodes continue to be abundant. (A 6-in.-wide, broken-out crevasse along bedding occurs about 7 ft [2.1 m] above the base of the unit.) The contact with the overlying Nedrow Member is transitional through a 4-ft/1.2-m interval of cherty limestones and calcareous shales, which is well exposed to the south in the East Stroudsburg Sewage Treatment Plant.

Overall, the Edgecliff Member at East Stroudsburg can be subdivided into three medial-scale, deepening- to shallowing-up cycles, the upper part of the third extending into overlying strata of the Nedrow Member. These three cycles (with bases at 0 ft/0 m, 21 ft/6.4 m, and 63 ft/19.25 m) are recognized in Edgecliff strata across much of the Appalachian Basin, as is one of two thin K-bentonites that occur in slickensided crevices at the base of and in the middle of the upper cycle.

**Nedrow Member, 41 ft/12.5 m (= McMichael Member of Epstein, 1984.)**

**Unit 3 (0 to 41 ft/0 to 12.5 m)**

As in central New York State, the Nedrow Member constitutes the most conspicuous shaly interval in the Onondaga of the Stroudsburg area (Figure 170). It is composed predominantly of medium-gray to medium-dark-gray, medium-gray weathering, calcareous, locally somewhat silty shale in beds 2 in. to 1 ft thick. Interbedded with the shale are 1 to 3 in.-thick beds, lenses, and nodules 1 to 3 in. thick of medium-gray, fine-grained, argillaceous limestone. Ostracode, brachiopod, coral, and crinoid fragments are common in the limestone, the only clearly identifiable fossils being *Pseudoatrypa* and small solitary rugose corals. A thin crevice in the middle of the Nedrow, visible on the north side of the tracks is a thin, widely correlatable K-bentonite layer (it is covered by a recent debris slide on the south side of the tracks).

The uppermost 6 ft/1.8 m of the Nedrow Member is marked by dark gray to black shales with little to no carbonate, and an intermediate, highly argillaceous limestone bed. This unit correlates with a widespread pair of black shales separated by a thin limestone that occurs at the top of the Nedrow Member throughout deeper portions of the basin (Brett and Ver Straeten, 1994; Ver Straeten, 1996a). In shallower areas it is recognized by distinctive deeper water facies relative to other Nedrow and Moorehouse strata below and above. The contact with the overlying Moorehouse Member is gradational and marked by the upward appearance of chert and the disappearance of argillaceous beds.
The middle to upper part of the Edgecliff Member (56 ft/17 m) and the lower approximate 2/3 of
the Nedrow Member (28 ft/8.6 m) are best seen in an outcrop on the north side of the East Stroudsburg
Sewage Treatment Plant, approximately 200 ft/60 m south of the railroad cut.

Moorehouse Member, 113 ft/34.4 m (= lower part of the Stroudsburg Member of Epstein, 1984.)

Unit 4 (0 to 55 ft/ 0 to 16.8 m)

The lower part of the Moorehouse Member consists of thin- to medium-bedded (in beds, pods,
and nodules 1 in. to 1 ft. thick), medium-gray to medium-dark-gray, light-gray to medium-light-gray
weathering, fine- to medium-grained, locally argillaceous, cherty limestone. Dark-gray (N3) to grayish-
black chert composes 15 to 25 percent of unit, occurring in irregular pods, lenses, and discontinuous beds
0.25 to 8 in. thick. Fossil fragments are relatively uncommon.

Two thin, calcite-filled crevices 9 and 12 ft (2.7 and 3.7 m)
above the base of Unit 4 represent a widely recognized pair of thin K-
bentonites in the lower part of the
Moorehouse Member (Figure 171). An additional correlatable K-
bentonite associated with a
slickensided calcite vein 51 ft/15.7 m
above base of Unit 4 (=7.5 ft/2.3 m
below top of Unit 4) underlies the
succeeding shaly Unit 5 across much
of the Appalachian Basin.

Unit 5 (55 to 59 ft/16.8 to 18.0 m)

This 4 ft/1.2 m-thick shaly interval is the “false Nedrow” of
eastern and central New York (Figure
172). As in other areas of the basin
caracterized by intermediate depth facies, it consists mostly of medium-
gray, silty, calcareous shale
containing lenticular beds and nodules
(to about 2 in. [5 cm] thick) of
medium-gray to medium-light-gray,
medium-light-gray weathering, fine-
to medium-grained, sparsely
fossiliferous limestone. Some
limestone nodules have thin cherty rims.

In deeper facies across the
basin (e.g., central Pennsylvania) the
"false Nedrow" consists of a thin
interval of black shale. In shallower areas of the basin, it is characterized by distinctly finer-grained limestones of deeper water aspect compared to under- and overlying strata.

**Unit 6 (59 to 113 ft/18.0 to 34.4 m)**

The upper half of the Moorehouse Member is composed mainly of thin- to medium-bedded (in beds 2 in. to 1 ft. thick), medium-dark-gray, medium-light-gray to light-gray weathering, fine- to medium-grained, very cherty, fossiliferous limestone. The chert, constituting 30 percent or more of the unit, occurs in elongate, discontinuous pods and ragged nodules some of which are 6 in. or more thick perpendicular to bedding. Overall, strata of Unit 6 show a shallower water character (grain size, bedding thickness, fauna) than underlying strata of the Nedrow and Moorehouse members. A 2-in.-wide crevice halfway up through Unit 6 is the Tioga A K-bentonite bed of Smith and Way (1983) and Way et al. (1986).

The false Nedrow and overlying strata of the Moorehouse Member, along with succeeding lower strata of the Seneca Member (= units 5 to 8), are better seen in a low outcrop approximately 60-100 ft./18-30 m south of the railroad cut. Part of Unit 6 and the overlying Seneca Member are also well exposed at STOP 8.

**Seneca Member, 31.5 ft/9.6 m (= uppermost part of Stroudsburg Member of Epstein, 1984.)**

**Unit 7 (0 to 1 ft/0 to 0.3 m)**

Marking the base of the Seneca Member is a 10 in. to 1 ft thick bed of greenish gray (5GY 6/1) to medium-dark-gray, medium-light-gray to moderate-yellowish-brown (10YR 5/4) weathering, non-calcareous, tuffaceous siltstone and very fine-grained sandstone (Figure 172). A thin section of the tuff bed from a nearby cut on I-80 at Stroudsburg reveals an abundance of ragged biotite flakes and finely disseminated pyrite (Inners, 1975, Figure 48). This is the principal altered volcanic tuff or K-bentonite (B) of the famous Tioga ash-bed zone (Smith and Way, 1983).

Recent study of the Tioga K-bentonites across the Appalachian Basin (Ver Straeten, 2001) shows that the Tioga A-G interval of Smith and Way (1983) does not correlate to, but overlies, the Tioga Middle Coarse Zone of Dennison (1961); for further discussion see Ver Straeten, this guidebook, p. 35). Based on overall and phenocryst-specific grain-size analyses of the Tioga Middle Coarse Zone, Dennison and Textoris (1978) and Dennison (1986) project a source in the vicinity of Fredericksburg, northeastern Virginia. In contrast, however, qualitative observation of the Tioga B bed across the basin appears to indicate that the phenocryst-rich, silt- to sand-sized lithology of the bed in the Stroudsburg area represents the coarsest facies of the Tioga B bed across the Appalachian Basin. This may point to a different, and more northwestern, source for the Tioga B bed in the Acadian magmatic arc proximal to the eastern Pennsylvania/southeastern New York region.

**Unit 8 (1 to 9.5 ft/0.3 to 2.9 m)**

The beds directly above the Tioga B are mainly thin- to medium-bedded (2 in. to 1 ft. thick beds), medium-dark-gray, light-gray weathering, medium grained, very cherty, fossiliferous limestone. Grayish-black to dark-gray chert, occurring in nodules 1 to 4 in., is very abundant. (Only the lower two or three feet of this unit are still present, the rest having been removed by road construction.) On the north side of the cut, a fault appears to repeat the Tioga B and overlying strata. The exact structural configuration of this part of the cut is—to say the least—unclear!

**Unit 9 (9.5 to 31.5 ft/2.9 to 9.6 m)**

Before road construction in the early or mid-1990’s, the distinctive coquinitic “Echo Lake” beds of Inners (1975) were well exposed here. They are now completely missing as a result of road construction. (But fear not, we will get a good look at them at STOP 8.) You will have to take our word
for it that once present at the top of the Seneca at this spot were about 22 ft/6.7 m of medium-gray to medium-light-gray, medium- to very coarse-grained, cherty, fossiliferous limestone in beds 2 in. to 1-foot thick—some of which were coquinites composed mainly of chonetid-brachiopod shells. Dark chert similar to that in the underlying unit was somewhat less common in these upper beds. (It may have constituted 50 percent of the interval directly above the Tioga.) The chonetid brachiopod in the coquinites is *Hallinetes*. Other fossils found here—mainly in the shell beds—include *Acrospirifer varicosus*, *Tentaculites bellulus*, and small rugose corals (Inners, 1975). The contact with the overlying Union Springs Formation of the Marcellus Group is not exposed, but it is probably just above the top of this unit.

### Significant Events, Processes, and Sequence Stratigraphy

Upper Lower and Middle Devonian strata of the East Stroudsburg railroad cut record several key events in the geologic history of the region, associated with tectonic, eustatic, and paleobiological processes active at the time of deposition.

Strata of the Schoharie, Onondaga, and Union Springs Formations record changes in the Appalachian Basin associated with two separate tectonically-active to quiescent stages in the Devonian to Mississippian Acadian orogeny (Ettensohn, 1985; Ver Straeten 1996a). Ettensohn (1985) and Ver Straeten (1996a) interpreted the post-Oriskany succession of basinal, dark gray mudstones and sandstones (Esopus Formation), mixed clastics and carbonates (Schoharie Formation), and succeeding Onondaga carbonates to record a period of active orogenesis, overdeepening of the Appalachian foreland basin, and input of clastic sediments, followed by increasing tectonic quiescence and a return to carbonate deposition over a period of approximately 18 million years (based on a new Devonian time scale of Tucker et al., 1998). Upper Schoharie and Onondaga strata in the railroad cut record the later, relatively quiescent stage of Acadian Tectophase I, immediately prior to the onset of a second phase of active tectonism, marked by Marcellus black shales (Union Springs Formation) that closely overlie the top of the Onondaga Formation (covered at STOP 7).

In the northeastern part of the Appalachian Basin (eastern New York), the transition from the widespread, shallow marine sandstones and limestones (Oriskany Formation and equivalents) to basinal dark gray mudstones and shales (Esopus Formation) is marked by a K-bentonite-rich interval (Sprout Brook K-bentonites). Similarly, a large number of K-bentonites are concentrated in the upper part of the Onondaga into the lower part of the Marcellus black shales (Bakoven Member). These appear in both cases to mark an increase in volcanic activity in the Acadian orogenic belt during, as previously stated, times of increased tectonism at the beginning of Acadian Tectophases I and II (Ver Straeten, 1996a).

Litho- and biofacies trends through the Schoharie and Onondaga Formations locally also reflect changes of relative sea level through the upper Emsian and lower Eifelian. General lithologic coarsening up and overall shallowing upward biofacies trends in upper Schoharie strata in the railroad cut indicate shallowing up to the basal coral-rich strata of the Onondaga (or the top sandstone of the Schoharie). An overall deepening through the overlying cherty limestones (Edgecliff Member) and mixed shale and limestones (Nedrow Member) culminates in maximum depths in the top "black beds" of the Nedrow. A return to chert-rich limestone facies through the lower to middle Moorehouse marks another upward coarsening/bed thickening trend to a series of medium-to thick-bedded limestones above the "false Nedrow shale." The occurrence of a relatively shallow water fauna in the upper Moorehouse (including *Paraspirifer acuminatus* and other forms) records a position of sea level lowstand prior to another reversal of both lithologic and faunal indicators, which marks a major deepening into the overlying black Bakoven shale, and on a larger scale, into the second major pulse of clastics of the Middle Devonian Hamilton Group.
These relative sea level trends can be discussed in the context of the sequence stratigraphic model (see also Ver Straeten, this guidebook, p.35). Examining the local strata, a relative water depth curve for the railroad cut can be established. Basinwide comparison of these trends (Ver Straeten, 1996a, 2001) permit construction of a relative sea level curve from which a sequence stratigraphic framework can be interpreted.

The shallowing up through the upper Schoharie in the East Stroudsburg railroad cut represents late highstand systems tract (LHS) deposition in the later part of a sequence that comprises the whole of the Schoharie Formation. The top (1 ft/0.3 m-thick) sandstone of the Schoharie and basal coral-rich limestone of the Onondaga may be interpreted as lowstand or basal transgressive systems tracts (LST or TST). LST is generally not recognized in shallower portions of an epicontinental sea (Brett and Baird, 1996); however, a sequence-bounding unconformity, noted widely at the base of the Onondaga in New York and towards Harrisburg, PA, is not recognized in the railroad cut.

Succeeding strata of the Edgecliff and Nedrow members represent the TST of a lower to middle Onondaga sequence. The so-called Nedrow black beds represent the position of maximum flooding basinwide. An initial slow, then more rapid shallowing through the overlying lower to middle Moorehouse represent early and late HSTs, with the LST and TST of the succeeding upper Onondaga-Union Springs sequence marked by upper Moorehouse (LST) and Seneca and Bakoven (TST) strata.

The interplay of tectonics and eustasy in a foreland basin and their effects on relative sea level change can be difficult to tease apart, as seen in vigorous debates over the recent years by researchers who favor one process over the other. However, through relatively high resolution correlation and subdivision of Emsian and Eifelian strata across the entire Appalachian Basin outcrop (Ver Straeten 1996a, 2001), and preliminary comparison with the European and North African successions, it appears that the varied influences of tectonism and eustasy can begin to be picked apart. In intervals where conodont biostratigraphy is of a high enough resolution for comparison between the Appalachian Basin, Europe, and North Africa, patterns of sea level change through the so-called "third order" sequences appear to be coordinated in time. Overall trends do not always match; for example, where the Esopus sequence represents a major transgression to depth greater than in the subsequent Schoharie and lower to middle Onondaga sequences, the record in much of Europe and North Africa show a greater deepening in a Schoharie-equivalent sequence relative to underlying Esopus-equivalent strata. This reflects greater subsidence of the Appalachian foreland basin during the early (Esopus-age), tectonically-active stage of Acadian Tectophase I. Other flexural effects of Acadian tectonism across the Appalachian Basin include the migration of a peripheral bulge during late Schoharie and Onondaga time, which resulted in the formation of pinnacle reefs in deeper, central portions of the foreland basin (Ver Straeten and Brett, 2000). One of the key paleobiological events in the Phanerozoic was the Late Devonian (Frasnian-Famennian) Mass Extinction, when as much as 70% or more of all species of marine animals became extinct (McGhee, 1996). Several lesser extinction-radiation events also occurred through the Devonian, including the lower Eifelian "Chotec Event," reported from the Devonian around the world (Chlupac and Kukal, 1986; Boucot, 1990; Walliser, 1996). The Chotec Event, which marks the extinction and radiation of different goniatite cephalopods and tentaculite-like dacyroconarids, occurs globally near the base of the Polygnathus costatus costatus conodont zone. In the Appalachian Basin, the base of the costatus zone occurs at the top of the Nedrow Member (Klapper, 1971, 1981). Comparison with sections in central Europe and Morocco indicate that the Nedrow black beds may represent this global bio-event in eastern North America.
STRUCTURE

The fold at STOP 7 is typical of folds in lithotectonic unit 2 of Epstein and Epstein (1967, 1969). This sequence of rocks, which extends upward from near the base of the Upper Silurian Poxono Island Formation to the lower part of the Middle Devonian Marcellus (i.e., Union Springs) Formation, is approximately 1900 ft (575 m) thick in the Delaware Water Gap area. More resistant units in the upper part (particularly the Esopus Formation and the Oriskany Group) underlie the crest of the long series of en echelon ridges, 900 to 1000 ft high, that extend from Saylorsburg, PA, past Stroudsburg and Bushkill into New Jersey (Cherry/Godfrey/Walpack Ridge) (Epstein, 1989).

Fold structures in lithotectonic unit 2 are typically third-order folds (Nickelsen, 1963), averaging 700 ft (215 m) in wavelength and 250 ft (75 m) in amplitude. The folds vary in style, from concentric, flexural and passive slip to similar flow folds, and are typically asymmetrical. They form en echelon trains, die out over short distances, and vary in direction and degree of plunge along strike. These latter characteristics account for the serpentine outcrop pattern displayed by this package of rocks on geologic maps. Northwest movement of the entire stratigraphic sequence is indicated by bedding-plane slippage, wedging, and faulting. Shortening of about 15% more than the shortening in underlying lithotectonic unit 2 due to this northwest translation has been computed from the folds in Godfrey Ridge (Epstein, 1969).

The top and bottom of lithotectonic unit 2 are apparently bounded by regional decollements, or, at least, local zones of decollement. These discontinuities are at or just above the very poorly exposed contact of the Poxono Island Formation with the underlying Bloomsburg Formation and in the lower part of the Marcellus shale. The upper decollement is well documented and will be seen at STOP 8B.

Figure 173 is a generalized cross section though the fold in the railroad cut at STOP 7. Bedding on the south limb dips a relatively uniform 20-30°SE; but on the overturned north limb, dips vary from 88°SE at base of the Seneca Member at the northwest end of the cut to mostly 50-60° within the underlying Moorehouse and Nedrow Members, and then back to 80°SE to vertical as the axis is approached. Cleavage dips a uniform 40-50°SE on the north limb, but shows considerable variation on
Figure 174. Stereonets of bedding (A) and cleavage (B) at STOP 7. A: solid dots = bedding planes; open circles = bedding-plane-slip horizons; axis plunges S40°W at 10°. B: open triangles = north limb; solid triangles = south limb; axis plunges S37°W at 10°.

Figure 175. Geologic map (A) and cross section (B) in vicinity of STOP 7 from Epstein (1989).
the south limb—possible due to the misidentification of some closely spaced joint fractures as cleavage. Interpretation of stereonets of bedding and cleavage indicate that the fold axis plunges 10° toward S40°W (Figure 174A, B). At least ten calcite-mineralized bedding-plane-slip horizons occur on the north limb (Figure 174A), indicating a flexural-slip mechanism of fold propagation. Some of these slip horizons seem to be localized along K-metabentonite beds.

Epstein (1989; Figure 175A, B) maps the fold at STOP 7 as doubly plunging, with the fold axis trending S45°W. Wavelength is about 1000 ft (300 m), and amplitude about 200 ft (60 m). The maximum amplitude of the fold occurs in the railroad cut, and the fold dies out to both the southwest and the northeast within a half a mile in each direction.

Leave STOP 7, turning right onto Forge Road.

0.4 5.7 Stop sign. Turn left onto Prospect Street, then immediately right onto ramp to I-80 East.
0.2 5.9 Exposures of Wisconsinan till in bank of on-ramp on right.
0.1 6.0 Merge with I-80 East.
0.7 6.7 Bear right onto US 209 at Exit 309 (Marshalls Creek).
0.2 6.9 Abandoned quarry in woods on right and adjacent railroad cuts contain the Onondaga Limestone, parts of the Schoharie Formation, and the Esopus Formation (Epstein, 1984).
0.1 7.0 Cross over I-80.
0.6 7.6 Behind the Dairy Queen to right is a large excavation in deeply weathered calcareous siltstone of the Schoharie Formation. The rock is leached of its calcium carbonate to depths of at least 20 feet, unique in this area where most bedrock has been stripped of its weathered regolith by Wisconsinan glacial scouring. This hill has escaped such scouring. Bedding at the locality is N47°E/22-40° SE. A pervasive cleavage cuts the rock: N 51°E/ 72° SE. A thrust fault—striking N85°E and dipping 66° SE—and a bedding plane thrust contort the cleavage (Figure 176).

0.55 8.15 Cleaved, Taonurus-bearing Esopus silty shale in cut to left.
0.2 8.35 Buttermilk Falls Road to right.
0.15 8.5 Cut in south-dipping lower Onondaga Limestone, Schoharie Formation, and upper Esopus Formation (see Figure 29, page 55). Buttermilk Falls on Marshalls Creek, the type locality of the Buttermilk Falls (=Onondaga) Limestone, is about 0.1 mile to the east of here.
0.4 8.9 Cross Marshalls Creek. Visible just upstream of here is a picturesque stone arch bridge (built 1910) on the old “creek” road. To the right on the northeast side of the creek is a ragged borrow pit in the Port Ewen Shale.
0.2 9.1 Deep road cut through the Port Ewen Shale. On the west side of the road is a 32-foot-long erratic of cherty limestone sitting on the shale (see Figure 42, page 67). The erratic was derived from the Edgecliff Member of the Onondaga Limestone, which has supplied scattered large erratics like this one throughout the area (see White, 1882, p. 46-48). A
large area of this unit was initially mapped as bedrock (i.e., Buttermilk Falls Limestone) in “Helderberg” terrain during initial mapping by Epstein. This required a fantastic structural interpretation, until its glacial origin was recognized.

0.2 9.3 Pocono Wild Animal Snake Farm to left. Partially buried on the slope to right are large blocks of shaly and silty limestone that have broken off Port Ewen and Shriver ledges near the top of the high ridge to the east and moved by gravity and gelification to their present positions.

0.3 9.6 Cut through highly fossiliferous Port Ewen shale.

0.2 9.8 Another deep cut, this one through the interbedded shales, limestones, and cherts of the Port Ewen/Shriver interval. Mineralized bed-parallel faults are common.

0.2 10.0 And still another deep cut in Port Ewen shale/Shriver chert.

0.3 10.3 Cross Marshalls Creek. Just upstream (to right) of here is an Amerind chert “quarry” in the Ridgeley Sandstone, recently discovered as a result of archaeological investigations for relocation of US 209 (P. La Porte, personal communication, 2001).

0.1 10.4 Outcrop of the cleaved siltstone of the Gumaer Island Member of the Schoharie Formation to left

0.1 10.5 To left is Riccobono’s “Quarry in the Schoharie,” operated by Route 209 Enterprises, a subsidiary of Haines & Kibblehouse, Inc. (See Inners and Ver Straeten, this guidebook, p.61).

0.3 10.8 Low cut through lower Onondaga Limestone (Edgecliff Member).

0.1 10.9 Traffic light. Turn right, following US 209 North.

0.1 11.0 Traffic light at PA 402 intersection; continue straight ahead. To right in the shopping center and northeast along the highway for about 0.5 miles are numerous cuts in an east-west ridge of Onondaga Limestone.

0.1 11.1 Road to right leads to an Amerind archaeological site along the proposed new route of US 209 (P. La Porte, personal communication, 2001).

0.3 11.4 On the ridge to right are numerous small excavations made by American Indians in extracting black chert from northwest-dipping ledges of Onondaga Limestone (P. La Porte, personal communication, 2001).

0.3 11.7 Oak Grove. Directly to the south (right) of here (behind Wendy’s) is a large kettle pond, at the east end of which is “Leap’s Bog,” site of the discovery of the Marshalls Creek Mastodon in 1968 (see Hoff, this guidebook, p. 146). The high ground between Wendy’s and the pond is an elongate kame that extends eastward from the bedrock ridge noted above. Left of the highway is the preglacial drainage divide of the Echo Lake lowland.

1.0 12.7 Hollow Road on right. Middle Smithfield School, just past the road on right, rests on glacial outwash (elev. 525 ft) laid down from the Echo Lake recessional position.

0.5 13.2 Old Church and cemetery to right (just past 209 Diner on left and Muller’s Diner on right).

0.9 14.1 Traffic light at Foxmoor Village. Continue straight ahead.

0.4 14.5 Middle Smithfield Church on left, sits on Echo Lake outwash. Low area west of church is a large ice-block depression. Erosional channels cut in outwash at the head of the lowland may have served as glacial lake sluiceways for meltwater dammed between the glacier’s margin and Echo Lake outwash. These lake outlets operated until the glacier’s margin retreated north of Bushkill, opening up drainage into the main stem of the glacial Delaware River. See STOP 9 for a detailed discussion on the deglaciation of the Echo Lake lowland.
Caesar’s Pocono Palace to left.

Small quarry in field to left contains calcareous siltstone of the Stony Hollow Member in the lower part of the Marcellus Shale. The rock is similar to the Schoharie Formation. Bedding dips 6° NW and a well-developed cleavage dips 56° SE (Figure 177). The dark gray shale of the Union Springs Member underlies these rocks, and is exposed at the next stop (STOP 8) where it is sheared along a bedding fault.

Pocono Indian Museum to left.

Turn right into entrance at Fairway.

Turn right onto road parallel to high tension line.

Turn right at Fairway Ridge sign.

Park along road at broad grassy area just beyond limestone outcrops.

STOP 8. FAIRWAY AND US 209 SHALE PIT: UPPER ONONDAGA LIMESTONE, UNION SPRINGS FORMATION, AND BASAL UNION SPRINGS DECOLLEMENT.

Leaders: Charles A. Ver Straeten, Donald H. Monteverde, and Jon D. Inners.

INTRODUCTION

STOP 8 consists of two parts: STOP 8A includes several rock cuts at Fairway on the southeast side of US 209, while STOP 8B is a small shale pit on the northwest side of US 209 about 500 feet to the west (Figure 178). STOP 8A is on property belonging to Resorts USA, Inc., near the “pre-development” Echo Lake measured section of Inners (1975, p. 522-524). The pit at STOP 8B is on the property of Yu Tian Chou, who resides in the house just to the east. Permission to visit these sites should be obtained from the property owners.

Strata exposed on both sides of the roadway at STOP 8A include cherty, fossiliferous limestones of the Moorehouse and Seneca members of the Onondaga Limestone (Figure 179). The strata were formerly assigned to the Stroudsburg member of the buttermilk Falls and Echo Lake Members of the Onondaga Limestone (designations abandoned; see Ver Straeten, this guidebook, p. 35). The Tioga B K-bentonite bed of Smith and Way (1983), which marks the base of the Seneca Member, is exposed partway up through the succession (Unit 7 below).

Among the key features of STOP 8A are two deformed zones in upper Moorehouse and lower Seneca strata. Sedimentary structures inside the zones (e.g., ball and pillow structures, convolute bedding) and a lack of brittle deformation features indicate early soft sediment deformation of the strata.
Nodular chert bands within the deformed layers are folded with the strata, and indicate genesis prior to soft sediment deformation.

Chertification remains one of the long-unresolved questions in sedimentary geology. The Onondaga Limestone has been and continues to be an excellent laboratory in which to explore problems of the origin, timing, and distribution of chert-rich strata in the rock record. Biogenic opal from siliceous sponges, radiolarians, and/or diatoms are often suggested as the source of excess silica for Onondaga chert. However, the increasing number of recognized K-bentonites (altered volcanic beds) in Onondaga strata basinwide draw attention to another potential source of silica.

Faunas through the upper

Moorehouse and Seneca members are, for the most part, dominated by medium- to low-diversity assemblages. The small chonetid brachiopod Hallinetes is the most characteristic form of the faunas. Within the upper Moorehouse, however (close to the base of the overlying Seneca Member), a very different fauna appears, characterized by a more diverse assemblage with larger forms, including Paraspirifer brachiopods and solitary and colonial rugose corals. The change is associated with a shift in paleoecologic conditions, driven by a regressive lowstand of relative sea level (= base of Depositional Sequence 4 of Ver Straeten, this guidebook p. 35).

The small pit on the northwest side of US 209 (STOP 8B) exposes black shales and buff-weathering calcareous shales of the Bakoven and Stony Hollow Members of the Union Springs Formation (lower part of the "Marcellus shale"). Fully laminated to slightly burrow-mottled textures (best seen in Stony Hollow strata) and a general lack of macrofauna indicate anaerobic to dysaerobic conditions at the time of deposition. A deformed zone at the top of the black shales, found widely throughout eastern Pennsylvania and eastern to east-central New York, represents a regional decollement surface.

MOOREHOUSE AND SENeca MEMBERS
OF THE ONONDAGA LIMESTONE AT FAIRWAY (A)

Biostratigraphy and Age

As reported for STOP 7, the Moorehouse and Seneca Members of the Onondaga Limestone occur within the lower Eifelian (Middle Devonian) Polygnathus costatus costatus conodont zone (Klapper, 1981). The Tioga B bed at the base of the Seneca Member has been dated at 390 ± 0.5 Ma (Roden et al., 1990).

Stratigraphy

Onondaga Formation, 34.1+ ft (10.4 m+)
Moorehouse Member, 14.6+ ft/4.5+ m (= lower part of Stroudsburg Member of Epstein, 1984).
Unit 1 (Base of exposure to 6 ft/1.8 m)

The lowest beds exposed north of the road consist of massive, medium-dark-gray (N4), medium-gray (N5) weathering, fine- to coarse grained, partly bioclastic limestone containing profuse (nearly 50 percent by volume) of grayish black (N2) in irregular nodules (or concretions) up to about 1 ft (0.3 m) in diameter and discontinuous beds up to 2 ft (0.6 m) long. Medium-sand to granule-size fossil fragments include brachiopod, rugose-coral, and crinoid debris. At the very top of the unit is a thin drape of bioclastic calcarenite. As in all higher units, abundant hairline fractures in the chert are filled with white quartz. From exposures of this same interval in the parking lot on the opposite side of the road, it is clearly evident that this interval has undergone some sort of deformation (Figure 180). For further discussion of this point, see Deformed Zones below.

The cherts in this interval (and elsewhere in the section) display “circular” structures that may represent some sort of chemical differentiation of layers within the nodules, probably during diagenesis. The circular rinds may be dolomitic horizons where full chertification has not taken place. They would then represent a kind chemical rim formed at the time of chert formation, associated with the expulsion of Mg (Selleck, 1985).

Unit 2 (6 to 6.5 ft/1.8 to 2.0 m)

In sharp contact with the unit below is an irregular bed of medium-gray, mostly leached, tan weathering, mostly fine-grained, non-cherty limestone contained abundant comminuted brachiopod fragments and crinoid ossicles.

Unit 3 (6.5 to 10.1 ft/2.0 to 3.1 m)

Unit 2 grades up into medium-bedded (in beds 2 to 6 in. (5 to 15 cm thick), medium gray, medium-light-gray (N6) weathering, mostly fine-grained, partly bioclastic limestone containing abundant (30 to 40 percent), grayish-black chert in irregular nodular “courses” 1 to 7 in. (2.5 to 17.5 cm) thick, the largest nodules occurring toward the top. Fragmental fossil debris occurs in numerous thin stringers and wisps.

Unit 4 (10.1 to 13.7 ft/3.1 to 4.2 m)

The next unit up is in sharp contact with unit 3, its base consisting of a thin band of leached, slightly cherty, fine-grained to bioclastic limestone similar to unit 2. The main part of the unit is very much like unit 3, though not quite as cherty—the nodular chert bands being 1 to 3 in. (2.5 to 7.5 cm) thick. Relatively complete and fragmental fossils are profuse in the lower 6 in. (15 cm), but fossil debris is scattered and patchy in the upper 3 ft (0.9 m): Paraspirifer acuminatus and Pseudatrypa sp. occur near the base of the unit.
Unit 5 (13.7 to 14.2 ft/4.2 to 4.3 m)
Above unit 4 is a 0.3- to 0.8-ft- (9 to 24 cm) thick bed of medium-dark-gray, medium gray weathering, mostly fine-grained, slightly cherty limestone that is intensely burrowed (Figure 181) and coarsens upward into a 2- to 3-in.- (5 to 7.5 cm) thick coquinitic layer composed of a profusion of *Hallinetes* valves. Grayish-black chert occurs as a few ragged nodules 1 in. (2.5 cm) or less in diameter.

Unit 6 (14.2 to 14.6 ft/4.3 to 4.5 m)
Capping the Moorehouse Member is a single, continuous, locally folded bed—0.3 to 0.6 ft (9 to 18 cm) thick—of medium-dark-gray, cherty, fine-grained to coarsely bioclastic limestone containing abundant fragments of crinoids, small rugose corals, and chonetid (*Hallinetes*) brachiopods. Grayish-black chert occurs as nodules 1 to 3 in. (2.5 to 7.5 cm) thick that coalesce into a single irregular bed at the top of the unit.

**Seneca Member, 19.5+ ft/5.9= m (= upper part of Stroudsburg Member of Epstein, 1984).**
(The next two units are part of a 6.5-ft- [2.0 m] thick, structurally disturbed zone—sharply defined at the top and bottom—composed of rather chaotically mixed limestone, chert, and volcanic ash.)

**Unit 7 (0 to 3 ft/0 to 0.9 m)**
At the base of the contorted interval is 2.5 ft (0.75 m) of massive, light-gray (N6), micaceous and finely siliceous, noncalcareaous, tuffaceous siltstone (Tioga B of Smith and Way, 1983). Wedged in above and below the tuff bed are layers of contorted, light-gray, fine-grained to bioclastic, cherty limestone locally containing abundant *Hallinetes* valves and other fossil debris. Soft sediment deformation structures are visible at various positions along the outcrop (Figures 182 and 183) (see further discussion below).
Unit 8 (3.0 to 6.5 ft/0.9 to 2.0 m)

The upper half of this structurally disturbed interval consists mostly of tightly folded beds of medium-bedded (in beds 3 in. to 1.5 ft [7.5 cm to 0.45 m] thick), light-gray, cherty, fine-grained to coarsely bioclastic limestone (Figure 184). Grayish-black chert occurs in pods to 6 in. (15 cm) in diameter and discontinuous beds (folded) to 2 in. (5 cm) thick. At one spot at the top of the unit is a pocket of contorted, dark-gray (N3), very fine-grained, calcareous, probably tuffaceous, siltstone.

Unit 9 (6.5 to 13.5 ft/2.0 to 4.1 m) (w/ Unit 10 = Echo Lake Member of Inners, 1975)

Sharply overlying the contorted beds is a gently northwest-dipping interval of medium-bedded (mostly in beds 4 to 8 in. [8 to 20 cm] thick), medium-gray, fine-grained to coarsely bioclastic, cherty limestone containing numerous 2- to 4-in.- (5 to 10 cm) thick, wavy, coarsening-upward, coquitic bands composed of profuse Hallinetes valves and abundant small rugose corals (probably “tempestites”) (see Figure 179). Chert, grayish-black and occurring in irregular nodules 1 to 3 in. (2.5 to 7.5 cm) in diameter, is more abundant in the lower half of the unit (where it makes up about 30 percent of the rock).

Unit 10 (13.5 to 19.5 ft/4.1 to 5.9 m)

Exposed farther up the road (not in direct contact with the beds below, but directly succeeding them) is a 6-ft- (1.8 m) thick interval of cherty, mostly highly fossiliferous limestone identical to unit 8. About 20 percent of the unit is chert. The contact of the Onondaga Limestone and Marcellus Formation probably occurs not far above the top of this ledge.

Discussion

Deformed zones

A thick zone of deformed cherty limestones in the lower interval, exposed in the road cut and in the parking lot south of the road, feature cherts nodule layers that shift from horizontal- to vertically-bedded along the outcrop (see Figure 180). Characteristic features associated with brittle deformation appear to be absent. The lower deformed zone is approximately 15 ft (4.5 m) thick in the parking lot exposure; it also crops out in the lower two meters of the cut on the north side of the road (Unit 1, above).

The second deformed zone, approximately two meters in thickness, incorporates an upward succession of a thin cherty limestone (ca. 0.6 ft/0.2 m), green to light-gray, tuffaceous siltstones of the Tioga B K-bentonite (ca. 2.5 ft/0.75 m), an additional interval of cherty limestones (ca. 3.5'/1.0 m), and a thinner bed of tuffaceous siltstones (ca. 0.3 ft/0.1 m) at the top. The folded strata again lack brittle structures; features diagnostic of soft-sediment deformation are clearly visible along the outcrop, including small to large-scale ball and pillow/load structures and contorted bedding (see Figures 182, 183, and 184). The thin K-bentonite at the top of the deformed interval may represent a widely correlatable bed that occurs basinwide closely above the Tioga B bed, which was not previously reported by Smith and Way (1983) and Way et al. (1986).

The prevalence of soft sediment deformation features and the absence of brittle structures indicate early deformation, while the sediments were not fully lithified. No plastic deformation was
noted in the cherts; this appears to indicate very early diagenesis of the chert, prior to lithification of the limestones and also prior to folding of the strata. Ball and pillow structures visible in the upper deformed zone may imply in situ deformation. However, the larger-scale rollover of the sediments in the thicker, lower zone could be indicative of slump processes.

The deformation of pre-lithified sediments can be the result of many different factors, such as: 1) simple foundering of denser, coarser sediment layers into underlying less dense and more fluid layers; 2) seismic shocks and resulting deformation (producing a "seismite" layer); and 3) slumping and removal of strata during a relative sea level lowstand. Upper Moorehouse strata were deposited close to the onset of orogenesis Acadian Tectophase II, and during a lowstand of relative sea level sea level (beginning of Sequence 5 of Ver Straeten, this guidebook, p. 35). Therefore, any of the three factors could potentially have contributed to the deformation of upper Onondaga strata here. At present, the geographic distribution of the disturbed zones is unknown, and no detailed, bed-by-bed correlation of upper Moorehouse strata has as yet been attempted (e.g., STOPS 7 and 8), to determine if any upper Moorehouse strata are missing locally at STOP 8 due to slumping. More work is needed to determine the cause or causes of the early, soft-sediment deformation of the Onondaga Limestone seen at this locality.

**Chert in the Onondaga Limestone**

The process and timing of chert formation remains one of the great problems in sedimentary geology. Some cherts are recognized as undergoing early lithification; others are interpreted to reflect secondary replacement of limestone during later stages of diagenesis. The source of silica for chert formation is also a significant question; is it of a biogenic or inorganic origin? Biogenic sources include siliceous oozes of opal formed from the skeletal material of radiolarians, diatoms, or siliceous sponges. Dissolved silica in natural waters, including ground waters, may originate from different processes (e.g., chemical weathering, dissolution of skeletal opal, devitrification of volcanic ash).

The distribution of cherts in Devonian strata across the Appalachian Basin indicate that its genesis is controlled by depth-related processes, as yet not well understood. The chert variously occurs in carbonates (nodular and occasionally bedded types), fine-grained clastics, or as massive cherts, most notably in the Esopus- and Schoharie-equivalent Huntersville Chert (Ver Straeten, 2001) in the southern part of the Appalachian Basin. Field observation and numerous studies show a predominance of chert in facies deposited below normal wave base, and its near total absence in shallow (littoral to shoal) facies (e.g., Wilson, 1975; Maliva and Siever, 1989; Brett and Ver Straeten, 1994; Ver Straeten and Brett, 2000).

In the Onondaga Limestone, chert is predominantly nodular, sometimes appearing to take the form of *Ophiomorpha*-like trace fossils. The concentration of chert varies both vertically through the formation and regionally. In eastern New York and eastern Pennsylvania (Catskill to Stroudsburg area) chert is abundant in the Edgecliff and Moorehouse and locally the Seneca Members. In western New York, chert again becomes a significant part of the Edgecliff and Moorehouse Members, in some intervals comprising over 80% of the rock. The cherty Edgecliff facies in western New York were formerly assigned to the "Clarence Member," now recognized as chert-rich "Clarence facies" of the Onondaga (Brett and Ver Straeten, 1994).

Selleck (1985) suggested a biogenic origin of cherts in the Moorehouse Member due to solution of skeletal biogenic opal deposited with finer-grained carbonate, followed by transport and precipitation in local microchemical environments conducive to silica precipitation. In his model, distribution of silica was controlled by hydrodynamics, where fine biogenic silica material settled out in finer-grained carbonate facies. The occasional observation of siliceous sponge spicules and radiolaria in Onondaga and Esopus cherts (Selleck, 1985; Rehmer, 1976) and the general lack of sponge fossils does not
necessarily support a biogenic origin for the silica in the Devonian cherts; their presence could also be a function of preferential preservation within the cherts.

April et al. (1984) projected that the small amount of clay minerals in the Onondaga Limestone would provide only a small amount of silica for chertification. However, since that time, workers have documented an increasing number of K-bentonite beds in the Onondaga Limestone and equivalents. Smith and Way (1983) reported up to 8 beds in the upper Moorehouse to Seneca and lower "Marcellus" equivalents in central Pennsylvania. Conkin and Conkin (1984) reported additional beds through the formation, as has Ver Straeten (1996a; see Figure 26, page 42). At least 15 separate beds are documented from the Onondaga and basal Union Springs Formations at present, with an additional concentration in the middle part of the Onondaga in the southern part of the basin (Ver Straeten, 2001). Additional ash material may have been deposited and mixed into background carbonate sediments, leaving no easily recognizable traces. The process of altering volcanic ash to clays (devitrification), especially ash of high silica rhyolitic composition as documented for the Devonian K-bentonites (Waechter, 1993; Hanson, 1995), releases dissolved silica. This could provide a significant source for chertification.

The distribution of both chert and K-bentonites through Lower and Middle Devonian strata throughout the Appalachian Basin supports this correlation. Three clusters of K-bentonites occur in lower to middle Devonian strata across the basin—in the Kalkberg-New Scotland (Bald Hill K-bentonites), lower Esopus (Sprout Brook K-bentonites) and Onondaga (Tioga A-G K-bentonites, the Tioga Middle Coarse Zone and additional beds) Formations and their equivalents across the basin. Each of these formations and immediately underlying rocks (Glenerie Limestone, Huntersville Chert) feature widespread nodular to massive chert facies. In contrast, analogous offshore facies in Silurian rocks in New York, where only rare K-bentonites occur, show surprising little chert development. Much further work is needed to resolve these issues.

BAKOVEN AND STONY HOLLOW MEMBERS
OF THE UNION SPRINGS FORMATION (B)

The small pit at STOP 8B on the northwest side of US 209 exposes the contact of the Bakoven (3.3 ft/1.3 m) and the overlying lower part of the Stony Hollow (9.5 ft/2.9 m) Members of the Union Springs Formation, in the lower part of the "Marcellus Shale" (Marcellus Subgroup of Ver Straeten et al., 1994; see Ver Straeten, this guidebook, p. 35). This exposure was formerly more continuous (Figure 185) but blocks of Stony Hollow recently separated along cleavage planes, slumped, and covered part of the Bakoven Member low in the outcrop. The base of the exposure is approximately 18 ft (5.5 m) above the Onondaga Limestone.

Recent studies of Lower to Middle Devonian (Pragian-Eifelian) strata better outline the relationships of Middle Devonian strata of the "Marcellus Shale" all along the Appalachian Basin outcrop belt. Two main successions occur within the former Marcellus Formation of New York, which overlies the Onondaga Limestone and is overlain by the Skaneateles Formation. Ver Straeten et al. (1994) defined two formation-level units within a redefined "Marcellus Subgroup"—a lower Union Springs Formation and upper Oatka Creek Formation. A pair of thin limestone units (Hurley and Cherry Valley Members) lie at the base of the Oatka Creek Formation. The two formations, various subunits and several distinctive, time-significant marker beds are widely correlatable across the Appalachian Basin, between New York, southwest Virginia, and central Ohio.
The term "Marcellus Shale" is used differently in Pennsylvania from the New York succession, where it is bounded by time-significant marker horizons. In Pennsylvania it is used as a lithologic term for all lower Hamilton Group black shale facies, and variously represents strata of only the Bakoven Member (Union Springs Formation) to what likely represents strata of the Skaneateles Formation in the middle part of the Hamilton Group. Widespread recognition of the various member and marker beds permits basinwide correlation of Union Springs and lower Oatka Creek strata. Ver Straeten (this guidebook, p 35) briefly outlines this new stratigraphic revision and applies it to strata in eastern Pennsylvania. The Union Springs Formation in eastern Pennsylvania is comprised of the Bakoven (lower black shales; formerly Union Springs) and Stony Hollow (upper calcareous shales and siltstones) Members. The Oatka Creek Formation in the area is comprised of the Hurley and Cherry Valley Members at its base (previously unrecognized at the top of the Stony Hollow Member) and overlying strata of the Brodhead Creek Member. The top of the formation at present remains poorly constrained, pending further work.

**Union Springs Outcrop Description and Regional Overview**

Organic-rich, dark-gray to black shales of the Bakoven Member generally appear laminated to locally burrow-mottled in the region. The laminated textures of the organic-rich Bakoven point to deposition under widespread anoxic conditions across much of the Appalachian Basin during the onset of the second of four major Tectophases (Ettensohn, 1985) of the Acadian orogeny.

Weathered surfaces of the more buff to medium-gray Stony Hollow exhibit the laminated to burrow-mottled textures characteristic of the lower to middle parts of this member throughout the region. When present, its fauna is characterized by small, elongate, conical dacryonarids and styliolinids, with rare small brachiopods and bivalves. Textures and macrofauna present indicate anoxic to dysoxic conditions succeeded by a trend of increasing oxygen availability and shallowing upward to the overlying Hurley Member (not seen in this outcrop). Another pit on the north side of US 209 a short distance to the south exposes middle strata of the member, characterized by calcareous silty mudstones and siltstones and a macrofauna including proetid trilobites (*Dechenella*) and the small rugose coral *Guerichiphyllum*.

In distal areas of the basin in central New York and south-central Pennsylvania, the Bakoven black shale facies extends higher in the section completely displacing the Stony Hollow and reaching up to the Hurley Member and equivalents. However, in more proximal outcrops across eastern New York to central Pennsylvania, the contact between the Bakoven black shale below and the overlying Stony Hollow, Purcell, or Turkey Ridge members is synchronous, and closely overlies a thin K-bentonite bed. The K-bentonite represents a significant marker bed within the Union Springs Formation and equivalent strata, and is correlatable from eastern New York through Pennsylvania to southwest Virginia and into
the middle of the Delaware Limestone in central Ohio. It has not as yet been located at this outcrop or in the Delaware Water Gap area.

Projecting across US 209 from the top of the exposed Onondaga Limestone, the Bakoven Member here is approximately 24 ft- (7.3 m-) thick. The total thickness of the Union Springs Formation is not known for this locality. Alvord and Drake (1971) estimate 200 ft (61 m) for the combined Bakoven (formerly Union Springs) and Stony Hollow Members in the Bushkill quadrangle. In the Stroudsburg quadrangle immediately south, Epstein (1973) projected 50 ft (15 m) of Bakoven at the base of the formation. Alvord and Drake's (1971) 200-ft estimate for the Union Springs Formation as a whole is probably high; they include the Hurley and Cherry Valley Members at the top of the Stony Hollow, which are now placed within the overlying Oatka Creek Formation. Alvord and Drake (1971) also estimate 750 ft (230 m) of post-Stony Hollow Brodhead Creek Member (Willard, 1938) in the Bushkill quadrangle. It is not presently known how much of this represents Oatka Creek-equivalent strata.

The relatively thin nature of the Union Springs and associated strata in Monroe County is in sharp contrast to subsurface data from Pike County to the north and the Hudson Valley in eastern New York. Well log correlations by Rickard (1989) indicates 300 ft (91 m) of Union Springs strata and a total of 1238 ft (377 m) for the Marcellus Subgroup from a well in Shohola Township, Pike County, PA. In western Pike County (Blooming Grove township), Rickard (1989) reports a thickness of 260 ft (79 m) for the Union Springs Formation. Near Kingston, New York, the Bakoven and Stony Hollow Members total approximately 310 ft (100 m) and 223 ft (68 m), respectively (Rickard, 1989; Ver Straeten, 1996a). The overlying Hurley and Cherry Valley members of the Oatka Creek Formation total 7 and 10 m, respectively (Ver Straeten, 1996a). The post-Cherry Valley Oatka Creek Formation strata near Kingston totals 1293 ft (394 m), which yields a combined Union Springs and Oatka Creek (Marcellus Subgroup) thickness of 1900 ft (579 m) (Rickard, 1989).

**Union Springs Structural Interpretation**

The Stony Hollow Member has slumped over the faulted Bakoven Member. Most of the deformed rock is now covered by slumped Stony Hollow debris. The Union Springs consists of at least 6 ft (1.8 m) of deformed medium-gray to dark gray, non-calcareous, poorly to thinly bedded shale overlain by at least 10 ft (2.9 m) of laminated to thin-bedded, medium-gray, calcareous siltstone (Figure 185). The bedding trends N60°E/12°NW. A cleavage (trending N56°E/60°SE) cuts through the entire section, as does a scattered second-generation crenulation cleavage (trending N38°/61°SE). The Bakoven also contains a phacoidal-like cleavage that contains sigmoidal drag at its extremities and down-dip slickenlines. These features portray a tops-to-the-northwest sense of rotation. This can be seen both within individual beds but also over the entire zone (Figure 186). Of particular note here is the near complete separation of the deformation from the overlying Stony Hollow. The uppermost Onondaga that is

![Figure 186. Close up of decollement in Bakoven shale at STOP 8B. Note the sigmoidal drag in the cleavage, indicating a tops-to-the-northwest transport direction. Hammer is approximately 1 ft (15 cm) long.](image-url)
observed at STOP 8A also appears to be relatively free of tectonic deformation. This suggests a thin layer-parallel deformation zone (“decollement”) totally within the Bakoven.

This exposure is important because outcrops illustrating the deformation between Pocono/Catskill Plateau rocks above and the more complexly folded Middle Devonian-Silurian rocks below are rare. Several other examples of Marcellus deformation exist in the region. Fletcher and Woodrow (1970) described two separate slip zones at 168 and 468 ft (51.2 and 142.6 m) above the Onondaga contact in the Hess #1 well in Shohola Township, Pike County. Across the Delaware River an isolated Bakoven exposure exhibits similar structures (Herpers, 1951). There, the deformed shale was shiny enough that it was mistaken for coal and an adit was constructed. The adit in northwestern New Jersey was excavated in 1818 (Figure 187). Herpers (1951) rediscovered the “mine” and described a 27-ft long, 3-ft wide and 5-ft high (8.2 x 0.9 x 1.5 m) adit. Strained, slickensided Bakoven shale was also mistaken for coal at different sites in eastern New York, and similar horizontal shafts driven into the rocks (Chadwick, 1944; Don Van der Zee, personal communication to CVS, 1990). The deformed Marcellus exposure in New Jersey is approximately 180 ft (54.9 m) above the Onondaga contact (Herpers, 1951). However, all these regional Marcellus slip zones appear to be stratigraphically higher than exposed at STOP 8B. Detailed local mapping of the Marcellus is lacking so other, higher slip horizons can not be verified.

The Marcellus Shale is a major regional slip/decollement horizon. Numerous examples of a deformed zone just below the contact of the Bakoven and Stony Hollow exist along the 120 mi (200 km) separating Kingston and Oriskany Falls, New York (Chadwick, 1944; Rickard, 1952; Pedersen et al., 1976; Bosworth 1984a, b; unpublished data of Ver Straeten). Bosworth (1984a,b) suggested this represented a regional decollement zone. Characteristics of this zone include duplex structures, discrete faults and associated slickensided and striate/polished shale fragments (Bosworth 1984a, b). Similar Marcellus examples exist along the Pennsylvania outcrop belt to the west (Wood and Bergin, 1970; Nickelsen, 1986; Zhou et al., 1996; Shumaker, R.C., 1997; Faill, 1998; Faill and Nickelsen, 1999). Epstein and Epstein (1967) and Epstein et al. (1974) also suggested that the Marcellus is a regional decollement zone in the Delaware Water Gap region. They proposed that due to difference in formation thickness and rheology, select sedimentary packages responded differently to the regional stress than other sediments. In his model the Poxono Island through Buttermilk Falls/Onondaga consists of thinner bedded units that vary considerably lithologically over short thickness. This allowed a stronger degree of folding than the more consistently bedded and lithology of the underlying Shawangunk and Bloomsburg package and overlying Mahantango and above formations.

Decollements in the Marcellus as exposed here relate to accommodation of strain translation originating lower in the crust. Herman et al. (1997) modeled approximately 1 kilometer translation strain on the leading-edge thrust mapped in the Paulins Kill window, just east of Kittitanny Mountain. This thrust could have ramped upsection through the Shawangunk and Bloomsburg and into bed-parallel
decollement horizons in horizontal lying, stratigraphically higher units. Unfortunately, the Shawangunk-Bloomsburg section where this might have occurred has all been removed by erosion; therefore, this idea is pure supposition. Shear strain seen at this location is consistent with northwest-directed motion possibly related to Paulins Kill structures. In this region the Poxono Island and Marcellus Formations, acting as weak layers separating stiffer Shawangunk-Bloomsburg and Helderberg-Onondaga rock packages, are susceptible horizons to accommodate this northwest-directed strain (Epstein and Epstein, 1967). If deformation at this location relates back to the Paulins Kill structures and not some unknown overlying decollements, only 1 km of translational strain would be partitioned between the Marcellus and Poxono Island detachment zones. Therefore total offset at this outcrop is relatively minor, less than 1 kilometer.

Several questions do remain, though. Are the Marcellus decollement zones the same throughout the region? In other words, is there one horizon that bifurcates into several zones along strike or several discrete horizons that can be traced along the outcrop belt? Drilling may be needed to answer this question as the Marcellus resides under the Delaware River from Bushkill to and past Port Jervis, New York. Is the northwest-directed thrust sense of motion consistent throughout the region? If this zone represents a detachment/decollement zone caused by large block differences in rheology and associated fold formation, would a consistent sense of motion be indicated?

DEPOSITIONAL HISTORY OF THE ROCKS AT STOP 8 (AND ASSOCIATED STRATA)

The limestones of the Onondaga Formation and overlying clastics of the Hamilton Group record a second major shift in deposition associated with alternating tectonism and quiescence in the Acadian orogeny (Ettensohn, 1985; Ver Straeten, 1996a). Onondaga and equivalent carbonates, which occur widely across eastern North America from James Bay, Ontario to Maine, southwest Virginia, and Illinois (Koch, 1981), were deposited during a late, relatively quiescent stage of Acadian Tectophase I. The sharp shift to fine-grained, organic-rich black shales in the overlying lower part of the Hamilton Group, especially in the Union Springs and Oatka Creek Formations, mark a period of subsidence of the Appalachian foreland basin system during the early stage of Acadian Tectophase II (Ettensohn, 1985). This indicates, in part, the foundering of the widespread Onondaga platform to ramp geometry of the basin and load-induced overdeepening of the foredeep. With decreasing subsidence during middle Hamilton time, the initial black shales are succeeded by coarser-grained progradational sandstone-dominated strata, which progressively young toward the basin center. On the proximal margins of the basin, the sand-rich facies grade through shelf and nearshore facies to at least temporary continental facies; this is apparently seen as low in the section as in Stony Hollow-equivalent strata of the Turkey Ridge Sandstone at Swatara Gap, PA, where plant root traces are found within the unit. In eastern New York, shallow marine to shoreface sandstones strata give way to continental deposits near the top of the Oatka Creek Formation.

Increased sandstone deposition characterizes middle Hamilton Group strata all across the central and northern Appalachian basin (e.g., middle member of the Mahantango Formation, PA; Ludlowville Formation, NY). However, upper Hamilton rocks (upper member, Mahantango Formation, PA; Moscow Formation, NY) are characterized by a decrease in sandstone and increased mudrock content, followed by a return to widespread limestone deposition or equivalent proximal sandstones (Tully-Gilboa-Sparrow Bush Formations, PA and NY). The succession of basinal black shales to coarser sandstones and the following return to carbonate-rich facies reflects an overall decrease in tectonism and its effects in the Appalachian foreland basin during Acadian Tectophase II.
A number of discrete, thin clay beds, representing altered volcanic ash falls, have been noted within the upper part of the Onondaga and lower part of the Union Springs Formations. These K-bentonite beds include the Tioga A-G zone of Smith and Way (1983) and additional beds that have been discovered by various workers since (e.g., Conkin and Conkin, 1984; Ver Straeten, 1996a, 2001). These appear to indicate an increase in magmatic activity in the adjacent Acadian orogen during the onset of Acadian Tectophase II.

Detailed study of litho- and biofacies trends across the basin reveal several scales of fining- to coarsening-up cycles, reflective of changes in relative sea level during the Middle Devonian. At STOP 8, a shallowing into the upper part of the Moorehouse Member of the Onondaga Limestone is reflected in a subtle coarsening-up and gradational shift toward shallower water carbonate lithologies. Also notable through the succession is a shift from fossil assemblages characterized by small- to medium-size brachiopods (especially *Hallinetes*) and other fossils characteristic of relatively deeper water "shelf" settings in the Onondaga to medium to large forms, including *Paraspirifer* and colonial rugose corals seen in intermediate depth Onondaga facies (seen in Units 3-5 here). Interpreted within the context of the sequence-stratigraphy paradigm, the transition to the shallower facies from the middle to upper Moorehouse represents late Highstand Systems Tract of a lower to middle Onondaga depositional sequence (Depositional Sequence 4 of Ver Straeten, this guidebook, p. 35). Relatively shallow upper Moorehouse facies are interpreted to represent the Lowstand Systems Tract of a fifth depositional sequence above the base of the Oriskany Sandstone.

Subsequent strata show an initial return to *Hallinetes*-dominated, relatively deeper shelf facies in the Seneca Member of the Onondaga Limestone, marking initial transgression (base Transgressive Systems Tract of Sequence 5). Covered at STOP 8A, somewhere very close to the top of the exposed Seneca, is a sharp transition into basinal black shale facies of the Bakoven Member of the Union Springs Formation. Within the overlying covered Bakoven lies a point of maximum flooding during Depositional Sequence 5; this marks the contact between the Transgressive and early Highstand Systems Tracts. Upper black Bakoven Shales and the lower part of the Stony Hollow Member, exposed in the pit at STOP 8B, mark aggradational to early progradational deposits in the upper part (late Highstand Systems Tract) of Sequence 5.

The sequence stratigraphy of the overlying remainder of the Hamilton Group and Tully are well defined in New York (Brett and Baird, 1996; Ver Straeten, 1996a; see Ver Straeten, this guidebook p. 35). Four additional "third order" sequences occur through the Hamilton Group; these comprise strata of the Oatka Creek, Skaneateles, Ludlowville, and Moscow Formations of New York (Sequences 6-9). A tenth sequence at the in uppermost Middle Devonian strata consist of the Tully Limestone and overlying Genesee Formation and their equivalents.

In Pennsylvania, post-Union Springs sequence stratigraphy is, in general, poorly constrained (however, see Prave and Duke, 1991, and Slattery, 1996, for a discussion of Mahantango Formation sequence stratigraphy in central Pennsylvania). These sequences can tentatively be identified through Middle Devonian strata in the southern part of the basin. As a guide to future work in Pennsylvania, a challenge will be to delineate four depositional sequences within the upper part of the "Marcellus Shale" and the Mahantango Formation throughout Pennsylvania, building on a simple framework laid out in Ver Straeten (this guidebook, p. 35).

Leave STOP 8, returning to US 209 North

0.6 17.7 Stop sign. Turn right onto US 209 North at Fairway entrance.

0.1 17.8 To right is an old abandoned sand pit in the Sand Hill delta, a late Wisconsinan ice-contact delta (see STOP 9).

0.2 18.0 Traffic light. Turn left onto SR 1016.

0.2 18.2 Turn left into gravel pit.
INTRODUCTION

This sand-and-gravel pit is one of several intermittently active operations in the Sand Hill delta, a prominent ice-contact feature developed in the Echo Lake lowland (Figure 188). This particular pit is on land owned by Resorts USA, Inc., and permission to enter it should be obtained at their nearby offices on US 209 at Shoemakers.

Geologic Setting

The Delaware River at Wallpack Bend (Figure 189) leaves a strike valley underlain by the Marcellus shale and cuts through Wallpack Ridge following the line of a large structurally controlled meander. From here, it continues to Delaware Water Gap following a narrow strike valley underlain by the Poxono Island Formation. The Echo Lake Lowland is the continuation of the “Marcellus” valley, forming the head of a strike valley that continues 12 miles southwest through Oak Grove to Brodhead Creek valley. White (1882) referred to this strike valley as the “Stroudsburg buried valley,” noting that it was filled with glacial drift that in places may have been as much as 100 feet thick. Topography around the Wallpack Bend area lends itself to the supposition that the Delaware River at one time flowed through the Echo Lake lowland on its way to the Delaware Water Gap. However, the narrowness of the lowland near Oak Grove does not support the location of a Delaware River paleovalley here during the Pleistocene. The Stroudsburg buried

Figure 188. Location of STOP 9, Sand Hill delta, and surficial geologic map near Shoemakers, Pennsylvania. Base map modified from Bushkill 7.5-minute quadrangle. Legend: af = artificial fill; Qal = alluvium; Qaf = alluvial fan; Qf = outwash fan; Qft = meltwater-terrace; Qd = Sand Hill delta; Qtr = thin till; and r = rock.

Figure 189. Physiography of the area around Wallpack Bend area and location of Sand Hill, Echo Lake Lowland, Stroudsburg buried valley, and features and places named in text.
valley is likely the product of glacial erosion, given the weak resistance of the Marcellus shale to corrasion and the fact that ice flow was directed down the valley.

Oak Grove also marks the former drainage divide of the buried valley. Southward, the buried valley floor slopes to the southwest toward Brodhead Creek. Northward, the valley’s rock floor slopes northeast, passing beneath Echo Lake and Sand Hill toward the Delaware Valley at Bushkill. The present drainage divide lies about 3 miles northeast of Oak Grove near Echo Lake.

The Echo Lake lowland is floored by thick deposits of stratified drift (Figure 190). Texture of the materials ranges from boulder-cobble gravel to silt and clay. Deposits are typically collapsed and kettles are common, showing that these materials were laid down against ice. A few level outwash plains exist around Echo Lake and Sand Hill. The Echo Lake outwash reaches an elevation of 535 feet, and the Sand Hill outwash reaches an elevation of 520 feet (elevation was determined from the 1998 revision of the Bushkill quadrangle). Northeast of Sand Hill, the elevation of the valley floor quickly drops off to less than 380 feet. Bushkill Creek enters the lowland just behind Sand Hill, flowing about 2 miles northeastward before emptying into the Delaware River near Bushkill, PA. White (1882) estimated that the bedrock floor of the valley near Echo lake was below 400 feet in elevation and may be much lower, given that the elevation of Bushkill Creek is about 340 feet before it enters the Delaware.

Discussion

Several investigations on meltwater deposits in Minisink Valley have argued that the late Wisconsinan ice sheet disappeared from this area either by regional stagnation or marginal retreat. This controversy is a recurring argument, and it is certainly not unique to the Minisink Valley area. Earlier research by 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), Depman and Parillo (1969) and Crowl (1971) favored a stagnation model where the uplands were deglaciated first with residual masses of ice left in the valleys. Large areas of collapsed topography in kames and kame terraces, many kettles, ice-contact scarps, and unpaired terraces were cited as evidence for stagnation. Epstein (1969) and Epstein and Koteff (in press) near Stroudsburg, PA, and Ridge (1983), Witte, (1997a; this guidebook, p. 81), 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 shown that deglaciation took place largely by the systematic melting back of the margins of the Kittatinny and Minisink Valley ice lobes. A narrow zone of stagnant ice bordered on the glacier’s margin, but it wasted back synchronously with the retreating active glacier margin.

Crowl (1971) surmised that glacial materials covering the Echo Lake lowland were chiefly the products of glacial stagnation. He based this on the many kames and kame terraces he identified, the existence of kame (collapsed) structures, kettles, and highly variable surface texture. Ice-retreat positions were not mapped. However, the end moraine near Dingmans Ferry and Montague in Minisink Valley did show that ice in Minisink Valley was active at times and possibly readvanced for several miles before depositing the moraines. The overall context of deglaciation, however, was one of valley-wide stagnation and not sequential ice margin or stagnation-zone retreat. Most ideas about deglaciation during the middle part of the 20th century involved models of stagnation. A common view was that uplands became deglaciated first leaving large residual masses of ice on the valley floors. Meltwater, derived from melting stagnant ice, deposited sediment over and around residual ice blocks left in the valleys. Upon their melting, kames, kame terraces, and kettles were formed. In addition, after residual ice masses melted, outwash from distant sources upvalley formed an extensive outwash plain, covering

Figure 191. Top-down model of glacial stagnation that was used to illustrate deglaciation and meltwater deposition in Minisink Valley. Modified from Depman and Parillo (1969, Figure 4).
the lower parts of the valley floors. This model of deglaciation is summarized in Figure 191. Locally it was used to explain deglaciation of the Delaware Valley around Tocks Island (Figure 192). The complex geometry of glaciolacustrine and glaciofluvial deposits illustrated in Figure 192 was a product of the chaotic environment in which they were deposited.

**SAND HILL DELTA**

**Local Geology**

Crowl (1971) suspected that the flat-topped deposit that made up the Sand Hill area was deltaic. However, because good exposures of foreset bedding were not available and the margins of the deposit exhibited ice-contact topography, he mapped the deposit as a kame that was laid down between ice blocks that lay in the Werry Lake depression and Bush Kill valley. The pit at STOP 9 shows that Crowl’s original suspicions were correct. It also shows why this area is so appropriately named Sand Hill. Materials exposed along the pit’s high, steep walls show that the Sand Hill delta consists of about 10 feet of gravelly topset beds overlying about 55 feet of sandy foreset beds (Figure 193). Topset beds, which in many places have been
partly stripped, consist of planar-bedded, framework-supported, cobble-pebble gravel (Figure 194). Elongated clasts exhibit imbrication that show a paleo-stream flow of south to southeast. Larger cobbles are as much as 8 inches in diameter, and many of the gravelly beds contain larger clasts at their base. The gravel is largely derived from local sources, and consists of gray shale, siltstone, and sandstone with secondary amounts of white quartz-pebble conglomerate, red sandstone, and limestone.

Foreset beds (Figure 195) consist chiefly of fine to medium sand with a few thin beds of pebbly sand. Dropstones are common. Individual beds are typically less than 2 inches thick and dip less than 11° in a northeast to southeast direction. Sedimentary structures consist of stacked sets of climbing ripple-drift sequences of Type A and B ripples capped by silt drapes. These rhythmic sets of sandy foresets are typically deposited on the middle part of the delta slope by underflow currents that flow down the delta front along distributary channels. The Sand Hill section represents a subaqueous environment where sedimentation was rapid (possibly on a daily or hourly scale). These structures also reflect rapid changes in current velocity along the prograding delta front.
The clay drapes are probably seasonally deposited, laid down during the winter months when meltwater production was minimal.

The thick mid-delta component represented here is probably a function of the small size of the lake basin and that the delta was chiefly built by meltwater from a distant source up Bush Kill valley. Pebble provenance, and direction of paleo-flow determined by the dip of foreset bedding, and gravel imbrication support a westerly meltwater source. A slight rise in elevation of the delta plain in a northwest direction (510 to 520 feet) also supports a Bush Kill source.

A second pit that lies about 1000 feet northwest of STOP 9 (see Figure 188) shows similar materials with the exception that the lower part of the pit contains a core of very coarse gravel. Bedding here dips about 12° to the northwest. This material does not look collapsed. Perhaps these coarse gravel beds were deposited at the mouth of a subglacial tunnel prior to their burial by the sandy foresets.

**ECHO LAKE LOWLAND**

**Summary of Deglaciation**

Meltwater deposits in the Echo Lake lowland delineate two ice-retreat positions, one near Echo Lake, the other near Sand Hill (Figure 190). The Echo Lake morphosequence consists of an extensive valley train that has its head located just northeast of Echo Lake. Reaching an elevation of about 535 feet, the Echo Lake sequence continues downvalley about four miles to Marshalls Creek, where it lies at an elevation of about 490 feet. Several lakes and many smaller kettles show that the Echo Lake outwash was deposited over and around small stagnant ice blocks. Because the floor of the Echo Lake lowland slopes northward, glaciolacustrine deposits probably lie beneath the valley train. Higher standing deposits downvalley from Echo Lake (Figure 190) are outwash fan deposits laid down by meltwater entering the lowland at the mouths of tributaries. There are a few kames between Meadow Lake and Oak Grove. These deposits are small mounds of sand and gravel that lie above the Echo Lake outwash. They predate the outwash, and represent places where depressions in the glacier’s stagnant margin were filled with sand and gravel. The Echo Lake margin has been tentatively correlated with the Zion Church margin in Minisink Valley.

Retreat from the Echo Lake margin to the Sand Hill margin resulted in the formation of a small proglacial lake between the Echo Lake outwash and the margin of the glacier. Collapsed topography on the northeast side of the Sand Hill delta places the Sand Hill margin near the modern location of Bushkill Creek. Later meltwater and postglacial drainage has eroded part of the Sand Hill delta, especially where Bushkill Creek enters the Echo Lake lowland. “Sand Hill Lake” drained over a rock-floored spillway (elevation 495 feet) near Middle Smithfield Church. The lake may have also drained out through a series of ice block depressions now marked by Echo and Coolbaugh Lakes. Since this spillway elevation is estimated at 505 feet, lake drainage along this path was probably short-lived. Downstream from Coolbaugh Lake, the lake’s outlet waters cut an erosional channel in the Echo Lake outwash.

Retreat from the Sand Hill margin resulted in meltwater draining the Bushkill Creek valley to flow northeastward, directly to the Minisink valley. Outwash fan deposits and meltwater-terrace deposits behind the Sand Hill delta mark the lowering of local-base level control in the lowland and eventual opening of meltwater through-drainage to the Minisink as ice retreated out of the Echo Lake Lowland.

The northeast, stepward style of ice retreat suggested for the Echo Lake lowland is similar to that suggested for Minisink Valley by Witte (this guidebook, p. 99) where the longitudinal profiles of valley-train terraces and their downstream continuation from their heads show that large masses of residual ice did not cover the valley floor. Collapsed topography and kettles do indicate deposition against and over
stagnant ice. However, these landforms are common components of stagnation-zone retreat, and there is no need to invoke regional or valley ice-tongue stagnation to explain deglaciation.

Leave STOP 3, turning right on SR 1016 to return to US 209 North.

0.2 18.4 Traffic light. Turn left on US 209 North.
0.1 18.5 Fernwood Resort Hotel to left. Resort sits on an outwash terrace about 50 feet below the Sand Hill delta. The terrace is a remnant of a large meltwater fan laid down at the mouth of Bushkill Creek valley after ice retreated from the Echo Lake lowland.

0.4 18.9 Enter Delaware Water Gap National Recreation Area.
0.3 19.2 Bushkill Creek to left, well below a steep fluvially cut scarp. US 209 runs along a remnant of the meltwater fan discussed at mile 18.5.

0.4 19.6 Pass Community Drive on right and descend onto flood plain of Bushkill Creek (elev. ~360 ft, about 120 feet below the Sand Hill Church delta plain).

0.3 19.9 Bushkill Visitors’ Center to right.
0.1 20.0 Cross Bushkill Creek into village of Bushkill, passing from Monroe County into Pike County. Bushkill was once a thriving community, boasting—with adjacent Maple Grove in Monroe County—two tanneries, two grist mills, a sawmill, at least two hotels, and passenger train service. On the creek bank to the left are the stone foundation ruins of the mill built by Henry Peters in early 19th century—and still standing in the 1970’s. A mill had existed on this spot since at least 1770 (Henn, 1975).

0.2 20.2 Traffic light (blinking). Road to Bushkill Falls (SR 2001) to left. The store on the corner (now “Deli Depot”) is one of the oldest enterprises in continuous service in Pike County. A store was first established here by Henry Peters (of the mill). After several changes of ownership, Ralph Turn and William Cook took over the operation—and the inscription “Turn and Cook” is still imbedded in the concrete step at the front entrance (Henn, 1975). Spared destruction by the Park Service in the 1970’s, the store now does a good business in food, sundries, antiques, and tourist items.

0.1 20.3 Reformed Church of Bushkill (1874) and old cemetery to right. The elevated tract north of the creek here has been mapped as a gravelly kame terrace, the top surface of which lies 70-80 feet above the Delaware River (Crowl, 1971). Alternatively, the terrace may be a remnant of much more extensive outwash laid down by meltwater in the Delaware River and Bushkill Creek valleys. As the glacier’s margin retreated up the Delaware Valley, the meltwater river adjusted to its longer course by eroding outwash downvalley.

1.1 21.4 Pass Brodhead Road on left. From time to time Brodhead Road has been closed after an oversteeped till slope (about 0.3 miles from Route 209) failed, sending an earth flow across the road. (Epstein, this guidebook, p.119)

0.1 21.5 Bushkill Access Area to right. Buses pull up along right side of road to discharge passengers, then drive ahead about 0.1 mile and park in large parking area to the left. Lunch will be served along the banks of the scenic Delaware River. Good restroom facilities are available.

LUNCH STOP. BUSHKILL ACCESS AREA: THE POWER OF WATER—A BRIEF HISTORY OF THE DELAWARE WATER GAP NATIONAL RECREATION AREA.

Host: Megan O’Malley.
Water—one of the most powerful and important forces on Earth: In the right amount, water is the giver of life, but too little or too much of it wrecks havoc. It is the multifaceted nature of water, its inherent life giving properties that exist alongside its ability to destroy that have motivated people for millennia to control its power. Battles have been fought and lives lost over control of water rights. Though not bloody battles, the story of the Tocks Island Dam Project, the beginning of Delaware Water Gap National Recreation Area, and the question of just who gets to control the water of the Delaware River are issues as dramatic and complicated as any military conflict (Albert, 1987).

With an average discharge of 11,750 cubic feet per second, the Delaware River is only the 33rd largest river in the United States. At 331 miles long and at places less than 150 yards wide the Delaware is certainly no Amazon or Nile. Why then the controversy? As they say in the world of retail—location, location, location. About 10 percent of the United States population lives within the Delaware River Basin. That's millions of people who could potentially benefit from the river as a source of drinking water and as a source of energy in the form of hydro-electric power. It is also thousands of people living along the river in towns and cities who could lose their homes and lives should the Delaware flood.

The issue of damming the Delaware River first appears in 1783 when the nation is still in its infancy and the states of New Jersey and Pennsylvania sign an anti-dam treaty. For the next 200 years or so, the issue would be revisited in the countless court cases and involve many millions of dollars. Pennsylvania, New Jersey, and New York battled well into the 20th century without conclusively resolving the issue of who got to control the water of the Delaware River.

Though the fight raged in the U.S. Supreme Court for years, it wasn't until 1955 that the damming of the Delaware River captured the attention of the general public in a very real and emotional way. The summer of 1955 was hot and dry. By August the region was in the middle of a drought that was quickly becoming severe. Water levels were low and the ground was bone dry. On August 12 the drought ended with what would become catastrophic results. Little can heave rain on an area with the intensity and violence of a hurricane. And that's exactly what hit the Delaware River Basin on the 12th. Hurricane Connie hit land in North Carolina and ended the drought that had plagued the East Coast. Because of the drought, water levels throughout the east were very low and as a result the water was absorbed by the soil with much less flooding than would otherwise have resulted. The drought ended, water levels were restored to normal and the parched soil was saturated. But the story does not end there.

Hurricane season is still peaking in August, and days later another large-scale storm, Hurricane Diane, pounded the East Coast. The path of Diane, which closely followed the Delaware River and its tributaries, caused a tremendous amount of severe flooding. Eleven inches of new rain fell on the Delaware River Basin on soil that was still saturated from the rains of Hurricane Connie. The flooding left nearly one hundred people dead. The issue that had been argued in the courts for years suddenly crystallized.

The issue of damming the Delaware River had for years been an issue debated among New York, Pennsylvania, and New Jersey. Building a dam to facilitate using the river for drinking water was an issue between these states, but flood control was the responsibility of the federal government. Thus, a new chapter in the story begins.

The U.S. Army Corps of Engineers, the agency responsible for building most major dams in the United States, was called in to survey the river and determine where to build the dam. With the involvement of the federal government, new elements were added to the project. From this point on the dam project would be designed to provide drinking water, hydro-electric power, flood control, and recreation. After several years of surveys, appropriations and debate Congress passed the Flood Control Act of 1962. Under the provisions of this act, the Army Corps would build a dam at Tocks Island,
located about 6 miles up river from the Delaware Water Gap. In 1965 a second act authorized the creation of Delaware Water Gap National Recreation Area which would be a unit of the National Park Service.

Though a definitive plan was now in place, the project was nowhere closer to being realized. New and old problems alike plagued the project. The escalation of the Vietnam War meant less money for Tocks Island Dam. Problems also arose with the dam itself. The unique geology of the area (Depman and Parillo, 1969) meant the plan for the actual structure of the dam would have to be revised. This revision would add millions of dollars to the project. New problems loomed for the Tocks Island dam.

The 1960's saw the beginning of the environmental movement—and soon the Tocks Island dam project was doomed. It began in 1965 with the sale of Sunfish Pond, one of the few glacial lakes in New Jersey. A pristine and unique environment, Sunfish Pond was sold to a consortium of power companies so that a storage reservoir affiliated with hydroelectric power could be established on top of Kittatinny Ridge. None other than Supreme Court Justice William O. Douglas spoke out against the destruction of Sunfish Pond. In fact, he led a hike to Sunfish Pond that brought national attention to the project. Sunfish Pond was saved. But new attention was focused on the impact the dam would have on the river itself and the surrounding area.

The Army Corps had been in the process of acquiring the privately held land along the river since the Flood Control Act of 1962 and in the process made quite a few enemies. The most vocal local and national opponents combined to form a coalition group called Save the Delaware. This coalition, which included the Sierra Club, Trout Unlimited, and the National Wildlife Federation—among others, became such a powerful voice that Congress began to listen. Federal legislation caught up with the environmental movement in the form of the National Environmental Policy Act of 1970. The Army Corps was required to research the environmental impact the dam would have on the surrounding area. Not surprisingly, it was concluded that the dam would have potentially devastating affects on water quality and the overall health of the adjacent area. The death knell was sounding for the Tocks Island Dam!

No one factor can stand alone as the reason the Tocks Island Dam project failed. Rather, a combination of all the issues mentioned worked together to prevent the dam from being realized. Though the dam was not officially de-authorized until 1992, the project was essentially killed forever with the Wild and Scenic River Act of 1978. The inclusion of the Middle and Upper sections of the Delaware River in this act guaranteed that the cultural and natural resources that would have forever been altered or destroyed had the dam been built are now protected and preserved indefinitely.

What we know today as Delaware Water Gap National Recreation Area is far from what planners envisioned over thirty-five years ago when Congress authorized the establishment of this National Park System unit. Though the Tocks Island Reservoir never came to be, the natural and cultural resources that were saved combine with the diverse recreational opportunities available at the Recreation Area to create one of the most dynamic and varied units of the National Park System.

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In the 19th and early 20th centuries, the area around what is now the Bushkill Access Area was the site of several hotels, the earliest dating from 1838. The later Riverside Hotel catered to “wheelmen” and advertised (in 1902) its access to “the finest natural shale roads, level and good for wheeling at all seasons.” In 1898, Philip S Rosenkranz operated a ferry in this area after moving the site about two miles upstream from Wallpack Bend (Decker Ferry). By the 1920’s the ferry had a capacity of two cars; it was the last one to operate in the upper Delaware Valley. Service ended on May 4, 1945, when a student pilot from Stewart Army Air Force Base, New York, accidentally cut the ferry’s guide cable during a training flight (Henn, 1975; Clemensen, 1996).
In 1996 a small log was retrieved by John Wright (NPS) from a deep pit dug near the restroom facility at the Access Area prior to its construction. The log was found at a depth of 21 feet, lying on gravel. Overlying materials consisted of 10 feet of channel-fill silt and clay overlain by distal and proximal levee sand. It yielded a radiocarbon date of 4,105 ± 90 yr B.P. (GX-22942).

Leave LUNCH STOP. Be careful crossing US 209 to parking area at shale pit where buses are waiting. Continue north on US 209.

0.1 21.6 Long exposure of lower Mahantango shale to left. Shale-chip rubble has been cleared out to form parking area. (See STOP 10 for discussion of importance of this shale-chip rubble to the “River Road” from here north to Matamoras). The over-steepened slopes in the weak shales and siltstones along this part of the Delaware River are the result of deep scouring by the Wisconsinan ice sheet (and probably even earlier glaciations)—similar to El Capitan in Yosemite Valley! The steeply southeast-dipping cleavage abetted the shape of the slope.

0.5 22.1 Another long exposure of Mahantango shale to left.

0.3 22.4 The elongated pond on the right is a beaver pond, formed where the busy little critters have constructed a dam across a narrow and shallow channel. These channels are common along the outer margins of postglacial stream terraces, typically forming where the backslope of the terrace meets the valley wall.

0.4 22.8 Small sand and gravel pit on left in outwash-fan deposit. To the right are a broad postglacial stream terrace and a good view of Wallpack Ridge in New Jersey.

0.3 23.1 Bear left onto park road (the former “River Road”) following signs to Toms Creek Picnic Area. Climb low rise onto large alluvial fan laid down at the mouth of Toms Creek. The surface of the fan is cut by two successively lower fans that show it is of multiple ages. These fans are graded to postglacial stream terraces in the main valley.

0.4 23.5 Little Egypt Road to left.

0.1 23.6 Cross Toms Creek. In the late 1700’s, the Nyce brothers built a grist mill here near the old road and the area became know as “Egypt Mills” by the pioneer settlers (because they obtained flour and meal at the mill and carried it home on pack horses, like the “brethren of Joseph” in the Old Testament) (Henn, 1975). The name still appears on the most recent USGS topographic map.

0.1 23.7 Toms Creek Picnic Area to left.

0.2 23.9 Park buses near entrance to old shale pit on left. Follow road into shale pit and gather near the back of the pit.

STOP 10. TOMS CREEK SHALE PIT: PLEISTOCENE SHALE-CHIP RUBBLE DEPOSITS, POSTGLACIAL STREAM SEDIMENTS, AND PRICKLY PEAR CACTI.

Leaders: Ron W. Witte and Jon D. Inners.

INTRODUCTION

STOP 10 is located in the southwest corner of the Lake Maskenozha 7 ½- minute quadrangle in Delaware Water Gap National Recreation Area (Figure 196).
Historical Context

At the base of the cliffs on the western side of Minisink Valley, there are thick deposits of shale-chip rubble that were mined to build River Road. I. C. White (1882) wrote:

- Immense quantities of debris from the higher portions of these cliffs have accumulated along the lower slope, making great heaps of small fragments broken by frost and the friction of sliding down the cliffs...

...which are much used on the public roads under the name of “slate gravel”. .... The unrivaled excellence of the Delaware Valley road between Matamoras and Bushkill being due to the fact that vast beds of this “slate gravel” are formed along its entire extent so convenient to the road that hauling is often unnecessary for several miles at a stretch.

Between Bushkill and Milford, there are a dozen large pits and several smaller ones that supplied the slaty material (Figure 197). The original construction of River Road began around 1750. By 1810 it had been completed between Bushkill and Milford (Henn, 1975). The New York Sunday Tribune called the road a “Wheelman’s Paradise” and compared the smoothness of its surface with macadamized roads such as the Merrick Turnpike in Long Island (Henn, 1975). At the height of the biking sport, over 600 cyclists were reported to have passed through Milford, on Memorial Day, 1896 (Henn, 1975). The thick deposits of shale-chip rubble and their fortuitous location on the west side of Minisink Valley, played an important role in the settlement of these lands, and its later growth by providing a ready source of natural aggregate to build River Road.
Shale-chip rubble or as it is known by its other names “slate gravel”, or “sharpstone colluvium” is common in Pennsylvania (Figure 197). It is not common, however, in areas glaciated during the last Ice Age (late Wisconsinan). The shale-chip rubble of Minisink Valley is the only large deposit found north of the late Wisconsinan glacial limit (Figure 198). The rubble, well described in Sevon et al., (1989), consists of angular, elongated, platy, prismatic and bladed clasts of the Mahantango Formation.

Average clast length varies between one and six inches. Larger clasts, up to boulder size, may be interspersed throughout the deposit. Typically, the rubble has very little matrix, although many of the clasts exhibit a thin coating of clay. The few beds that do have a substantial matrix component display a coarsening upwards of shale clasts, suggesting that they were deposited as a slurry flow. Bedding is slope parallel, and averages between one to four inches thick. However, in many places the homogeneity of the rubble makes it difficult to discern bedding. Most of the elongated fragments are oriented down slope. Bedding, sorting, and clast orientation of the rubble suggests that most of this material, after it has fallen off the outcrop and accumulated at the top of the apron, moves downslope as a massive sheetflow. Bedding and grading show that this downslope transport is episodic and in some cases may have involved water.

Glacial erosion and the lithology and structural elements of the Mahantango Formation have created a geologic setting that is conducive to the formation of very large volumes of shale-chip rubble over a short time. Glacial erosion over the course of at least three glaciations has cut back the west side of Minisink Valley and formed a very steep rock face that is as much as 500 feet high. Mechanical weathering of the rock by frost shattering has formed an extensive apron of shale-chip rubble that has accumulated since Minisink Valley was deglaciated about 18,000 years ago. The steep southeast-dipping cleavage of the Mahantango Formation, which is well formed at the Toms Creek pit, the thin, northwest-dipping beds of shale and siltstone, and the vertical joints form weak zones and the extensive surface area required for rapid
fragmentation. Places on the cliff face where vertical joints intersect cleavage are especially prone to attack by frost riving (Figure 199). Weathering of the rock surface in these areas results in the fragmentation of the rock as a series of sheets that propagate away from the joint. This step-like exfoliation or sheeting occurs across the cliff at Toms Creek, and it appears to be the major mechanism in which shale-chip rubble is made. The size of the rubble clasts is directly related to cleavage spacing, bedding thickness, and joint penetration. The shale-chip rubble formed at Toms Creek (Figure 200) is typically bladed in habit (‘pencil shale’) due to the local closely spaced cleavage.

Near Dingmans, located twelve miles upstream, rubble clasts are more platy because cleavage is absent, or at least poorly developed (Figure 201). Sandy beds typically form platy or flaggy fragments when weathered, because cleavage in these coarser beds is typically more widely spaced than the shale and siltstone beds, and bedding thickness is greater. Clast habit is also a function of joint spacing and depth. Most of the joints at Toms Creek are nearly vertical, have a cross-strike orientation, and only penetrate the rock face a few inches to several feet.

**SHALE-CHIP RUBBLE AND OTHER DEPOSITS**

**STOP 9** consists of three sites—*a*, *b*, and *c* (Figure 196). Due to the large number of field trip participants, please spread out and break off into smaller groups. *Sites a, b, and c* do not have to be visited in sequential order.

*a. Shale-chip diamicton and postglacial stream sediments.*

*Site a* is located 350 feet southwest of the pit entrance along Toms Creek Road (Figure 196). Here a road cut at the distal edge of the shale-chip rubble apron reveals a four-unit stratigraphy that consists of shale-chip rubble and diamicton overlying postglacial fluvial overbank deposits (Figure 202).

The lowest unit (d) consists of more than 29 inches of planar, thinly bedded, very fine sand (10 YR 4/3) and silt (7.5 YR 4/4), with minor beds of fine sand (10 YR 5/3). A few subangular and subrounded shale pebbles less than 0.5 inch in diameter are in some of the coarser-grained sand beds. In places, reddish lamella (5 YR 4-3/4) a product of oxidation, are found along the contact between an upper more porous sand bed, and a lower, much less porous silt bed. Bedding in the upper 18 inches of
unit (d) is highly deformed. Bedding contacts have irregular to wavy traces that vary from 0.25 to 4.0 inches. Pinch-and-swell bedding is common and near the top of the unit, there are a few folded beds. Intensity of deformation attenuates rapidly with depth. Deformation in the upper part of unit (d) was due to compaction and dewatering caused by the weight of the overlying colluvial mass. The alluvial materials here lie about fifteen feet above the Qst2 terrace, which forms the high-standing terrace across US 209. This height, absence of paleosols, and indication of episodic overbank sedimentation shows that unit (d) is part of the Qst3 terrace, the oldest flood plain deposits in the valley, and not glacial outwash.

Unit (c) is 14 inches thick and consists of massive very fine sand and silt with a few deformed beds and lenses of very fine sand. Angular shale pebbles make up about 3 percent of the sediment by volume. These pebbles are typically concentrated at the base of some of the poorly formed beds. Unit (c) is a transitional unit that consists of a mix of alluvium and shale-chip diamict. It marks a zone of interbedding and mixing that occurred at the edge of the colluvial apron.

Unit (b) is more than 3 feet thick, and it consists of very fine sand and silt that contains by volume 5 percent shale clasts. Shale clasts vary in habit from pebble-size blades and chips to cobble-size platy fragments. Most of these lie on their flat sides, dip downslope, and exhibit faint bedding. This unit appears to be a distal facies of the shale-chip rubble apron formed where wind-blown sand and silt became mixed with the rubble at the edge of the apron.

Unit (a) forms the surface deposit. It is shale-chip rubble that consists of more than 25 percent by volume of small bladed and prismatic shale chips and some larger more platy clasts. Very fine sand and silt form a loose matrix. The hillslope above the road bank slopes about 28°. Farther up, where the slope becomes steeper, the volume of shale chips greatly increases. The large boulder, five feet to the right of the outcrop, appears to lie at the base of unit (a). It is a large joint block that made its way to the edge of the apron by either rolling or sliding down the steep face of the apron.

b. Mahantango outcrop and the formation of shale-chip rubble.

Site b is located on the far-left (southwest) side of the pit about twenty-five feet above the pit floor (Figures 196 and 199).

Shale-chip rubble is the product of intense frost shattering along the cliff face. The size and shape of the fragments and rate of fragmentation is controlled by the attitude and spacing of cleavage, joints, and bedding and grain size. Bedding (N64°E/32°NW) and cleavage (N70°E/54°SE) largely control the shape of the fragments. In places where cleavage is closely spaced and bedding is thin, fragments will be bladed to prismatic. In places where cleavage is more widely spaced or absent, platy fragments are more common. Grain size also controls size and shape of fragments. The finer-grained
shale and siltstone beds form thinner fragments, while the coarser, sandstone beds form thicker fragments.

The subvertical, nonpenetrative, cross-strike joints found throughout the Mahantango formation hasten rubble formation by providing additional surface area and access to cleavage partings. This results in the disintegration of rock by cross-cliff sheeting, a step-like removal of rock away from the joint face (Figure 203). Shallow joints result in the weathering of thinner sheets, while deeper joints result in the weathering of larger sheets. At times, large joint blocks also become dislodged from the cliff. Several of these may be seen throughout the pit, either partially buried in the excavated rubble or lying on the modern rubble apron, which formed after the pit was abandoned. Up close, rock fragments are loosened individually or in small aggregates. Typically, there is a joint-side downward rotation of the fragment (Figure 203). Eventually, the tilted fragment dislodges and falls to the apron’s surface. Another form of small-scale fragmentation includes flaking, a process where small rock chips and flakes on the exposed cleavage face become dislodged. This type of fragmentation occurs around shallow surface fractures, many of which were glacially formed.

c. **Glacial erosion and puzzling vertical striations.**

*Site c* is located on the far-right (northeast) side of the pit, about 100 feet above the pit floor (Figure 196).

Glacially polished rock and glacial striations are rare because the “rubble machine” has been very effective in weathering the cliff face. The glacially polished and sculpted rock, downvalley striations, and *sichelwannen* preserved at this location strongly hint that this part of the cliff was not exposed to weathering as other parts of the cliff were. In most of the shale pits, fresh to lightly weathered, glacially scoured rock occurs at the base of the cliff where it was quickly buried by rubble following deglaciation. Preserved polish and striations higher on the cliff face would have taken much longer to be covered were exposed to weathering for a much greater length of time. Perhaps stagnant ice may have protected the cliff. Another explanation is that weathering of the cliff face was lessened in areas that had fewer joints, resulting in decreased rates of weathering.
Striations at site c have two orientations. The first set nearly parallels the cliff face in a downvalley direction. These are very faint, consist of scratches to small grooves, and are tilted upwards in a downvalley direction by as much as 20° from a horizontal plane. These features are glacially cut; their tilted orientations were due to outward and compressive flow at the valley side, caused by thicker ice in the center of Minisink Valley. The second set of striations has a subvertical orientation (Figure 204) that varies about ten degrees from the vertical. They occur only on glacially polished rock, and in a few places crosscut the first set. They are very fine, are continuous for only a few inches or less, and they were not found elsewhere in the valley. How did they form? Are they a product of glacial erosion or postglacial slope processes? Individual rock fragments sliding down the cliff face, or fragments at the slip face of a larger slide may have formed these striations. Because both of these processes are common, it is puzzling that the striations were not observed elsewhere in the valley. A second possibility is that the scratches may have formed when debris-rich stagnant ice along the cliff face collapsed, due to undermining by meltwater. Because the shale and siltstone here is relatively soft, quartz sand embedded in the ice may have cut the striations as the ice slid down the valley wall. Because the effects of glacial erosion are so well preserved here, high above the pit floor, stagnant ice may have initially protected parts of the cliff from erosion. Perhaps, you have a better idea about the formation of these enigmatic scratches.

Other glacial erosional forms include several large crescentic gouges (Figure 205). These formed where a large chip of rock was dislodged at the base of the glacier. They typically form where shallow surface fractures, a product of glacial loading, cut the rock face.

CACTI ON MINISINK ESCARPMENT

Cacti were first discovered on the Minisink escarpment in the 19th century, reportedly by a Dr. Barrett of Port Jervis (White, 1882). In fact, White observed that "cactus plants of the species Opuntia Vulgaris" cover the cliffs of "Hamilton sandstone" (Mahantango Formation of modern usage) in Westfall Township, Pike County, between Matamoras and Milford (p. 195-196). Nearly a hundred years after White’s report, R. C. Smith, II, of the Pennsylvania Geological Survey—in searching this same area for lead, zinc, and copper mineralization (also, but mistakenly, reported by White)—rediscovered the cacti (Smith, 1973). Smith’s botanical sources identified the plants as Opuntia compressa (Salisb.) Macbr. (O. vulgaris), the prickly pear cactus. He noted that such cacti had recently been reported in New Jersey, but that there was no correlation of the occurrences to slope, soil composition, soil pH, organic matter, or other plant species. His sources stated that “this variety of prickly pear makes its home on rocks, sand dunes, or sandy prairies from eastern Massachusetts to southern Ontario and southern Minnesota and south to Georgia, Alabama, and Missouri”—but is not found in the far west (Smith, 1973).
Prickly pear cacti have also recently been found in some profusion near the top of the escarpment above the southwest end of the Toms Creek shale pit—GPS location 41°07′44″N/74°57′08″W at an elevation of approximately 750 feet (Figure 206). The cacti here occur at the very top of the shale-chip rubble slope where the deposit feathers out against a low shale cliff, in the lowest part of the cliff face, and at the very top of the cliff. The roots of most of the plants appear to be anchored in small fractures in the bedrock, though some cacti are growing where the rubble is very thin (less 6 inches thick). Many of the plants are concentrated in thin shale and rubble strips extending about 50 feet downslope from the main bedrock cliff, with the intervening strips filled with “mobile” shale-chip rubble. At least one of these strips is directly downslope from a large tree that blocks rubble from covering the area (Figure 206). The slope of the rubble surface in the area of the cacti is about 30°; the face of the cliff is along cleavage (attitude N72°E/66°SE).

Rhoads (1993) and Rhoads and Block (2000) identify the prickly pear cactus that occurs in Pennsylvania as the eastern prickly pear, *Opuntia humifusa* (Raf.) Raf. It is described as an herbaceous perennial that puts out bright yellow flowers in July (Figure 206) and produces a red to red-purple fruit (“pear”) that is fleshy, many seeded, and 3-5 cm long (Rhoads and Block, 2000). *O. humifusa* grows on dry, shaly cliffs and barrens and is reported from numerous places across the Commonwealth, including—in addition to Pike—Bucks, Montgomery, Berks, Lancaster, Cumberland, Franklin, Bedford, Huntingdon, Snyder, and Allegheny Counties (Rhoads, 1993). In the early 1980’s, it thrived on a porous siltstone rubble slope at a fossil site in the Trimmers Rock Formation southeast of Danville, Montour County (site described, but cacti not mentioned, in Hoskins et al., 1983, p. 137-139)—but the plants apparently disappeared from this area several years ago.

Leave STOP 10, turning left onto park road.

0.1 24.0 Stop sign. Turn left on US 209 North.

0.2 24.2 To left in the woods is more of the old River Road. From this point north numerous stretches of the old road can be seen roughly paralleling US 209.

1.0 25.2 George Nyce farm to left (see Henn, 1975). An 18th-century stone house on this property is constructed of blocks of Onondaga limestone and Schoharie siltstone. The farm buildings rest on glacial outwash, thinly covered by shale-chip colluvium. On the right beyond the entrance to Eshback Access Area are two postglacial stream terraces. The higher terrace (Qst3) remnant is located upstream from the access entrance where it lays against a lower terrace (Qst2). Hand auger holes show that these deposits consist chiefly of very fine sand, and fine sand. The few surface cobbles and pebbles presumably have
an anthropogenic placement. Although some of this material may be ice-rafted, carried onto the terrace during winter floods. (See Witte, this guidebook, p. 99.)

0.9 26.1 The farm to left is “Wheat Plains,” homestead of the Brodhead family. A farm was established here in the 1770’s by Gabriel Brodhead, a Revolutionary-War soldier, and bequeathed down through six generations (passing out of the family’s hands only between about 1871 and 1894) (Henn, 1975).

0.3 26.4 Cross Elfin Gorge creek. To left, just to the south of the bridge, is a parking lot from which a trail leads back to the beautiful Elfin Gorge falls (over Mahantango shale and siltstone) and the provocatively named Diana’s Bath plunge-pool just below. The tract along the creek northwest of US 209 was once the homestead of the Van Gordon family. One of the family daughters, Savanah, married Henry C. Ford, who as Chairman of the State Fishery Commission in the late 1800’s, was instrumental in restoring the once much-depleted shad fishery to the Delaware River (Henn, 1975).

0.6 27.0 Briscoe Mountain Road (to Pocono Environmental Education Center) to left.
0.1 27.1 To left is a beautiful stone arch bridge on the old “River Road.” Postglacial stream terrace (Qst3) remnant on right. Hand auger in field showed 3 ½ feet of silty very fine sand over presumably a cobble-gravel lag.

0.2 27.3 Big pit exposing Mahantango shale to left.
0.3 27.6 Cut in lower Mahantango shale to left.
0.4 28.0 Another cut in Mahantango shale to left.
0.2 28.2 Abandoned Mahantango shale-chip-rubble pit to left. Outwash terraces on right (60 to 80 feet above river). Old sand and gravel pit (1500 feet east of US 209) revealed about 12 feet of fine cobble-pebble gravel overlying sand and pebbly sand. The outwash terrace surface is gravelly, unlike postglacial stream terraces, which may have only a few pebbles and cobbles.

0.4 28.6 Park maintenance and storage yard to left on floor of large shale pit. More than 50 feet (angle of repose) of framework-supported shale-chip colluvium is exposed at the pit’s north end. Average size of chips is about 1.5 in., and there is very little matrix. Layering is slope parallel. Postglacial stream terraces on right.

0.6 29.2 To left is an unmarked gravel lane leading to a parking lot at the head of a well-maintained trail along Hornbecks Creek. Follow this trail for a mile through the woods and you come to two picturesque waterfalls over ledges of Mahantango shale and siltstone—Cascade Falls and Indian Ladder Falls (see Henn, 1975). (The latter is well named for it consists of many low falls in the gorge above the 25-foot-high Cascade Falls.)

0.2 29.4 Cross Hornbecks Creek.
0.4 29.8 To right near the highway is a remnant of Qst3 (highest postglacial stream terrace in valley). A hand auger hole revealed more than four feet of very fine sand with no gravel. In some places the flanks of these higher terraces are gravelly suggesting that the overbank materials are thin, possibly removed by erosion as the river cut down to a lower level. The high bank to the left is part of the Hornbecks Creek outwash fan.

0.2 30.0 To the right is a low bedrock knob (Mahantango Formation) that was a small nearshore island in the glacial (and early postglacial) Delaware River. On the east side is a steep scarp, along the top of which are numerous moderately dipping (25° SW) ledges of dark gray, silty shale (with very thin sandstone beds) (see Sevon et al., Plate 1). The scarp borders an old braid channel of the river. A narrow kame extends to the south in the hill’s protected “current shadow” (Crowl, 1971). The knob was at one time known locally as “Bald Hill” because its thin soil cover supported few trees (see Henn, 1975).
Bald Hill Farm on left. This was the former site of Pine Hill Golf Course in the early 1900’s (Henn, 1975). High terraces to the right near highway are meltwater-stream terraces.

Large abandoned shale pit in Mahantango to left.

Another large Mahantango shale pit to left.

Wilson Hill Road on left. On right just before the road is an interesting cobblestone building. While such buildings are common in the glaciated terrain of central New York (McKinney, 2000), this is one of the few, if not the only one, in this area.

To get to Dingmans and Silver Thread Falls turn left on Johnny Bee Road (TR 325) here and follow signs.

Cross Dingmans Creek. High-standing terrace on left just past the bridge is part of the Dingmans Creek outwash fan.

Traffic light at intersection with PA 739. Continue straight ahead.

Historical Marker to right reads:

**DINGMAN’S FERRY.** Here was located one of the earliest ferries across the Delaware. Andrew Dingman in 1750 built the flat boat he used as a ferry with his own handaxe. Dingman was one of the pioneer settlers.

To right, the classic building with four Doric columns across the front is the former Dutch Reformed Church, erected 1850. Just beyond on the left is Delaware Cemetery. It lies in outwash and shale-chip colluvium. Buried here among many pioneer families and more recently deceased individuals is Thunder Cloud (1856-1916), a half-breed Blackfoot Indian who served as a U.S. Army scout in the Sioux Wars of 1872-76 and toured with Buffalo Bill’s Wild West Show. Later he became the foremost Indian model in the country, his likeness appearing in such famous works as “The Treaty of Traverse de Sioux” (a mural in the Minnesota State Capital at St. Paul) and “Father Hennepin discovering the Falls of St. Anthony.” He married the artist Henrietta Hasagen and lived on Mill Street in Dingman’s Ferry (Henn, 1975)

Higher terraces to the left are outwash. Postglacial steam terraces to the right.

Cross Adams Creek. A hike up this stream, though rather rigorous, is well worth the effort. About 1.2 miles upstream is a spectacular gorge and waterfall (called Liberty Falls by Henn [1975]) in upper Mahantango shale and sandstone. Downstream from the gorge the stream valley exhibits at least two outwash/alluvial terrace levels. About 750 feet northwest of US 209, the creek runs through a low, 500-foot late glacial/postglacial bedrock gorge, the old channel appearing as steep, cobble-covered slopes just upstream and downstream on the southwest side of the gorge.

Climb back onto high-standing outwash terrace.

Cross over Dry Brook.

Pass Zimmerman Road on left. Nearby terraces are outwash while those farther out in valley are postglacial stream deposits.

Start descent toward Conashaugh Creek. High-standing terrace on the left is the Conashaugh Creek outwash fan.

Cross Conashaugh Creek.

To left is another typical cut in the lower Mahantango with low banks of shale-chip rubble.

Large shale-chip-rubble pit to left.

Good view of Mahantango cliffs on the Minisink escarpment ahead. Road cut on left in Raymondskill Creek outwash fan. Low terraces to right are postglacial age.
STOP 11. RAYMONDSKILL CREEK SURFICIAL SECTION: GLACIAL AND POST GLACIAL GEOLOGY AND AMERIND HISTORY.
Leaders: Ron W. Witte and John Wright.

INTRODUCTION

STOP 11 is on the northwest side of the Minisink Valley in a reentrant at the mouth of Raymondskill Creek in the upper part of the Delaware Water Gap National Recreation Area (Milford 7 ½ - minute quadrangle; Figure 207). In the mid-nineteenth century, Henry Depue developed the Raymondskill House on this site as accommodations for raftsmen and possibly hunters. Around 1880, Emile Schanno bought the property and renamed it the Schanno House. In 1892, the three sons of Emile Schanno enlarged the hotel to service the coaches that traveled River Road. Because of its reputation for fine dining it became known as the “Little Delmonico” (Henn, 1975). In 1989, the hotel, long abandoned, burned. The damaged structure was razed by the National Park Service in 1995.

Raymonds Kill is also the site of an infamous skirmish where thirteen colonialists were killed trying to attack a small group of Indians who had retreated up the Raymonds Kill. Apparently, two thirds of the attacking party fled after being ambushed near Bastian Spring, leaving the rest to unsuccessfully fend off the attack (Henn, 1975).

Geologic Setting

Minisink Valley is a deep, glacially scoured river valley underlain by a complex stratigraphy consisting of late Wisconsinan glaciolacustrine, glaciofluvial, and postglacial alluvial sediment. The low ridge across the river (eastward) is Wallpack Ridge. The high cliff on the western side of Minisink Valley is held up by the Mahantango Formation, which is Upper Devonian age and consists of repetitive sequences of shale and siltstone.

A high-standing terrace to the south, locally known as Indian Point, rises about 70 feet above the parking area. The terrace is the remnant of a large outwash fan (Qf on Figure 207) that was laid down at the mouth of Raymondskill Creek during deglaciation. At one time, the fan filled the reentrant. Later, it
was eroded down to its present level by meltwater and postglacial Raymondskill Creek. High, gravelly terraces up Minisink Valley are part of the Montague valley train, laid down from an ice margin position about two miles upstream from Raymondskill Creek. The low broad plain that forms the valley floor across US 209 is an abandoned flood plain of Holocene age (Qst2 on Figure 207). The terrace lies 20 to 30 feet above the Delaware River (measured above the river’s mean annual elevation here of 370 feet). The terrace is underlain by overbank deposits of fine sand and silt that are as much as 20 feet thick. In places, the Qst2 deposits comprise several levels that represent flood-scoured areas, and younger Qst2 subsets. These terraces of intermediate elevation, found between the Qst2 terrace and the modern flood plain, are found throughout the valley. Most of these appear to be related to local riparian conditions controlling erosion and deposition.

A recent slump, near the top of bluff overlooking the parking area, exposed a 15-foot section of outwash (area mapped as Qf, Figure 207). We won’t climb there today. (What, and miss another opportunity to dirty our hands in glacial drift!) However, Indian Point affords an excellent view of the Montague valley train and the lower meltwater and postglacial terraces in Minisink Valley. Sediment exposed along the face of the slump consists of four feet of planar-bedded cobble-pebble gravel overlying, pebble gravel, and pebbly sand with minor cross-bedded sand. Clasts chiefly consist of gray shale, siltstone, and sandstone. The direction of paleoflow is east to southeast, based on the orientation of elongated clasts and crossbedding. This shows that the outwash was laid down by meltwater draining the Raymondskill Creek drainage basin, rather than the Delaware mainstem. The dominant lithologic suite of shale, siltstone and sandstone clasts also shows that most of the material here was largely derived from local upland sources. Small exposures dug out below the slump revealed cross-stratified sand, pebbly sand, and pebble gravel. Topography of the fan’s south side suggests it was laid down against stagnant ice. However, the fan appears to be graded to the surface of the Montague valley train (see Figure 63c in Witte, this guidebook p. 108), showing that the ice was only of local extent and did not fill the valley as a tongue of dead ice.

**Question.** Does the distribution of valley-train and outwash-fan deposits in Minisink Valley indicate deposition against large blocks of stagnant ice, or have meltwater and postglacial streams eroded most of this material?

**Directions to Raymondskill Bluff**

Carefully cross US 209 (does the video game *Frogger* seem like an appropriate analogy here?) and follow a fishing-access road across the Qst2 terrace toward the river. At large right-hand bend in road (about 450 feet from US 209) turn left (north) and follow the footpath through wooded area to the bluff overlooking Raymondskill Creek. Please note that the bluff face is extremely precarious. Stay back at least 5 feet from its edge. The low bluff (Figure 208) exposes about 15 feet of overbank sand and silt, replete with buried A and B soil horizons that preserve a history of episodic alluviation, vertical flood plain accretion, and prolonged periods of land stability marked by soil formation. Beneath this material is coarse gravel and sand, part of an old alluvial fan. Minisink Island (Figure 207), across the Delaware, is also a Qst2 terrace, except for a few narrow strips of modern flood plain that lie on the flanks of the island.
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. In contrast to the late glacial river, the modern Delaware is a meandering stream of very low sinuosity (so low that many sections of the river channel are straight) flanked by two abandoned flood plains. These flood plains are mapped in the valley as Qst3 and Qst2 terraces (see Witte, this guidebook, p. 99). The deep, dated, alluvial sequences at the Shawnee-Minisink (McNott et al., 1985), and Upper Shawnee Island (Stewart, 1991) sites show that the form of the Delaware River at the end of the Pleistocene was also a non-braided one with a well-established flood plain. The river was also at or slightly above its present elevation. For the sake of brevity in this discussion, the postglacial form of the Delaware will be referred to as meandering.

RAYMONDSKILL BLUFF

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. In contrast to the late glacial river, the modern Delaware is a meandering stream of very low sinuosity (so low that many sections of the river channel are straight) flanked by two abandoned flood plains. These flood plains are mapped in the valley as Qst3 and Qst2 terraces (see Witte, this guidebook, p. 99). The deep, dated, alluvial sequences at the Shawnee-Minisink (McNott et al., 1985), and Upper Shawnee Island (Stewart, 1991) sites show that the form of the Delaware River at the end of the Pleistocene was also a non-braided one with a well-established flood plain. The river was also at or slightly above its present elevation. For the sake of brevity in this discussion, the postglacial form of the Delaware will be referred to as meandering.
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 stream 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 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, after deglaciation, the Delaware River cut down 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 that the gravel beneath the postglacial stream terraces is glacial outwash, laid down by meltwater during the latter stages of deglaciation of the upper Delaware River drainage basin. Based on the oldest dates at the Shawnee-Minisink and Upper Shawnee Island sites (see Table 4, p. 114), 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 sufficiently 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 postglacial stream-terrace deposits is not glacial outwash, but outwash that had been reworked and incised by the postglacial Delaware River.

Postglacial alluvial terraces in Minisink Valley consist of two abandoned flood plains that cover large parts its floor. These are designated Qst3 and Qst2. The higher position, slightly coarser strata of the Qst3 terrace show that it is older than the Qst2 terrace and that it was laid down by a river that was higher than it is today. The Qst3 terrace represents the oldest flood-plain deposits in Minisink Valley. They were probably laid shortly after ice had withdrawn from the Delaware Valley, about 15,000 to 14,000 yr B.P. The lack of multiple buried soils show that the period of Qst3 deposition was probably short-lived. Incision to the Qst2 level may have been initiated by the onset of delayed isostatic rebound, and a reduction in sediment supply due to the transition from 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) may be in accordance with the estimated age of the Qst3 terrace.

The Qst2 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 so limits the elevation of its surface.” Due to the narrow width of Minisink Valley and low sinuosity of the Delaware River channel, the Qst2 flood plain outgrew its fluvial setting, receiving sediment only during the greatest of floods.

Deep, well-exposed, alluvial sections, like the one at Raymondskill bluff are rare in Minisink Valley. Most of the information about these postglacial fluvial deposits comes from the many archaeological investigations conducted over the last fifty years. Because all these investigations involved the digging of pits, recurring visits to the study sites, once the work was completed, only yielded a filled excavation. Fortunately, Raymondskill Creek changed its course about thirty years ago, providing archaeologists, soil scientists, and geologists with a highly informative outcrop, and causing a dilemma for cultural and natural resource preservationists.
Raymondskill Creek, a Change in its Course

Delaware Water Gap National Recreation Area contains numerous archeological sites, several of which are included in archeological landmark districts. One archeological landmark district, the Minisink Historic Landmark District, is comprised of 19 archeological sites and one standing historic structure. This district is important due to its ability to yield information that has significance on the historic contact between the Indian and European people in Munsee County, a region which encompasses southern New York through northern New Jersey to northeastern Pennsylvania. The Manna Site (36Pi04), located about 100 feet south of the bluff, is one of the 19 archeological sites in the district. Historically, the Manna Site contains evidence of ancient Native American populations from the Early Archaic cultural period (ca 8000 B.C.) through the Contact cultural period (A.D. 1750).

Stereoscopic air photos from 1979 (scale 1:12,000, on file at NJ Geological Survey, Trenton, NJ) show that Raymondskill Creek flowed through a much straighter channel at its mouth than it does today (Figure 209). Prior to 1979, the creek used a more northerly channel on its way to the Delaware. The exact time of abandonment is unknown. However, the straight nature of the 1979 channel Figure 209) suggests the creek changed its course in the 1970’s, possibly during June, 1972, when the remnant of Hurricane Agnes devastated the Delaware Valley. From 1979 to the present, the creek has carved out a sinuous course, eroding substantial parts of the lower Qst2a and Qst2b terraces in its mouth, and Qst2 along the bluff at STOP 11. Based on a comparison of 1979 and 1997 photos of the creek’s mouth (Figure 209), the bluff face has retreated as much as 50 feet from its 1979 location. A large point bar

STOP 11 - Raymondskill Bluff

Figure 209. Surficial geologic map and color-infrared photograph (1997) of the Raymondskill bluff site, showing the change in the course of Raymondskill Creek near its mouth over the last thirty years.
was also deposited just upstream from our overlook as the creek channel quickly evolved from straight to a meandering one.

Raymondskill Creek has probably changed its course numerous times throughout the Holocene. Based on the amount of erosion that has occurred since 1979, and the current configuration of the channel, Raymondskill Creek will probably cut a new channel (Figure 209) to the Delaware within the next twenty-five years. When this happens, undercutting of the bluff section at STOP 11 will diminish, slowing the rate of bank retreat. Eventually, the abandoned channel may become filled with overbank sediment, a process mostly facilitated by hydraulic ponding during the larger flood events on the main stem. “Ah yes, geologic processes working in harmonious accord with the natural surroundings.” However, “letting nature take its course” does present a specific set of problems. Because the terrace that forms the bluff contains a diverse assemblage of Amerind cultural components that date back to the Late Archaic, including burials, what is the role of the National Park Service regarding the preservation of this site? Do they shore up the bluff face, which would severely diminish the site in terms of its geologic and archaeological value or just let it be?

Stabilization efforts for this type of erosion are complex in nature. One option is the traditional, hardened containment of the stream channel (the utilization of rip-rap). This method is expensive and distracts from the stream’s natural setting. Using this type of stabilization would reduce fisheries habitat and would accelerate stream velocity to downstream reaches. An alternative stabilization technique would involve utilizing natural channel design principles, which in turn is more cost efficient. The lower cost changes would include improving fishery habitats, avoiding downstream impacts and retaining the natural appearance of the stream.

Archeological Setting

The Delaware Water Gap National Recreation Area lies within the Upper Delaware River Basin, a unique drainage containing abundant physical evidence of a rich cultural past. Riparian lands along the Delaware River enticed the earliest North American inhabitants to exploit its resources. Changing climate at the close of the last Ice Age to modern time has fostered varied econiches that maintained a wide diversity of flora and fauna. These plants and animals supported human occupations spanning a continuous period from approximately 10,500 yr B.P. to the present day. This rich history is preserved in both historical and archeological record of the Delaware Water Gap National Recreation Area.

In the late 1950’s, the proposed construction of the Tocks Island Dam stimulated historical and archeological interest in this valley. Historians and archeologists were summoned to identify, record, and salvage data before the valley was inundated. Archeological investigations from 1959 through 1975 involved archeologists and historians from academic institutions from Pennsylvalnia, New Jersey, New York, and the District of Columbia. During this time, these investigators mounted the largest and most complex research programs ever directed to a specific location along the Mid-Atlantic seaboard. The results of their efforts have yielded information concerning the evolution of human settlement and environmental adaptation within the valley over the last 10,000 years. By the mid 1960’s, recognized that this area offered a rich and well preserved record of prehistoric occupation, beginning with the Paleo-Indian, the earliest known culture in the New World. Current theory suggests that during the late Wisconsinan glacial period, a land bridge existed between Asia and Alaska, vanishing around 10,000 yr B.P. Hunter-gatherers migrated across this land bridge following herds of caribou and other large mammals. This early culture is recognized archeologically by distinctive fluted points, called Clovis, which are found in eastern North America as isolated finds. Three sites in the Delaware Water Gap Recreation Area contain evidence of this culture. The oldest radiocarbon dates place the earliest occupation at 8640 B.C. ± 300 years near the Delaware Water Gap.
The intensive archaeological investigations also documented the record of the changing environment through the Archaic Period (8,000 yr B.P. to 3,000 yr B.P.) and how the Archaic populations adapted to those pressures. The data is being used to reconstruct, in more detail than any other location within the Mid-Atlantic area, the Archaic settlement distributions of this region.

Over 478 archeological sites have been documented within the Delaware Water Gap National Recreation Area. Among these sites, several have been placed into specially designated districts based on their unique and nationally significant composition. This district, which includes the area around the mouth of Raymondskill Creek, entitled the Minisink Historic District, received National Landmark status on April 19, 1993. Sites such as these are of extreme interest to scholars of the northeastern United States because they provide information about the immediate ancestors of the Algonkian Indians, who occupied the Atlantic seaboard before European contact in the 17th century. In the Upper Delaware River Valley, archeological evidence shows that the initial Late Woodland occupations represent the ancestral tradition of the Minisink Indians, an Algonkian group of the terminal Late Woodland (Late Woodland Period—A.D. 900 to A.D. 1550). Further research may link the Owasco culture with the New York and New Jersey Owasco cultures. The earliest evidence for domesticated corn in this valley is documented in this period from an archeological site in New Jersey within the recreation area. At this same time, archeological evidence indicates these sites show an increase in size and number, suggesting a rise in population.

Since the Delaware River is not navigable north of Trenton, New Jersey, Euro-American settlement in the Upper Delaware Valley occurred almost a century later than the coastal areas of New York, the Hudson Valley, and the Delaware Bay. Thirty-nine of 478 documented archeological sites within the park are contact period sites or contain contact period cultural components. These sites contain European-made objects traded inland by other Native American Indians. This evidence offers opportunity to study how this introduction changed Native American Indian society in this area.

The first Euro-American settlers of the valley in the early 18th century claimed the land for agriculture. As transportation routes expanded, villages were built to support a growing agricultural base. After the American Revolution, this area witnessed a boom in population growth. There are several of these sites within the boundary of the park which offer an archeological record that reconstructs the life ways of these early frontier settlers. In the post-Civil War period, agriculture declined and was supplanted by recreation affording people the opportunity to view the natural scenic splendor of the Water Gap. Continued research and preservation of archeological and historic sites within the Delaware Water Gap National Recreation Area promises to clarify and aid in the development of a systematic understanding of American prehistory and history in the northeastern United States.

**Evolution of the Holocene Flood Plain at the Mouth of Raymondskill Creek**

The Raymondskill bluff site records episodic periods of alluviation between extended periods of landscape stability marked by buried soils. Radiocarbon dating of charcoal fragments, which were found concentrated along well-defined alluvial strata, show that flood plain materials preserved here may be as old as 5000 yr B.P. Most of the buried soil horizons and alluvial beds could be traced across the face of the bluff. Bedding on the far right side of the bluff (west) dipped westward and in some places pinched out. This geometry suggests that this part of the terrace may represent an overlapping alluvial sequence that covered a lower terrace along the creek. Proof that some of the stream terraces grow by lateral accretion. The left side of the bluff lies truncated against the younger floodplain deposits that form a narrow, low terrace along the flanks of the modern river.

The bluff section may be divided into two parts (Figure 208). (Apologies to those wishing to discuss each buried soil horizon, alluvial bed, and oxidation lamella in detail. Hey, I usually map all this
The lower part (below 3.55 m) consists of a sequence of closely spaced buried A horizons that represent extended periods of soil formation, and infrequent flooding. Overall, alluvial beds here are finer grained, thinner and represent multiple periods of land stability. A date of 4500 ± 40 yr B.P. (GX-28162-AMS) was determined from a piece of charcoal collected at a depth of 4.55 m, about 0.45 m above the base of the terrace.

Above 3.55 m, there are fewer buried soil horizons, and the alluvial beds are typically thicker, and coarser-grained. A date of 3230 ± 40 yr BP (GX-28163-AMS) was determined from a piece of charcoal collected from a very rich charcoal layer, section B. This horizon, projected back to section A, gives a depth of 3.25 m or 0.3 m above the contact between the upper and lower parts shown in Figure 209. Minimal soil development, increased bedding thickness, and increased grain size point to a substantial rise in flood frequency and rate of alluviation in the upper part of the bluff.

Stewart (1991) has observed a similar, but older alluvial pattern in the Delaware Valley showing that depositional rates for the period between 4200 B.C and 800 B.C. were consistently higher than anytime before or after. The cause of the increased alluviation may be due to a change in vegetative cover that coincides with the Holocene Altithermal, a period of rapid climatic fluctuations in which the overall trend was toward a warmer and dryer climate. During this time, vegetative cover may have changed from a closed temperate forest to a mix of forest and patchwork grasslands. This led to increased erosion along the Delaware’s tributaries and increased alluviation along the mainstem (Stewart, 1991). The lower part of the Raymondskill bluff section seems to be too young to coincide with the beginning of Holocene Altithermal (~6000 – 3000 yr B.P.). It does, however, correlate with the late part of the Altithermal, the two sections of the bluff marking the transition from the Atlantic Period (8000 to 4000 yr B.P.) to the Sub-Boreal Period (4000 to 3000 yr B.P.). Clearly, the upper part of the section shows a substantial increase in the frequency and rates of alluviation, presumably representing change in the hydraulic and sedimentologic characteristics of the drainage basin. These changes appear to be climatically driven and not the result of local riparian conditions. Continued work at the 5000-year old-Raymondskill Bluff site will hopefully yield a higher resolution record of alluviation and soil development and comparison to other dated sections in Minisink Valley.

The coarse gravel and sand deposits beneath the Qst2 flood-plain sediments are part of an old alluvial fan laid down at the mouth of Raymondskill Creek. Because the elevation of this material lies well below nearby glacial outwash, it probably is of late Pleistocene age, or early Holocene age deposited well after the late Wisconsinan ice sheet retreated out of the Delaware River drainage basin. The eastward dipping surface of the fan, and its high percentage of clasts derived from local sources (mostly the outwash fan that sits just upstream in the reentrant) show that these materials were deposited by Raymondskill Creek rather than a glacial meltwater stream.

The lower terraces at the mouth of the creek, Qst2a and b (Figures 209 and 210) are no higher than 15 feet above the Delaware, and they consist of massively to thickly bedded fine sand, very fine sand, and silt. Soil development here is very weak, and it appears that depositional rates here are very high based on thickness of bedding. A date of 330 ± 30 yr B.P. (GX-28164-AMS), determined from a piece of charcoal collected at a depth of 2.15 meters (Figure 210) shows that the terraces in the mouth of Raymondskill Creek are very young and that they were built very quickly. The history of channel abandonment, erosion, and high rates of sedimentation at the mouth of Raymondskill Creek show that this area of confluence is a highly dynamic environment, susceptible to rapid change over periods of x10 to x10² years.

If you wish to view the bluff face follow the footpath toward the mouth of Raymondskill Creek and approach the lower part of the bluff from its left (eastern) side. Continue along the face of the outcrop and exit along its right side, climbing back onto the Qst2 terrace. Please, limit scrapping and digging to the cleaned-off areas on the bluff—and remember collecting artifacts is not permitted.
After exiting the right side of the bluff, follow the top of bluff back toward the Delaware and retrace route back across US 209 to the parking area.

Leave STOP 11, turning left on US 209 North to cross Raymondskill Creek. The Historical Marker on right reads:

\textit{WYOMING-MINISINK PATH. Here, an important Indian trail connecting the Delaware and Susquehanna Rivers ascended Indian Point to Powwow Hill. The path was used by Delaware Indians in their migration to the Wyoming Valley, and later by Connecticut settlers.}

0.1 36.8 Turn left on SR 2009.
0.2 37.0 To right over the next 0.2 mile are several cuts exposing gently northwest-dipping, fossiliferous shale and siltstone beds in the Mahantango Formation.
0.3 37.3 Turn left into parking lot at the “Million-Dollar Outhouse.”

\textbf{STOP 12. RAYMONDSKILL FALLS: MAHANTANGO FORMATION, ORIGIN OF WATER FALLS, AND PLANT ECOLOGY OF THE GORGE.}

Leaders: Ron W. Witte, Donald H. Monteverde, and Jeanine Ferrence.

\textbf{INTRODUCTION}

STOP 12 is located on Raymondskill Creek about 0.6 mile above its confluence with the Delaware River, in Dingman Township, Pike County (Milford 7 ½-minute quadrangle; Figure 211).

\textbf{Geologic Setting}

Raymondskill Falls, one of many waterfalls in the Delaware Water Gap National Recreation Area, is located where Raymondskill Creek has cut a deep glen in the Mahantango Formation. The falls consists of three closely spaced plunges that drop 127 feet to the floor of a rock-walled chasm. Raymondskill Creek, spilling over a series of low cascades, enters an oversized plunge basin about 70 feet above the upper falls through a narrow rock-walled gorge (Figure 212). The upper falls (Figure 213) plunges 49 feet over a lip held up by a thick sandstone bed into a small, narrow plunge basin. During periods of low flow, Raymondskill Creek flows through a small notch cut back 15 feet from the upper lip of the falls. The creek quickly exits the middle plunge pool, dropping 38 feet over the middle
falls (Figure 213) into another oversized plunge basin. From here the creek is diverted around either side of a large dislodged joint block boulder to where it plunges 40 feet over the lower falls (Figure 214) into another large plunge basin. Directly across from the lower falls is Bridal Veil Falls, a narrower cascade of similar height, formed where an unnamed tributary meets Raymondskill Creek. The creek flows from the lower plunge basin through a narrow rock-walled gorge, dropping 20 feet over low (< 2 foot) cascades for a distance of about 600 feet. The mouth of the gorge is guarded by a 12-foot high rock promontory that juts out from the north bank of the creek. Several other rocky “shoulders” are found throughout Raymondskill glen, their eroded forms suspiciously resembling parts of the modern falls. Downstream from the gorge, Raymondskill passes through a large valley-side reentrant on a more leisurely course to the Delaware River, a distance of about 2600 feet. Once outside the lower gorge, the bedrock floor of the valley quickly dives beneath thick glacial outwash mantled by thin alluvium.

Bedrock Geology

Although the main topic of STOP 12 is the origin of Raymondskill Falls, it must be admitted—even by the most narrow focused of glacial types—that the local bedrock plays a significant part in this evolutionary story. The water has to fall from somewhere. So what about the bedrock? The host rock of Raymondskill Falls is the middle Devonian-age Mahantango Formation, part of the Hamilton Group. The Mahantango and its equivalent crop out across central Pennsylvania to Port Jervis, New York. North of Port Jervis, the Mahantango correlates to the Skaneateles, Ludlowville and Moscow Formations in eastern

Figure 211. Location of Raymondskill Falls, Raymondskill Creek, and nearby streams (Milford 7 1/2-quadrangle). Filled shaded-contour base map was constructed from USGS 10-m DEM. Preglacial or early glacial course of the upper part of Raymondskill Creek is shown in red.

Figure 212. Rock-cut ravine above upper plunge pool believed to be late Wisconsinan age. On left side of photo above cascade is an abandoned falls. View from upper observation deck. Scale to right of upper cascade is two feet.

Figure 211. Location of Raymondskill Falls, Raymondskill Creek, and nearby streams (Milford 7 1/2-quadrangle). Filled shaded-contour base map was constructed from USGS 10-m DEM. Preglacial or early glacial course of the upper part of Raymondskill Creek is shown in red.
New York and the Hudson Valley (Ver Straeten, personal communication, 2001). It gradationally overlies the various Marcellus units along this strike belt (see discussion at STOP 8).

Recent study of the Mahantango in the Raymondskill area began with Fletcher and Woodrow (1970). They described three members, a more sandy, middle member sandwiched by an upper and lower shaly members. Raymondskill Falls lies entirely within the middle sandy member. Based on mapping farther to the south, Alvord and Drake (1971) mapped an upper and lower member of identical lithology that was separated by the “Centerfield Reef,” a coral-bearing calcareous-siltstone biostrome. Epstein (1973) did not subdivide the Mahantango, but noted the existence of the Centerfield biostrome. Sevon et al. (1989) tried to follow the subdivision of Fletcher and Woodrow (1970), but were unable to trace the units on a regional scale. They noted that the Centerfield does not extend very far from Monroe into Pike County. All these authors indicated that Mahantango was deposited under shallower water conditions than the underlying Marcellus.

Prave et al. (1996) in their study of the regional facies of the Mahantango analyzed the Raymondskill and two other sections approaching Port Jervis, New York. They identified six coarsening-upwards cycles in the Raymondskill Creek section, each of which begins as a mudstone-dominated interval. These grade upwards through hummocky and into hummocky/swaley cross-stratified sandstone bodies capped by abraded fossiliferous wackestone (Figure 215). The mudstone beds characterize normal offshore-shelf deposition. The hummocky cross-stratification signifies a shallowing relative sea level with the sediment deposition now above storm wave base. Fletcher and Woodrow (1970) noted ball-and-pillow structures in the sandstone bodies (Figure 216). These deposits represent storm-dominated shelf deposits off a northwest facing shoreline (Figure 217).
Figure 215. Measured sections along the eastern Mahantango outcrop belt from eastern Pennsylvania and southern most New York (modified from Prave et al., 1996). Sections measured at B = Bowmanstown, R = Raymondskill, S = Sawkill Creek, P = Port Jervis. Sections depict coarsening-upwards succession that represents prograding strata from offshore (1+2) through lower to upper shoreface (3+4) paleoenvironments. A flooding event followed by coarser-grained transgressive lag sediment (7) caps each of the cycles. Hummocky cross stratification (modified from Maill, 2000) indicates that sediment rose above storm wave base during the shallowing cycles. Prave et al. (1996) represent these sections as indicating storm-dominated marine-shelf depositional environments.

RAYMONDSKILL FALLS

Waterfalls are common along many of the tributaries of the Delaware River in Delaware Water Gap National Recreation Area. Many of these are found on the western side of Minisink Valley along a northeast-trending strike belt of the Mahantango Formation. Waterfalls are also the most photographed and visited geologic feature in the Recreation Area. Dingmans, Raymondskill, and the many other falls throughout and near the park (including the commercially operated Bushkill Falls) are places people seek, hoping to provide themselves with at least a moment of serenity in their very hectic lives. Actually, most people seem to visit Raymondskill Falls, because they want to photograph the $800,000 outhouse (Figure 218).

What Got It All Started?

Prior to the onset of continental glaciation in the Northern Hemisphere (> 2.3 ma), the Delaware’s tributaries probably had gentle profiles, the result of millions of years of erosion in a passive continental margin setting. Waterfalls may have existed, but only in places where streams crossed or cut back across rock layers that varied greatly in their ability to withstand erosion. Most of the falls in this area formed on thin, interlayered sequences of shale, siltstone, and sandstone of the Mahantango and

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overlying Trimmers Rock Formations. Because their ability to resist erosion is similar, waterfalls probably did not exist, except for a few, low cascades.

Glacial erosion over the course of at least three glaciations has deeply scoured the floor of Minisink Valley and cut back its walls, especially on the western side of the valley where the Mahantango Formation was readily eroded. Because ice flow was generally directed southward down the axis of the main valley, erosion here was much greater than it was along the tributary valleys, which were oriented obliquely to glacial flow. Witte and Stanford (1995) have estimated as much as 150 feet of valley-bottom scour in the Minisink over the last two (late Wisconsinan, and Illinoian) glaciations, and Braun (1989) has suggested that as much as 450 feet (150 m) of land may have been removed in eastern Pennsylvania due to glacial erosion.

Due to glacial scour, preglacial tributaries were truncated by glacial erosion where they entered the Minisink. After the first ice sheet retreated northward, the Delaware’s tributaries, now found themselves in hanging valleys and falling over a cliff in a waterfall that was not there prior to the

Figure 216. Example of ball-and-pillow structure photographed from the upper falls observation platform. View is looking northeast. Due to density contrasts between the heavier sandstone bed and lighter siltstone bed, the sand has settled into and displaced the silt. This is characteristic of soft sediment deformation that occurred prior to lithification. Sand bed is approximately 1-1.5 feet thick.

Figure 217. Paleogeographic reconstruction of a northwest-facing shoreline during Mahantango deposition (from Prave et al., 1996). Lettered locations are same as sections in Figure 215. Number regions correspond to different facies identified in measure sections. Prave et al. (1996) identified two other coarser facies (5+6) in central Pennsylvania. Facies 5 contains tidal-channel-mouth shoals and facies 6 consists of subtidal to intertidal flats and tidal channels. This two facies represent farther shoreline progradation than observed here at Raymondskill and indicate deposition under tide-influenced conditions.
glaciation. Over time, erosion of the falls face by hydraulic plucking formed a knickpoint along the slope of the tributary. During the many tens of thousands of years before the next glaciation, the waterfall retreated slowly upstream. Below the falls, a narrow rock-walled gorge marks this retreat. It seems reasonable that this process repeated itself during successive glacial and inter-glacial periods. During each glaciation, a new and lower fall closer to the main valley would form. Over the course of three glaciations, one would expect to see three sets of falls, with the oldest one cut back farthest upstream. Both Dingmans Creek and Sawkill Creek have widely-spaced falls that may correlate to different periods of glaciation. Raymondskill Creek does not. However, an “upper knick point” above Raymondskill Falls (Figure 219), formed by many closely spaced cascades, may represent an older and eroded falls. One reason for the absence of multiple, widely spaced falls on Raymondskill Creek may be that it is much younger than Dingmans and Sawkill Creek. Another explanation may be that the hanging valleys cut by the younger glaciations lie beneath thick deposits of meltwater sediment, far below the modern stream. The rock promontory at the mouth of the gorge 600 feet downstream from the lower falls is probably an abandoned falls that was notched and eroded. The former falls is probably a younger knickpoint related to a younger glaciation. Its position near the head of the valley-side reentrant lends itself to the possibility that a large part of this knickpoint may lie beneath alluvium and late Wisconsinan outwash. Both Conashaugh Creek and Dry Brook (Figures 211, and 219), which lie southwest from Raymondskill Creek, do not have waterfalls (other than a few very low cascades). These streams appear to be much younger than Raymondskill based on their much smaller drainage basins and narrower valleys. The lack of falls along these streams shows that the later glaciations were probably not nearly as effective forming waterfalls along the tributaries of the Delaware.

If waterfalls result directly from glacial erosion in Minisink Valley, then their heights may be used to estimate glacial erosion. Raymondskill Falls drops 125 feet (the actual height of the pre-notched

Figure 218. Comfort Station at Raymondskill Falls.

Figure 219. Stream profiles of Raymondskill and Conashaugh Creeks (panel A) and detailed profile of Raymondskill Falls (panel B).
falls may have been as much as 150 feet). The bedrock floor of Minisink Valley lies 150 to 200 feet below the Delaware River (unpublished data on file at New Jersey Geological Survey, Trenton, NJ). Added together, they show that glacial scour since the earliest glaciation to have reached this area (either 2.3 ma or .83 ma, based on the offshore $^{18}$O record discussed in Braun (1989) may have been as much as 350 feet. This estimate is reasonably close to the 450 feet (150 m) suggested by Braun (1989).

Raymondskill Falls: a Product of Multiple Glaciations

The morphology of Raymondskill Falls suggests that at one time it may have consisted of one larger falls or cataract, rather than the three smaller falls we see today. Abandoned falls, notched falls, and over-sized plunge basins in Raymondskill glen all show that the current falls are a product of multiple cycles of erosion and knickpoint retreat. Because waterfalls retreat largely by hydraulic plucking (the removal of rock by the impact of running water), increased rates of stream discharge due to the addition of meltwater, probably had a dramatic effect on erosion and shaping of the falls. During the late Wisconsinan, meltwater flowed from several sources, including nearby, small upland proglacial lakes, and stages five and six of glacial Lake Wallenpaupack (Duane Braun, personal communication, 2001). Although, it is tenuous at best to link these specific geologic events to the various erosional forms of the falls, there appears to have been at least six major periods of erosion (Figure 219; see also Sevon and Inners, this guidebook, p. 136).

1) Retreat from the original hanging valley to a valley-head position (late Pliocene to middle Pleistocene age),
2) Notching and retreat from the early glacial valley head falls position, and the formation of an upper and lower falls. Formation of large plunge basin below the greater upper falls (Illinoian).
3) Retreat of upper falls to form the middle and upper falls and formation of the middle plunge basin (late Wisconsinan).
4) Retreat of upper falls and formation of upper plunge basin (late Wisconsinan).
5) Cutting of narrow gorge above upper plunge pool and erosion of the falls that formed the upper plunge basin (late Wisconsinan).
6) Minor notching (< three feet) and retreat of the falls’ lips (< 10 feet) (postglacial late Wisconsinan and Holocene).

Lithologic and Structural Control of the Falls

Sedimentological and structural features control the morphology of the falls. Grain size variation within the Mahantango at this STOP is subtle. Thick packages of finely interbedded siltstone and fine sandstone do not portray much apparent differential weathering and yield a uniform weathering profile. Thinly bedded, slightly coarser grained sand beds do weather in relief but these units are uncommon. The most resistant beds are generally thicker, fine-grained sandstones. Still sediment grain size does play a sizeable part in the falls development. This is evident in two major ways:

1) Just above each major falls, the creek flows under a gentle gradient across a nearly flat bed. Once it reaches the edge, the water cascades over the falls. This topmost or “cap” bed is strong enough to resist the further recession of the falls. Layers that are more resistant originate from thicker uniform sand beds without intervening silts that create weaker bedding partings. Disrupted bedding caused by load casts or some other soft sediment deformation seems to prohibit cleavage development and thereby strengthens the bed. The cap bed on the upper falls is fine sand that shows ball-and-pillow and disruption of the overlying bed caused in part by the differential subsidence.
2) The cleavage development is variable within the Mahantango. In this region, cleavage is weakly developed and not strongly penetrative. Finer grained beds have a more persistent cleavage that leads to easier paths for water infiltration and subsequent frost shattering.
Preliminary mapping of waterfalls within the Mahantango suggests that joints play a major factor in their evolution. Joints appear more penetrative than the regional cleavage. Therefore they become avenues for water flow and subsequent enhanced erosion. A study of the upper falls (Figure 220) shows that joints have largely controlled its development. The main face of the upper falls is marked by a steeply southeast-dipping joint. The joint face is straight and smooth. Above this trace, the face appears to be receding by the selective removal of material along cleavage planes within individual beds. The slope angle above the joint is less steep. The same joint trend appears on the block jutting out over the falls. A second joint trend projects to the south and creates a "joint cave" similar to what is seen at site 3 on the loop trail. Again, the more penetrative nature of the joint as opposed to cleavage apparently receives higher water flow and enhanced erosion. All along the course of Raymondskill Creek this same feature can be seen.

Once sufficient weathering occurs and weakens the rock along a joint it will drop off and further set back the slope. An example of this process can be seen by looking down towards the plunge pool of the upper falls. South of the plunge pool a joint controlled setback appears in the wall rock. Visually tracing this trend up a small saddle appears. This joint location probably sited a previous falls location.

Regionally the joints do not parallel the cleavage trends. Figure 221 shows how the joints are subparallel to bedding strike and are straighter and more penetrative. Farther up stream from the upper falls, the joints and cleavage are more parallel to each other (Figure 222). Here the joints still control the falls development. Again, the more penetrative fabric of the
joints causes them to be more dominant feature. Cleavage planes change orientations or defract when entering beds of finer or coarser grain size. Joint dominance can be seen by their uniform wider spacing along the stream channel. These planes stand out stronger than the cleavage planes.

**Modification of the Falls by Glacial Derangement**

During deglaciation, meltwater streams may cut new stream courses in areas of thick glacial drift and modify preglacial drainage by damming valleys with drift. I.C. White (1882) suspected that the upper part of Raymondskill Creek in preglacial time flowed to Sawkill Creek through the valley of Motts Run (Figure 211). During one period of deglaciation, meltwater deposits (stratified drift) presumably blocked drainage to the Sawkill. Meltwater found a new path to the Minisink Valley, along the path that is the modern Raymondskill. Several pieces of evidence support White’s hypothesis: 1) Present-day Motts Run valley is filled with late Wisconsinan outwash. This material filled in a small, upland proglacial lake that spilled into Conashaugh Creek and later the lower part of Raymondskill. The upper part of the Raymondskill Valley above Silver Spring appears to be an older valley because it has a broad floor and low gradient, unlike its lower part, which is deeply incised and has a steep slope (Figure 211). Of the larger falls in the Recreation Area (see below), Raymondskill has retreated the least distance from the Delaware (retreat distance measured from base of falls to far edge of valley reentrant, determined by drawing a line along base of cliffs on either side of reentrant).

Retreat distances for various falls are as follows:

- Raymondskill Falls = 0.44 miles
- Pinchot Falls (Sawkill Creek) = 1.54 miles (Pinchot Falls also looks to be a product of glacial derangement.)

**Dingmans Falls (Dingmans Creek) = 1.36 miles**
- Bushkill Falls (Little Bushkill Creek) = 1.66 miles

Since Raymondskill Creek only ranks behind Sawkill Creek in terms of its discharge, its minimal retreat distance supports a change in the size of the Raymondskill drainage basin sometime during the Pleistocene as postulated by I.C. White.

**The Hemlock Glen: Unwelcome Guests**

The Raymondskill Falls trail leads to the falls area through one of Delaware Water Gap National Recreation Area's most valuable treasures, a natural hemlock ravine. While geologists may be intrigued by the story in the rocks, there is another story, equally compelling, in the struggle for life in the shallow...
soil and rocky environment. Hemlocks, the most common tree in the ravine, have shallow roots, which enable them to live in the shallow soil. Their probing roots wrap themselves around rocks and grow into any available crevice, grasping for a foothold. The ravine protects the trees from strong winds; however, erosion or damage to their roots can weaken their already tenuous foundation and fell a majestic forest.

The cold water of Raymondskill Creek chills the air in the glen creating a cool, damp environment. The dense foliage of the hemlocks also helps cool the glen by filtering out 95% of the sunlight before it reaches the forest floor (Battles et al., 2000). Streams entering a hemlock glen may drop in temperature by as much as 9°F (5°C) (Evans et al., 1996). This shade provided by the hemlocks is the difference between life and death for many species of insects and fish living in the stream. Because the temperature of their environment controls their body temperature, even a small change in water temperature can stress or kill them. One fish that requires cool water is the native Brook Trout. Small streams in hemlock forests are three times more likely to support reproducing Brook Trout populations than similar streams in hardwood forests (Snyder et al., 1998). These streams also support a greater variety of aquatic insects (35% more), an important food source for the Brook Trout (Snyder et al., 1998).

Several species of birds also depend on the hemlocks. Blackburnian warblers, black-throated green warblers, and blue-headed vireos all nest in hemlock trees, each using a different part of the tree for their nest. In Delaware Water Gap National Recreation Area, these three species are found almost exclusively in hemlock forests (Evans et al., 1996).

The hemlocks in the park’s ravines are in some cases 150 years old, survivors of a time when hemlocks were prized for bark and timber. A natural chemical known as tannin is found in hemlock needles and bark. This chemical is an important natural pesticide that protects the trees from insect pests. When fallen bark and needles decompose these natural pesticides are released, making the soil and water acidic and staining the water the color of tea. Tannins were useful to settlers as part of the tanning process. In the mid to late 1800s, most (70%) of Pennsylvania's extensive, mature hemlock forests were cut down for bark to supply tannins to the world's largest leather tanning industry (Whitney, 1990).

Perhaps the biggest threat to the hemlocks today, is the hemlock woolly adelgid (Adelges tsugae), an aphid-like insect that sucks the sap from the trees. Hemlock woolly adelgid is an exotic species, and because of a lack of predators in the United States, its populations are largely uncontrolled. Trees infested with hemlock woolly adelgids often lose most of their needles and may eventually die because of the infestation. The weakened trees are vulnerable to attack by a native beetle, the hemlock borer beetle (Melanophila fulvoguttata) that attacks and kills the trees within a few years. Hemlocks infested with hemlock borer beetles attract woodpeckers, which chip of the bark in an attempt to capture and eat the beetles. This chipping gives affected trees a characteristic red trunk.

In an attempt to save the trees, the park has initiated a program in which Japanese ladybird beetles (Pseudocymnus tsugae) are released into infected hemlock stands. Because the ladybird beetle is a wooly adelgid predator, it is hoped that the beetle will control the adelgid populations in the park.

THE FALLS LOOP TRAIL

*Please follow the loop trail shown on Figure 223. Because the observation areas around the falls are small, allow the groups in front sufficient time to move along the trail. Stepping-stones along the trail are slippery; proceed with caution.*
Site 1. Trail intersection and NPS placard describing several plants that thrive in these shaded, cool, and moist hemlock glens. The rock promontory here may represent the former position of a waterfall that has long since been abandoned. Erosional slope processes have further reduced the size of the falls. Rock promontories (shoulders of rock that jut outward from steep slopes) are common in Raymondskill glen. Many of them are former waterfalls that have been deeply notched; their former faces have retreated far upstream. In some places, the intersection of closely spaced vertical joints and cleavage in the Mahantango Formation forms a zone of weakness that is easily eroded. Over time, differential erosion along the steep walls of the ravine will form a high-standing shoulder of rock. Weathering and moss have obscured many of the sedimentary structures in the nearby outcrops. If one looks closely, the bedrock is a thinly, interbedded siltstone and fine sandstone, probably from facies 2 of Prave et al. (1996) (Figure 215). Subtle hummocky cross stratification is also present.

Site 2. Upper observation deck and top of the upper falls. The small observation deck is partly built on the lip of the upper falls. Postglacial erosion has carved a 25-foot wide notch, 15 feet back from the lip. The Mahantango shows some good sedimentary structures here. Firstly, ball-and-pillow structures can be seen on the northern wall, just below the path you just descended (Figure 216) and just across the falls on the block that juts out over the falls. Next to the overlook is a strange bedding feature believed to be related to the differential subsidence related to the ball-and-pillow structure. It resembles pinch-and-swell textures with the arcuate lines termed rib-and-furrow and suggesting northwest-directed transport direction (Ricci Lucchi, 1995). Joints can be seen to control development of the fall faces. Joints trend differently (N46°/71°SE on face of falls and N65°E/73°SE into the joint cave on opposite shore) than the cleavage (N73°E/70°SE). These joint faces appear to control the weathering and retreat of the falls. There are several interesting features above the upper falls (Figure 223) that tell an intricate story of knickpoint retreat, and notching. The first is the abandoned plunge basin that forms the upper pool. Clearly, this basin is not related to the modern Raymondskill Creek. Thirty feet above the creek, and to the left of the narrow ravine at the head of the plunge basin, is an abandoned falls. It is similar in elevation to the high surface on the far side of the pool. The rocky surface behind the abandoned falls is devoid of gravel or till clasts, having the appearance of being washed by a great torrent of water. In places low rock ledges mark former cascades. Based on the width of this upper surface, imagine a stream two to three times wider than the upper pool and plunging over a falls 25 feet above our heads. If
one goes back a little farther in time, imagine this falls in its former position near the lip by the observation deck, a cataract of about 80 feet in height. What a great river this must have been! Based on the size of the Raymondskill Creek drainage basin only glacial Raymondskill Creek could have carved these features. The rock-cut gorge upstream from the plunge basin (Figure 212) represents a later stage of meltwater erosion that presumably was caused by a decrease in stream discharge and channelization behind the abandoned falls. The timeline for this story is unclear. The washed surface above the abandoned falls (Figure 223) shows that late Wisconsinan meltwater scoured this surface. Otherwise, it would have been covered by till, or outwash. Based on this evidence, the story of knickpoint retreat above the upper falls (70 feet) and notching (30 feet) is one that was played out during the last glaciation.

Stop 3. Joint cave/rock shelter (Figure 224): Entrance is 5 feet high x 2 1/2 feet wide; inside dimensions are 7 feet high x 7 feet wide x 16 feet long. The roof of the cave is partly held in place by hemlock roots. Joint caves are common in the Mahantango Formation in places where there are steep slopes. Rock fragments were preferentially removed in places where the orientation of cleavage, joints, and bedding plane intersections, formed highly fractured zones. The fractured rock is loosened and dislodged largely by frost shattering. Root growth in places plays a secondary role. Most of the debris in the cave was removed to build the nearby trail.

Stop 4. Lower observation deck. Good view of upper and middle falls (Figure 213), and top of lower falls. Several abandoned falls positions including area around observation area can be seen from here. Raymondskill Creek does not plunge over the upper and middle falls in two consecutive drops, but rather over a series of steep, widely spaced steps. The step-like character of the falls and the lower cascades in Raymondskill glen is closely related to bedding thickness, grain size, and the orientation and spacing of joints and cleavage. The large plunge basin below the middle falls is also not a product of the current geometry of the falls. The key to its origin lies well above our heads. Left of the upper falls and partly screened by vegetation is a high rock knob that is similar in height (6 feet lower) to the lip of the upper falls. This knob is an abandoned falls whose position marks the location of the upper “greater” falls. Prior to knickpoint retreat from this position, the middle falls did not exist. Formation of the middle falls involved at least two phases of erosion. Phase 1 consisted of retreat to a position a marked by the face of the middle falls. The deep plunge pool below the observation deck marks the former location of the falls. Phase 2 consisted of retreat to the position marked by the face of the upper falls, which included the formation of the upper and middle falls, and small plunge basin between the two falls. The age of these erosional phases is unclear. Based on the amount of erosion upstream, it appears that the middle falls may have also formed during late Wisconsinan time. The timing of retreat from the lower falls is tenuously placed as pre-Wisconsinan. The large joint-block boulder near the lip of the lower falls has channelized stream flow, resulting in additional scour of the channel bed and erosion along the lip of the lower falls.
**Stop 5.** Small notch cut in the Mahantango cut by late Wisconsinan meltwater. This cut offers another example of highly moss covered and weathered outcrop that hides the nature of the rock. On the south side of the exposure, a six-inch thick sandstone bed stands out against the otherwise siltstone and fine sandstone. Around the corner on the uphill (west) side of the path, there is an example of differential loading that creates a circular weathering pattern on the rocks. Above this, if one climbs up some, the rocks show a thinly interbedded nature and yield hints of laminations and hummocky cross bedding.

**Stop 6.** High tech, environmentally friendly, on-site waste composting facility (Figure 218). Private, spacious, his and her rooms, naturally cooled, energy efficient, mostly built with natural materials, including imported roofing slates. Cost $800,000 to a cool $million.

Leave STOP 12, returning to US 209 via SR 2009.

- 0.5 37.8 Turn right on US 209 South.
- 5.0 42.8 Traffic light at intersection with PA 739. Continue straight ahead.
- 11.7 54.5 Traffic light (blinking) in Bushkill.
- 0.3 54.8 National Park Visitors’ Center to left.
- 0.3 55.1 Bear left onto Community Drive (Middle Smithfield Township 633).
- 1.2 56.3 Stop sign. Turn left on River Road (SR 2028).
- 0.2 56.5 National Park Headquarters to left.
- 7.1 63.6 Turn left at entrance to Shawnee Inn.
- 0.3 63.9 Front of Shawnee Inn. End of Day-2 field trip. Have a safe trip home!
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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.

<table>
<thead>
<tr>
<th>Miles</th>
<th>Int.</th>
<th>Cum.</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>Parking area on left side before traffic light at entrance to Old Mine Road, north end of Delaware Water Gap.</td>
<td></td>
</tr>
<tr>
<td>0.4</td>
<td>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.</td>
<td></td>
</tr>
<tr>
<td>0.6</td>
<td>Trailhead of the Farview Hiking Trail.</td>
<td></td>
</tr>
<tr>
<td>0.2</td>
<td>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 <em>Vernonaspis</em> and <em>Americaspis</em>. Because the outcrop lies within the National Park, collecting is strictly verboten.</td>
<td></td>
</tr>
<tr>
<td>0.4</td>
<td>Outcrop of Bloomsburg Red Beds on right.</td>
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<td>0.2</td>
<td>Pass Shawnee Inn and Golf Resort across Delaware River on your left.</td>
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<tr>
<td>0.5</td>
<td>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.</td>
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<td>1.0</td>
<td>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).</td>
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<tr>
<td>0.8</td>
<td>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.</td>
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<tr>
<td>0.6</td>
<td>Tocks Island (upstream end).</td>
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<td>0.2</td>
<td>Exit Worthington State Forest.</td>
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<tr>
<td>0.6</td>
<td>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.</td>
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</table>
| 0.8   | 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°W and S78°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.
GEOLOGIC FEATURES ALONG THE RED DOT-BLUE BLAZE-DUNNFIELD CREEK TRAILS
DELWARE 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 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: Dunning 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.

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

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

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

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

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

**Syncline**: U-shaped downfold of rock layers.

**Talus**: An apron of irregular rock fragments derived from and lying at the base of a cliff.

**Till**: Unsorted mixture of clay, sand, and boulders deposited beneath a glacier.
GEOLOGIC FEATURES ALONG THE ARROW ISLAND TRAIL

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 quarry.

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

Stop 5. Original horizontal sediment layer (bedding; solid line) is now tilted. The rock breccia 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 (terming 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 (CaCO3) 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 (MgCaCO3) 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 Pennsylvania 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.


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.


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


10. Smithfield Beach. Tocks Island Discussion, pullout.
## Appendix D

### Table D1

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</tbody>
</table>
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>
<thead>
<tr>
<th>Direction of wind flow</th>
<th>Number of readings</th>
<th>Temp A 2</th>
<th>Temp B 2</th>
<th>Temp C 2</th>
<th>Wind Speed</th>
</tr>
</thead>
<tbody>
<tr>
<td>In</td>
<td>17</td>
<td>42.5</td>
<td>45.1</td>
<td>45.8</td>
<td>173</td>
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<tr>
<td>Out</td>
<td>44</td>
<td>34.3</td>
<td>44.9</td>
<td>54.3</td>
<td>175</td>
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<tr>
<td>All</td>
<td>64</td>
<td>36.8</td>
<td>45.5</td>
<td>52.1</td>
<td>170</td>
</tr>
</tbody>
</table>

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